

## Preliminary preface to NMSOP-2

Ten years have passed since the first printed edition of the IASPEI New Manual of Seismological Observatory Practice (NMSOP 2002). Some 2,000 hard copies are currently in use in more than 100 countries at seismological observatories, data and analysis centers, in teaching, research, and field applications, national and international seismology training courses, or by private enterprises and individual scientists. In 2006, Seismological Press Beijing published a two-volume hard cover edition in Chinese. In 2010 the Badan Meteorologi Klimatologi dan Geofisika (BMKG) in Jakarta, with support of the Japan International Cooperation Agency (JICA), published the first 3 Chapters of NMSOP in (Bahasa) Indonesian language. Three more Chapters are currently being translated and scheduled for publication in 2012. The first four Chapters have been translated into Turkish, with more being in preparation, and Russian translations of several NMSOP Chapters are used at seismological observatories and analysis centers in Russia and other countries of the Commonwealth of Independent States (CIS) already since 2006.

Such global adaptation and use, signals a growing demand for the kind of information provided by the NMSOP and time for a rigorous update. NMSOP-2 will be published as a pure electronic edition. It is again the result of a great cooperative international effort: 73 authors from 18 countries have contributed to (or are still working on) it. More than 70 experts from 20 countries provided peer reviews, and Dr. Siegfried Wendt, himself author and co-author of several contributions to this Manual, has compiled a long-period filtered teleseismic broadband record section of the Collm (CLL) observatory in Germany, which now embellishes the front cover page of NMSOP-2. In this spirit, with this aim, the NMSOP needs to be maintained and further developed under the auspices of IASPEI and its Commission on Seismological Observation and Interpretation (CoSOI). Only then is it assured that the seismological community can always refer to an up-to-date and IASPEI-authorized guidance in observatory practice in the widest sense of modern applications.

The electronic version of NMSOP-2 will be realized in several steps during 2012, as manuscripts become available after completed review, revision and editing. Future revisions of NMSOP will have to adopt this approach of piece-wise realization anyway. An overall revision of such a huge and topically complex publication within a fixed time frame and in mostly voluntary work is no longer feasible and practicable, and the current editor will no longer be available for it. Therefore, anyone who wishes to propose further improvements or new topics or to make such a contribution himself is highly welcome and should contact the editor Torsten Dahm ([torsten.dahm@gfz-potsdam.de](mailto:torsten.dahm@gfz-potsdam.de)) and/or the CoSOI chairman, Dr. Dmitry Storchak, Director of the ISC ([dmitry@isc.ac.uk](mailto:dmitry@isc.ac.uk)).

NMSOP-2 and future follow-up versions will be freely accessible and downloadable. The newly chosen format (see *Editorial Remarks*) permits easy upgrading and complementing of NMSOP in accordance with rapid developments and changing requirements. It also minimizes the need for an overall editorship by emphasizing personal initiative and responsibility of all NMSOP authors, both current and future ones, for their contributions under the auspices of CoSOI/IASPEI and ISC. Thus, the IASPEI New Manual of Seismological Observatory Practice now becomes a **truly dynamic publication**.

NMSOP-2 provides access and download options for many valuable computer programs on:

- Event location
- Ray tracing
- Seismogram analysis
- Parameter determination
- Instrument calibration
- Record filtering
- Fourier analysis
- Network modeling

Moreover, we have complemented several topical contributions and exercises by animations (movies). They can be activated or downloaded (as the programs) from the listing *Download programs and files* (see overview on the NMSOP-2 cover page). These movies illustrate striking features, such as:

- Ray propagation of different seismic phases through the Earth interior;
- Great slowness differences of their propagating wavefronts on the Earth surface;
- Formation of related seismic records and travel-time curves in a regional network;
- Dependence of source radiation patterns and directivity on source azimuth, epicentral distance, fault strike, dip angle, and slip direction.

We are strongly interested to expand such new NMSOP features and invite other authors, who have developed similar modules of high educational value, to make them available as complementary material to relevant topics of the electronic NMSOP edition. Proposals should be addressed to the editor Torsten Dahm ([torsten.dahm@gfz-potsdam.de](mailto:torsten.dahm@gfz-potsdam.de)).

### **What necessitated a second edition of the NMSOP?**

Since the first NMSOP edition in 2002 rapid developments have taken place in many fields of observatory seismology, concerning both sensor and data acquisition technologies, the growing automation of data analysis and interpretation, widening applications of seismological technologies and methods in volcano monitoring, rupture tracking, microzonation, hazard assessment, earthquake and tsunami early warning, strong-motion, ocean bottom and engineering seismology as well as other fields. This necessitates upgrading of several existing NMSOP (2002) chapters and auxiliary materials as well as the addition of new ones.

At the 2007 IUGG/IASPEI General Assembly in Perugia, Italy, the Editor proposed to guide the elaboration of a purely electronic, significantly amended version of the NMSOP. CoSOI accepted this initiative and re-established a related NMSOP Working Group. The detailed NMSOP-2 project plan was presented two years later at the 2009 IASPEI General Assembly in Cape Town, South Africa. An essential component in this plan was the commitment of the library of the GFZ German Research Center for Geosciences in Potsdam to host and professionally maintain on a long-term basis the electronic versions of NMSOP. This commitment was demonstrated in the same year by putting the editorially slightly revised and corrected 2002 edition as NMSOP-1 on the Internet and allocating DOI-numbers to all reviewed contributions.

NMSOP-1 will continue to be available on the GFZ website also in future. This assures preserving many valuable older pieces of information which NMSOP-2 will not duplicate but refer to. Since October 2011, all electronic versions of NMSOP have their own, very easy to access website address: <https://nmsop.gfz-potsdam.de/>, which is mirrored on the Websites of



IASPEI (<http://www.iaspei.org/projects/NMSOP.html>) and of the International Seismological Centre (<http://www.isc.ac.uk/standards> ).

We have opted for a web-only edition of NMSOP-2. The reasons are pragmatic and economic: Increased mailing costs for bulky hard copies, out-rating even production cost, would not assure that NMSOP-2 will be affordable for those most in need of it. On the other hand, rather cheap and speedy computer and Internet facilities have meanwhile become available on a global scale, even in most of the earthquake-prone low-income developing countries.

To ease the orientation of users in the current, still incomplete preliminary version of NMSOP-2, we have compiled a complete *List of* (all expected) *NMSOP-2 titles* (see cover page). Already finalized and downloadable items, preliminary versions and those still in the pipeline are marked therein in different color. For more information consult on the cover page *Editorial Remarks* and *Rights\_Permissions\_Acknowledgments\_References\_NMSOP\_2*.

Special thanks go again to Ms. Margaret Adams (UK/USA) for final English proof-reading of all manuscripts of the first edition, and to Ms. J. Suckale, Ms. A. Sachse, Ms. U. Borchert, Ms. R. Milkereit, Mr. Ch. Nerger and Mr. L. Gabrysch for their valuable technical assistance in consistent formatting, drawing of figures and compiling the lists of acronyms, the glossary and the index for the first NMSOP (2002) edition. Although the current Web edition has been significantly amended, much of these early efforts have still been preserved in NMSOP-2. For this 2<sup>nd</sup> edition, however, the undersigned had to accomplish all editorial work alone, without any such technical or clerical support. Therefore, any short-comings in this regard are solely his responsibility. The electronic version makes it now easy to eradicate both errors in substance or annoying typos. Manual users which spot such errors or typos are kindly requested to inform the Editor.

NMSOP authors and NMSOP Working Group members would like to express their gratitude to IASPEI for entrusting this important task to them and for the continuous interest taken and encouragement provided by CoSOI. Particular thanks go to the many external reviewers (see *List of Reviewers*). Their constructive criticism and suggestions have greatly facilitated the completion of this work, helped to improve original drafts and thus to assure high standards of both professionalism and comprehensibility of the texts for the very broad NMSOP user community. More specific acknowledgments are given at the end of individual NMSOP chapters or auxiliary items.

Most important for the implementation and maintenance of NMSOP-2 on the web has been, and will continue to be for years to come, library of the GFZ ([bib@gfz-potsdam.de](mailto:bib@gfz-potsdam.de)). The library team, in close collaboration with the GFZ computer centre, will be the future key point of contact for both the authors and CoSOI in assuring the correct and comfortable long-term representation and upgrading of the IASPEI Manual. The editor has greatly benefitted from his expertise and cooperativeness in developing a suitable concept tuned to the specific potential of the web edition of NMSOP-2 at the GFZ. Gratefully acknowledged in this context is also the financial support granted to the editor by the GFZ in this final phase of completing NMSOP-2.



Peter Bormann (NMSOP Editor)

Potsdam/Kleinmachnow, January 2012

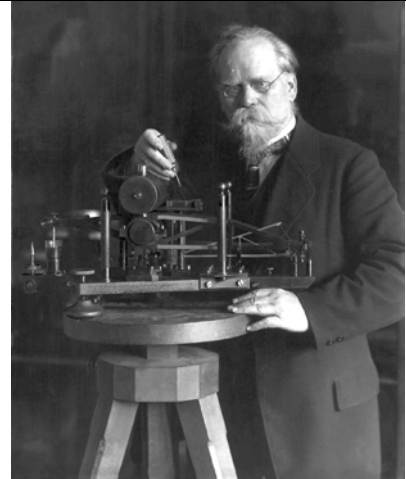
The 2<sup>nd</sup> Edition of the  
IASPEI New Manual of Seismological Observatory Practice  
is dedicated to the

**150<sup>th</sup> birthdays**  
**of the founders of modern observatory seismology**

**Emil WIECHERT (1861-1928)**

- Born on 26.12.1861 in Tilsit
- 1881-89 study of Physics at Königsberg (now Kaliningrad)
- 1896 discovers the electron, determines its ratio mass to electric charge and proposes that the Earth's core consists of iron
- 1898 world-wide first appointment as Professor of Geophysics (Göttingen Univ.)
- 1900 first WIECHERT pendulum seismometer
- 1901 attendance of the International Seismological Conference at Strasbourg, becoming one of the "founding fathers" of the International Association for Seismology, the forerunner of IASPEI
- 1902 opening of the seismological station Göttingen
- 1903 publication of the "Theory of automatic seismographs"
- 1904 17t horizontal pendulum at Göttingen station
- 1910 WIECHERT-HERGLOTZ formulas, allowing the exact analytical solution of the inverse problem for calculating 1D velocity-depth models from continuous prograde travel-time curves.
- Passed away at Göttingen on 19.03.1928
- About 180 WIECHERT seismographs of different type were operated at more than 110 observatories world-wide, some of them still being in use today.

and



Emil Wiechert with one of his vertical component seismographs

**Boris Borisovich GALITZIN (1862-1916)**

- Born on 02.03.1862 in Saint Petersburg
- Study of physics and mathematics at Strasbourg University
- 1893 "Research into the Mathematical Physics" with which he paved the way for the emergence of quantum mechanics
- Since 1894 Director of the Physical Cabinet of the Russian Imperial Academy of Sciences
- 1901 proposes and builds the first vibration table for instrument parameter calibration
- 1902 General theory of the horizontal pendulum; proposes electromagnetic transducer and pendulum damping with galvanometric-photographic recording
- 1907 proposes at the meeting of the International Seismological Association the general introduction of damped pendulums
- Design and testing of a piezoelectric accelerometer
- 1911 first determination of the energy of an earthquake by estimating the energy density from distant seismic recordings
- 1912 publication of his pioneering "*Lectures on Seismometry*"
- 1913 elected as Director of the Main Physical Observatory
- Developed the first program on earthquake prediction research and hypothesized about triggering effects of seismicity
- Detailed study of microseisms for forecasting cyclonic activity
- Galitzin seismometers were used world-wide until mid 1970s
- Passed away on 17.05.1916



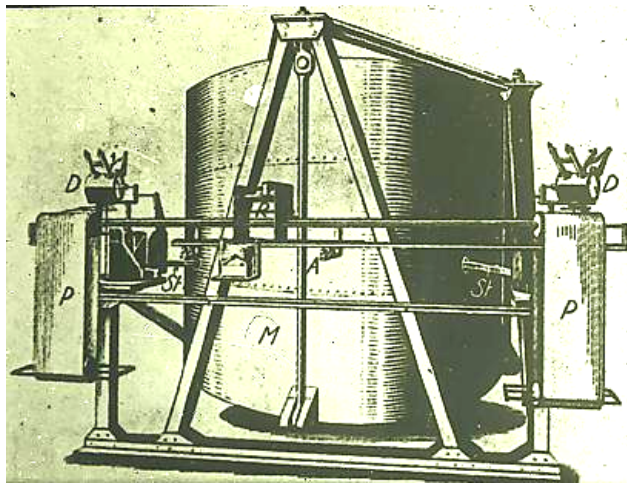
## Emil WIECHERT's motto:

(written in 1902 over the entrance to the seismological station Göttingen, Germany)

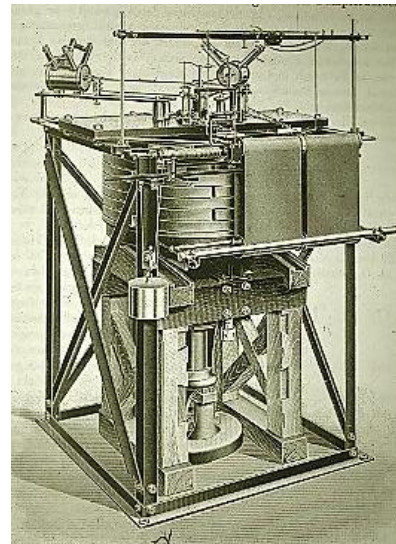


*The trembling rock  
bears tidings from afar -  
Read the signs!*

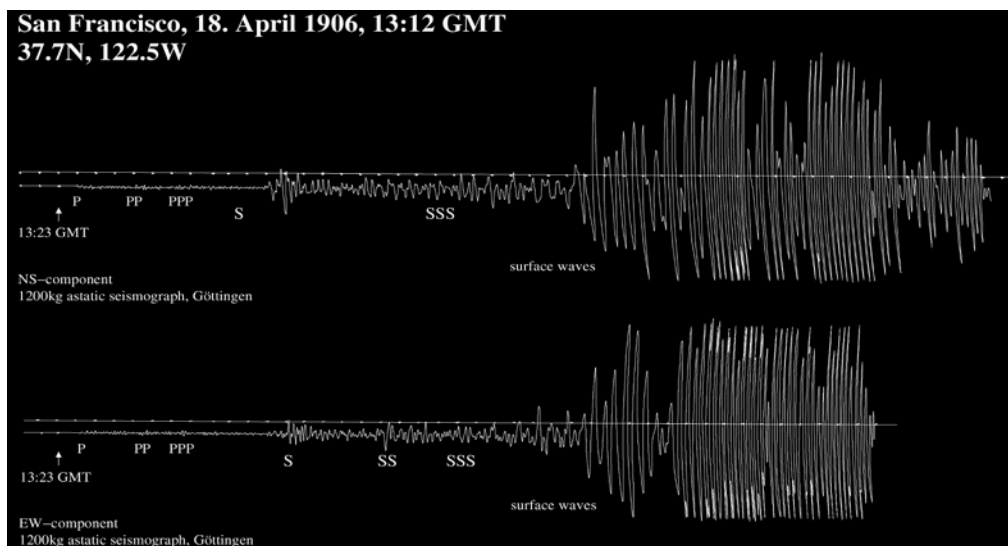
## Seismographs and recordings



The 1904 17t WIECHERT pendulum ( $T_s = 2.2$  s, magnification  $\approx 2000$ )

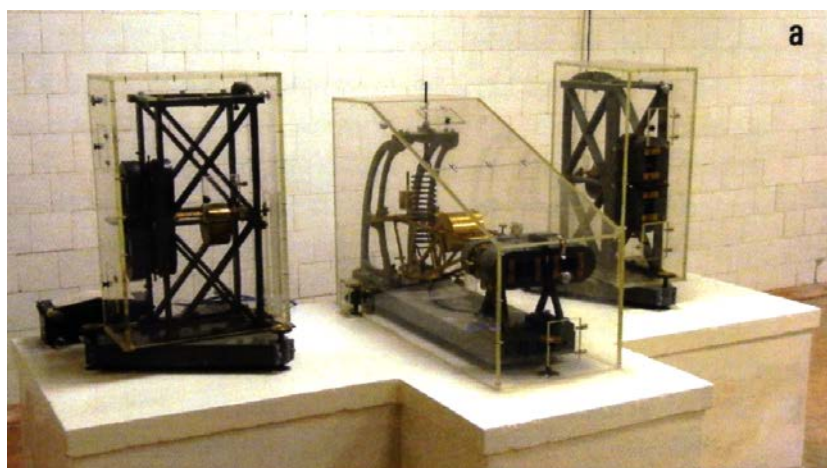


The 1904 WIECHERT astatic horizontal component pendulum (mass = 1.2t,  $T_s \approx 8.5$  s, magnification  $\approx 200$ ) and record of the  
↓ 1906 San Francisco earthquake at Göttingen

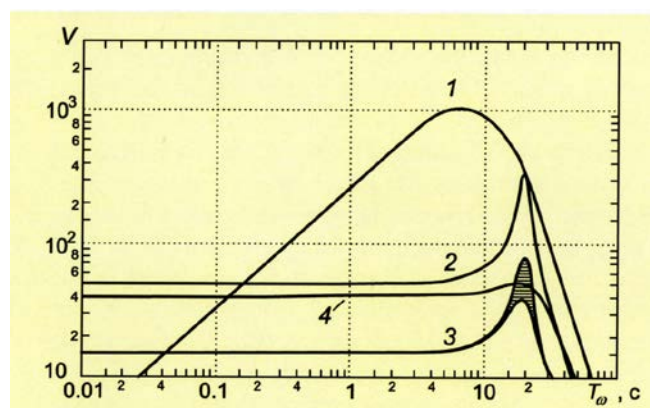




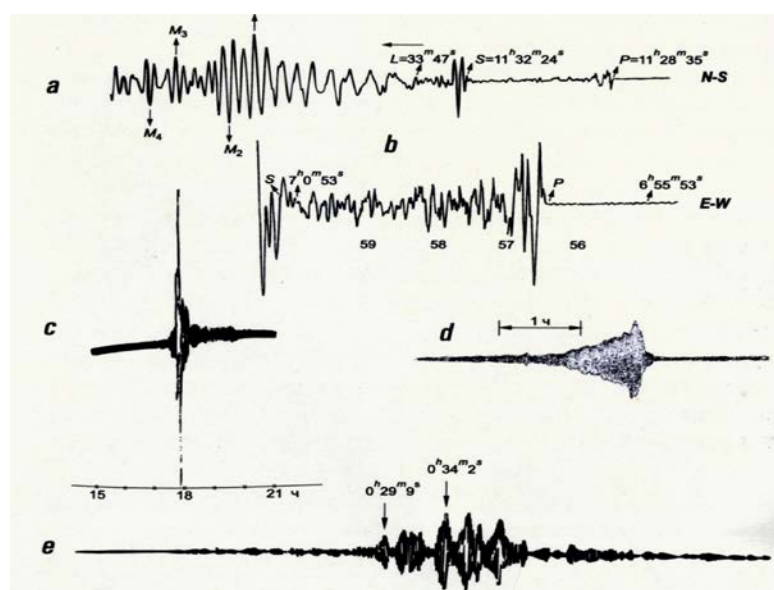
## Boris Borisovich Galitzin's seismographs and records



3-component set of electro-dynamic GALITZIN seismographs at Pulkovo station  
(Photo by A. Ya. Sidorin)



Amplification curves (1) of the electrodynamic GALITZIN seismograph with galvanometric recording, (2) of the Rebeur-Paschwitz seismograph, (3) of the Milne seismograph, and (4) of the heavy horizontal GALITZIN pendulum. Note the resonance peaks of the unattenuated seismographs (2) and (3).



Record examples (a-b) of the electrodynamic GALITZIN seismograph, (c-d) of the Rebeur-Paschwitz instrument and (e) of the Milne seismograph.

## **Complementary remark**

The year 2012 is also the 150<sup>th</sup> anniversary of **Radó Kövesligethy (1862-1934)**, the founder of the seismological observatory Budapest (1905) and Secretary General of the International Seismological Association (ISA; 1905-1922). Kövesligethy made several important theoretical and methodological contributions to seismology, amongst them the macroseismic procedure of depth determination of earthquakes and, in 1905 at the Hague conference of the ISA, he proposed the calculation of the emergence angle of seismic rays based on 3-component records. This stimulated B. B. Galtizin to develop his famous vertical-component seismograph. For a detailed account of the work of R. Kövesligethy see on the IASPEI website <http://www.iaspei.org/newsletters/newsletters.html> Peter Vargas contribution to the IASPEI Newsletter of November 2012.

## **Acknowledgments**

The dedication has been compiled by the Editor, P. Bormann, using different sources. The Wiechert portrait photo and the San Francisco earthquake record were copied with kind permission of the University of Göttingen, Germany, and the drawings of the Wiechert pendulums from Galitzin's (1910) publication on seismometry.

The portrait of B. B. Galitzin, of his instruments as well as of the comparative responses and record examples were copied, with kind permission of the authors Alexander Ponomarev and Alexander Sidorin, from a commemorative leaflet prepared on the occasion of the 33rd General Assembly of the European Seismological Commission (ESC), Moscow, 19-24 August 2012. For more details see also the IASPEI Newsletter of October 2012.



The New Manual of Seismological Observatory Practice (NMSOP 2002) was originally published in two printed volumes. Volume 1 comprised 13 topical Chapters, while Volume 2 contained different kinds of complementary annexed material. Annexed to most of the Chapters were Datasheets (DS), Exercises (EX), Information Sheets (IS), Program Descriptions (PD) and under Miscellaneous Acronyms, Glossary, the summary list of References and an Index. The only slightly revised first web edition of the Manual in 2009, thereafter termed NMSOP-1, reproduces the contents and outlook of NMSOP (2002)

Although the second electronic edition NMSOP-2 (2012) keeps the general classification of content and the internal structure of earlier Manual issues, it is no longer subdivided into two volumes. It also no longer requires an *Index* and a summary list of *References*. Powerful electronic search tools make a typed Index obsolete, and the saving of printed pages by avoiding repeated quotations of the same sources in the reference lists of individual papers is also no longer a valid argument for an electronic manual. Any reviewed individual contribution to NMSOP-2 now has its own DOI-number and can be downloaded separately. Therefore, more benefit is seen in considering each Manual contribution as a self-contained publication with its own reference list and the complete addresses of the authors. Accordingly, also the NMSOP (2002) summary *List of Authors* and *List of Contents* have been skipped. Instead, each NMSOP-2 item with more than 10 pages (with the exception of DSs and PDs) is now preceded by its own list of contents with page numbering. This gives readers a quick overview from the outset and eases the targeting of topics of specific interest.

What else is new in NMSOP-2? Four Chapters (14 to 17) and many new complementary materials have been added, including power point presentations, educational animation films and software, which also can be downloaded individually. Most other items of previous editions have been revised and often significantly amended as well. Both new and amended old contributions are recognized by DOI-numbers which contain “NMSOP-2\_...”, in contrast to unchanged earlier contributions. The latter are linked by their DOI-numbers, containing “NMSOP\_r1\_...”, to the only in some items slightly revised NMSOP-1 (2009) edition.

The original numbering of Chapters and complementary materials remains unchanged in NMSOP-2. Each Manual Chapter is broken down into sections and sub-sections, e.g., Chapter 5, Section 5.1, Sub-section 5.1.2, or even further. The first number always relates unambiguously to the respective Chapter. The same applies to the numbering of figures, tables and equations. Also annexed DS, EX, IS, and PD are numbered according to the Chapter to which they refer, followed by a sequence number, e.g., DS 3.1 is the first Data Sheet belonging to Chapter 3 whereas EX 5.4 is the Exercise number 4 related to Chapter 5. All Manual items have thus been linked with each other editorially, easing mutual cross-references, balancing of contents and avoiding unnecessary duplications of issues.

Not numbered, however, are texts and listings that relate to the Manual as a whole, such as *Acronyms*, *Editorial Remarks*, *Glossary*, *Prefaces*, and *Rights\_Permissions\_Acknowledgements\_References*. These files can be opened from the Manual content **Overview** on the NMSOP-2 cover page. The acronyms used in the Manual sometimes relate to institutions, organizations or programs that have their own web site; in those cases, we have added the web addresses in the alphabetical list of *Acronyms*.

The Manual is utilized by a very broad user community, including beginners, high-school students, teachers, disaster managers or specialists from other disciplines, who may not be sufficiently familiar with specific seismological terminology. Therefore, the *Glossary* in this

Manual includes also terms that may be common knowledge to seismologists but require a simple explanation for other users. When compared with the first edition, the *Glossary* in NMSOP-2 has been tremendously expanded by merging and harmonizing the glossaries in NMSOP (Bormann, 2002 and 2009), by Aki and Lee (2003) and by Lee (2009), and by adding additional entries of the editor and several other Manual authors. With nearly 200 pages this is at present the largest-ever published glossary explaining seismology related terms. [Therefore, users who have difficulties to understand some technical terms out of the context should consult the \*Glossary\* for a more comprehensive definition and explanation.](#) Yet, still the editor considers this current version to be a preliminary and maybe forever an open-ended one. Therefore, proposals for further amendments or revisions are highly welcome. [The same applies to the \*Acronyms\*.](#)

[Throughout the NMSOP-2 many links to external website have been given and their functioning been tested several times. However, we have realized that many institutions currently restructure rigorously \(or even close\) their previous websites. Thus, at the time of this writing, we found already some links not working that were still fully operational a few months ago. Therefore, neither the GFZ German Research Centre for Geosciences, nor the Editor or the authors can guarantee the correctness and functioning of any links given to external websites of institution, programs, firms of product. In case of link failures users are advised to “google“ for the respective “term”.](#)

In NMSOP-2 the names of authors and their addresses are given below the title of each Manual item. Several Chapters are collaborative work of all named authors, making it difficult to single-out individual contributions. The name of the main contributor or coordinator is then usually given first. Questions or comments should be addressed to him with a copy to the Editor. At times co-authors have contributed only to specific parts. When this case arises, their names are given in the list of contents behind the title of the respective sub-chapter/section. Questions or comments may then be addressed directly to these co-authors, again with a copy to the Editor and the main author of the Chapter for information. The same applies to annexed material.

The Manual contains many cross-references to figures, tables and equations in both Chapters and annexed items. The following nomenclature eases unambiguous cross-references:

- Annexed items are referred to by the acronym of their “class name” (e.g., DS; EX, IS, or PD) followed by the Chapter number to which they relate, and the sequence number, e.g., IS 2.1, EX 3.3, or PD 5.3.
- However, when referring to any sub-division of a Chapter we give only its respective number, e.g., see 4.2, or see 11.2.3.4.
- When referring to annexed items which relate to the Manual as a whole, we refer to the name of the folder (e.g., see Acronyms, Glossary, Prefaces, ...).
- Equations, figures and tables are numbered individually within each Chapter or annexed item; equation numbers are always given in brackets.
- Equation numbers in Chapters give first the Chapter number, followed by the sequence number of the equation within that Chapter, e.g., (5.7) or (3.43), whereas equations in annexed items are given by their sequence number only, e.g. (6).
- The same applies to figure and table numbers, with the exception that these numbers are not put in brackets, e.g. Fig. 5.3 or Tab. 3.1.
- References to figures, tables or equations in annexed items are made as follows: e.g., see Figure 4 in DS 11.2, or see Table 1 in EX 3.2, or see Equation (12) in IS 3.1.

NMSOP-2 offers access and download options via hyperlink to many computer programs, animations (educational movies), power point presentations and other types of files. The respective folder or file names are printed in the related texts in bold black. All these names are summarized in a listing of *Download programs and files* (see NMSOP-2 cover page). They can be downloaded and/or activated via hyperlink from this summary listing by right mouse click on the respective link names, which are printed in bold blue letters. Sub-routine or file names which are printed in this listing in bold black letters are only part of larger program packages. Related information, also about authors and the type of programs or files are given in this listing as well.

We use American English as consistently as possible, however some modifications may occur and hopefully will be tolerated. Experts from many nations have contributed both to the Manual drafts and the review process, thus different opinions about correct style, grammar and punctuation have been unavoidable. For example, there are pros and cons for writing either “Earth”, or “earth” depending on whether one refers to the whole planet or only parts of it. We opted for “Earth”, as in the old Willmore (1979) Manual. Accordingly, we speak of Earth models, Earth tides, etc. We avoided using the term ‘Earth’ when specifically referring to the Earth’s crust, lithosphere, mantle, or core. No consensus could be reached between American, Australian and English reviewers as to when to use the apostrophe (e.g., as in Earth’s Interior) and when not (as now common in Earth models or Earth systems). Therefore, we have tried to avoid its use. Despite all efforts, and taking into account that the majority of authors, including the Editor, do not have English as their mother tongue, typos, misnomers and occasionally clumsy phrasing may still be found. Some references to figures, equations, tables, publications or sub-sections may still be missing or incorrect; the Editor will be grateful for any suggestions and/or corrections.

Ten years ago, the IASPEI Working Group on Phase Names agreed on a list of standard phase names. It includes old, partially changed and new names. The list is reproduced in IS 2.1. It has been approved by IASPEI and was published by Storchak et al. in 2003, and, slightly modified, in 2011. As far as possible we have applied the new phase names, e.g., Pdif instead of Pdiff and PKPdf instead of PKIKP. Yet, some of the figures have been reproduced from other sources or have been plotted by programs for seismogram analysis that still use the old terminology. This explains some inconsistencies in the Manual with respect to phase names, but we strongly favour the consistent use of the standard phases names as given in IS 2.1.

With respect to magnitude symbols, different nomenclatures are used in NMSOP-2:

- a) the generic NMSOP (2002) nomenclature, which largely agrees with that in most pre-2010 seismological publication, with the exception of writing no subscripts and Ml instead of ML.
- b) the NMSOP (2002) specific nomenclature (see IS 3.2), because it describes unambiguously the type of seismic wave used for magnitude determination, the record component on which the amplitudes have been measured and the response type of the seismograph;
- c) the IASPEI (2011) nomenclature for standard magnitude and amplitude readings (see [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf)), which we strongly advocate. The reasons are outlined in Chapter 3 and IS 3.3.

Furthermore, in all NMSOP editions we generally use the symbol D for epicentral distance instead of  $\Delta$ , which is more common in older literature. However, in magnitude formulas and

## Editorial Remarks

tables of calibration values we keep the  $\Delta$  as in the original formulas. Physical units are given consistently in accordance with the International System of weights and measures (SI units). Only occasionally reference is made to equivalent American units.

Finally, since several chapters and other manual items are still under major revision, some references in the now published manual contributions to other section, figure and/or table numbers are still followed by ???. This means that these numbers still relate to the respective items in the NMSOP-1 (2009) web edition or the related printed edition of 2002. The question marks will be deleted when the finally revised numbers of these items become available with the completed version of NMSOP-2.

Potsdam/Kleinmachnow, December 2011

  
Peter Bormann (NMSOP Editor)

## References

- Aki, K., and Lee, W. H. K. (2003). Glossary of interest to earthquake and engineering seismologists. In: *International Handbook of Earthquake and Engineering Seismology, Part B*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 1793–1856.
- Bormann, P. (ed.) (2002 and 2009). Glossary. In: IASPEI New Manual of Seismological Observatory Practice. ISBN 3-9808780-0-7, *GeoForschungsZentrum Potsdam*, Vol. 2, 26 pp.; see <http://nmsop.gfz-potsdam.de>; doi:10.2312/GFZ.NMSOP\_r1\_glossary.
- IASPEI (2011) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf)
- Lee, W. H. K. (Compiler) (2009). A glossary for rotational seismology. *Bull. Seismol. Soc. Am.*, **99**(2B), 1,082–1,090.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2003). The IASPEI Standard Seismic Phase List. *Seism. Res. Lett.*, **74**, 761–772.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2011). Seismic Phase Names: IASPEI standard. In: Gupta, H. (Ed.) (2011). *Encyclopedia of Solid Earth Geophysics*, Springer, Vol. 2. 1162–1173.



## List of Reviewers

(as of January 2014)

All contributions in NMSOP-1 and 2 have been peer reviewed by international experts in the respective fields. Their names are given below in alphabetic order together with the numbers (in brackets) of the respective Chapters, Sections or auxiliary materials (DS, EX, IS, PD) reviewed by them. The Editor expresses his deep appreciation and thanks to all for spending their precious time in reviewing one or more contributions to this Manual, thus assuring the topicality and correctness of contents.

|                              |                                                  |
|------------------------------|--------------------------------------------------|
| Adams, R. D., UK             | (IS 2.1, 11; EX 3.1; EX 11.1-11.3)               |
| Babkova, Ye. A., Russia      | (11.5)                                           |
| Bard, P.-Y., France          | (14)                                             |
| Bean, C., Ireland            | (13)                                             |
| Bock, G., Germany (†)        | (IS 3.1)                                         |
| Bormann, P., Germany         | (All)                                            |
| Braunmiller, J., USA         | (IS 3.9)                                         |
| Brüstle, W. , Germany        | (3)                                              |
| Čepkuna, L. S., Russia       | (11.5);                                          |
| Choy, G. L., U.S.A.          | (3, 3.2.7.2, IS 3.1, 3.3, 3.7 )                  |
| Coyne, J., U.S.A./CTBTO      | (9)                                              |
| Dahm, T., Germany            | (7.5)                                            |
| DeSalvo, R., U.S.A.          | (IS 5.3)                                         |
| Dewey, J. W., U.S.A.         | (3; 12; DS 3.1; IS 3.2; IS 3.3; IS 3.4, IS 11.7) |
| DiGiacomo, D., Italy         | (IS 3.2; IS 3.5; 11)                             |
| Di Virgillio, Italy          | (IS 8.3)                                         |
| Donner, S., Germanz          | (IS3.9)                                          |
| Douglas, A., UK              | (11; IS 11.1; EX 3.1; EX 11.1-11.3)              |
| Elgamal, A., U.S.A.          | (7.4.6)                                          |
| Engdahl, E. R., U.S.A.       | (IS 11.1; IS 11.5)                               |
| Fogleman, V. K. A., U.S.A.   | (8)                                              |
| Forbriger, Th., Germany      | (4)                                              |
| Gabsatarova, I. P., Russia   | (11.2; 11.5)                                     |
| Graizer, V., U.S.A.          | (IS 5.3)                                         |
| Grünthal, G., Germany        | (12)                                             |
| Gusev, A. A., Russia         | (3.1-3.3 and 3.7 in NMSOP-1, 3.2.4.6)            |
| Hanka, W., Germany           | (11.4)                                           |
| Harms, J., U.S.A.            | (IS 8.3)                                         |
| Havskov, J., Norway          | (5)                                              |
| Hjortenberg, E., Denmark     | (4)                                              |
| Hort, M., Germany            | (13)                                             |
| Hutt, C. R., U.S.A.          | (7.4.4);                                         |
| Jäckel, K.-H., Germany       | (6)                                              |
| Jiménez, M.-J., Spain        | (12)                                             |
| Kanazawa, T., Japan          | (7.5)                                            |
| Kennett, B. L. N., Australia | (9; 11)                                          |
| Kilgore, B., U.S.A.          | (IS 8.3)                                         |
| Klein, F., U.S.A.            | (13)                                             |
| Kolomiyez, M. B., Russia     | (11.2; 11.5)                                     |
| Kozák, J., Czech Republic    | (IS 5.3)                                         |
| Kraft, T., Switzerland       | (IS 8.7)                                         |
| Krüger, F., Germany          | (3.2.7.1)                                        |

## List of Reviewers

|                               |                    |
|-------------------------------|--------------------|
| Kühn, D., Norway              | (IS 3.9)           |
| Lahr, J., U.S.A.              | (8)                |
| Lomax, A., France             | (3.2.8.5)          |
| Malischewski, P., Germany     | (2)                |
| Meier, Thomas, Germany        | (1)                |
| Michelini, A., Italy          | (3.2.8.5)          |
| Moschetti, M., USA            | (IS 3.3, 3.5)      |
| Mayer-Rosa, D., Switzerland   | (7)                |
| Nizhizawa, A., Japan          | (7.5)              |
| Nolet, G., France             | (3.2.6.1)          |
| Ohrnberger, M., Germany       | (11.6)             |
| Oth, A., Luxembourg           | (14)               |
| Ottmöller, L., Norway         | (10; 11; IS 3.4)   |
| Paulssen, H., Netherlands     | (9)                |
| Pirli, M., Norway/Greece      | (Glossary)         |
| Plešinger, A., Czechia        | (5)                |
| Poygina, S. G., Russia        | (11.2; 11.5)       |
| Presgrave, B., U.S.A.         | (10);              |
| Ritter, J. R. R., Germany     | (4)                |
| Saul, J., Germany             | (8; 16; IS 11.2 )  |
| Scarpa, R., Italy             | (13)               |
| Scherbaum, F., Germany        | (9)                |
| Schreiber, U., Germany        | (IS 5.3)           |
| Schweitzer, J., Norway        | (11)               |
| Shearer, P., U.S.A.           | (2)                |
| Sipkin, S. A., U.S.A.         | (1; 11)            |
| Sleemann, R., Netherlands     | (10; 16)           |
| So, E., U.S.A.                | (IS 3.5)           |
| Stammler, K., Germany         | (11.4)             |
| Steidl, J., U.S.A.            | (7.4.6)            |
| Stewart, R., U.S.A.           | (7)                |
| Storchak, Dmitry, UK/Russia   | (IS 2.1; IS 3.2, ) |
| Strauch, W., Nicaragua/SICA   | (11)               |
| Stromeyer, D., Germany        | (IS 3.1)           |
| Suhadolc, P., Italy           | (1)                |
| Taylor, L., U.S.A.            | (9)                |
| Teisseyre, R., Poland         | (IS 5.3)           |
| Theophylaktov, D. (✠), Russia | (11.2; 11.3)       |
| Tilling, R., U.S.A.           | (13)               |
| Udias, A., Spain              | (3)                |
| Veith, K., U.S.A.             | (3.1-3.3)          |
| Virgilio, A., USA             | (5.3)              |
| Wang, R., Germany             | (IS 3.1)           |
| Wassermann, J., Germany       | (IS 5.3)           |
| Webb, S., U.S.A.              | (7.5)              |
| Weber, B., Germany            | (11.4)             |
| Weichert, D. H., Canada       | (All)              |
| Wendlandt, K., Germany        | (IS 8.2)           |
| Wielandt, E., Germany         | (4)                |
| Zsiros, T., Hungary           | (12)               |
| Zweifel, P., Austria          | (7)                |

# Chapter 1

## History, Aim and Scope of the 1<sup>st</sup> and 2<sup>nd</sup> Edition of the IASPEI New Manual of Seismological Observatory Practice

(Version August 2013; DOI: 10.2312/GFZ.NMSOP-2\_ch1)

Peter Bormann

Formerly GFZ German Research Centre for Geosciences, Department 2: Physics of the Earth,  
Telegrafenberg, 14473 Potsdam, Germany; Now: E-mail: [pb65@gmx.net](mailto:pb65@gmx.net)

|            |                                                                                                  | <b>Page</b> |
|------------|--------------------------------------------------------------------------------------------------|-------------|
| <b>1.1</b> | <b>Introduction to the history of the manual editions</b>                                        | 1           |
| <b>1.2</b> | <b>Scope of the NMSOP</b>                                                                        | 6           |
| 1.2.1      | Observatory seismology: Historically and regionally changing concepts, conditions and approaches | 6           |
| 1.2.2      | Creation of awareness                                                                            | 15          |
| 1.2.2.1    | The magnitude issue                                                                              | 16          |
| 1.2.2.2    | Consequences of recent technical developments                                                    | 18          |
| 1.2.2.3    | The need for secondary phase readings                                                            | 21          |
| 1.2.2.4    | New seismic sensors, data acquisition systems and sensor calibration                             | 23          |
| 1.2.2.5    | What has to be considered when installing new seismic networks?                                  | 24          |
| <b>1.3</b> | <b>Philosophy of the NMSOP</b>                                                                   | 24          |
| <b>1.4</b> | <b>Contents of the NMSOP</b>                                                                     | 27          |
| 1.4.1      | The printed NMSOP (2002) and its electronic edition NMSOP-1 (2009)                               | 27          |
| 1.4.2      | The 2 <sup>nd</sup> electronic edition NMSOP-2 (2012-2013)                                       | 28          |
| <b>1.5</b> | <b>Outreach of the NMSOP</b>                                                                     | 29          |
|            | <b>Acknowledgments</b>                                                                           | 30          |
|            | <b>Recommended overview readings on seismology</b>                                               | 30          |
|            | <b>References</b>                                                                                | 30-32       |

### 1.1 Introduction to the history of the manual editions

Most of what we know today about the internal structure and physical properties of the Earth, and thus about the internal forces which drive plate motions and produce major geological features, has been derived from seismological data. Seismology continues to be a fundamental tool for investigating the kinematics and dynamics of geological processes at all scales. With continued advances in seismological methods we hope to better understand and assess their current status as well as the diverse related potential benefits, hazards and risks for mankind.

Geological processes neither know nor care about human boundaries. Accordingly, both the resources and the hazards can be investigated and assessed effectively only when the causative phenomena are monitored not only on a local scale, but also on a regional and global scale. Moreover, geological phenomena typically must be recorded with great precision and reliability over long time-spans corresponding to geological time-scales. Such data, which are collected in different countries by different research groups, have to be compatible in subtle ways and need to be widely exchanged and jointly analyzed in order to have any global and lasting value. This necessitates global co-operation and agreement on standards for operational procedures and data formats. Therefore, it is not surprising that the international seismological community saw the need for developing a Manual of Seismological Observatory Practice (MSOP) already many decades ago.

This matter was taken up by the scientific establishments of many nations, finally resulting, in the early 1960s, in a resolution of the United Nations Economic and Social Council (ECOSOC). In response, the Committee for the Standardization of Seismographs and Seismograms of the International Association of Seismology and Physics of the Earth's Interior (IASPEI) specified in 1963 the general requirements of such a Manual as follows:

- act as a guide for governments in setting up or running seismological networks;
- contain all necessary information on instrumentation and procedure so as to enable stations to fulfil normal international and local functions; and
- not to contain any extensive account of the aims or methods of utilizing the seismic data, as these were in the province of existing textbooks.

The first edition of the Manual of Seismological Observatory Practice was published by the International Seismological Centre (ISC) in 1970 with the financial assistance of the United Nations Educational, Scientific and Cultural Organization (UNESCO). A sustained demand for copies and suggestions for new material prompted the Commission on Practice of IASPEI in 1975 to prepare a second edition. The authors worked to achieve balance between western and Soviet/Russian traditions of seismological practice. This resulted in the 1979 version of the Manual, edited by P. L. Willmore, in which the basic duties of seismological observatories were envisaged as follows:

- maintain equipment in continuous operation, with instruments calibrated and adjusted to conform with agreed-upon standards;
- produce records which conform with necessary standards for internal use and international exchange; and
- undertake preliminary readings needed to meet the immediate requirements of data reporting.

The "final" interpretation of seismic records was considered to be an optional activity for which the Manual should provide some background material only, with no attempt to fully cover it. On the other hand, the Manual did provide more detailed guidance for observatory personnel as required for the occasional, but at those times most important, collection and classification of macroseismic observations. In general, the international team of authors "...sought to extract the most general principles from a wide range of world practice, and to outline a course of action which will be consistent with those principles."

Even as the 1979 Edition of the Manual was published, it was obvious that there existed significant regional differences in practice and that the subject as a whole was rapidly

advancing. Since this implied the need for continuous development, it was decided to produce the book in loose-leaf form and to identify chapters with descriptive code names so as to allow for easy reassembling, updating and insertion of new chapters. This useful concept was not achieved, however, and no updating or addition of new chapters happened after the 1979 edition. Yet, the general aims of the old MSOP are still quite valid, although the scope of modern practice has broadened significantly during the last few decades, and although old analog recording stations and data analysis procedures have meanwhile been replaced by the corresponding digital ones practically everywhere. But for a deeper understanding and full appreciation of these recent tremendous developments in seismological practice it has been important that, along with the IASPEI Centennial International Handbook on Earthquake and Engineering Seismology (2002), the 1979 edition of the MSOP has been made available on CD-ROM, and is now accessible, together with the electronic NMSOP editions, on the IASPEI home page (<http://www.iaspei.org/projects/NMSOP.html>) as well.

Since the last edition of the MSOP, seismology has in fact undergone a technological revolution. This is driven by cheap computer power, by the development of a new generation of seismometers and digital recording systems with very broad bandwidth and high dynamic range, by digital analysis tools that run on complex algorithms based on most recent theoretical concepts of seismic source processes, wave generation and propagation as well as signal processing and, finally, by the advent and breathtaking global progress of the Internet as an effective vehicle for rapid, large-scale data exchange and communication. This, however, made more and more sections of the 1979 Manual obsolete or irrelevant, providing no more guidance in many areas of critical importance for modern seismology.

In a workshop meeting organized in late 1993 by the International Seismological Observing Period (ISOP) in Golden, Colorado, entitled "Measurement Protocols for Routine Analysis of Digital Data", it was acknowledged that existing documents and publications are clearly inadequate to guide routine practice in the 1990s at seismological observatories acquiring digital data. It was concluded that a new edition of MSOP is needed as well as tutorials showing examples of measuring important seismological parameters (Bergman and Sipkin, 1994). This recommendation prompted the IASPEI Commission on Practice (CoP) at its meeting in Wellington, New Zealand, 1994, to establish a MSOP Working Group (WG) entrusted with the elaboration of an IASPEI New Manual of Seismological Observatory Practice (NMSOP). Peter Bormann was asked to assemble and chair the working group and to elaborate a concept on the aims, scope and approach for a new Manual.

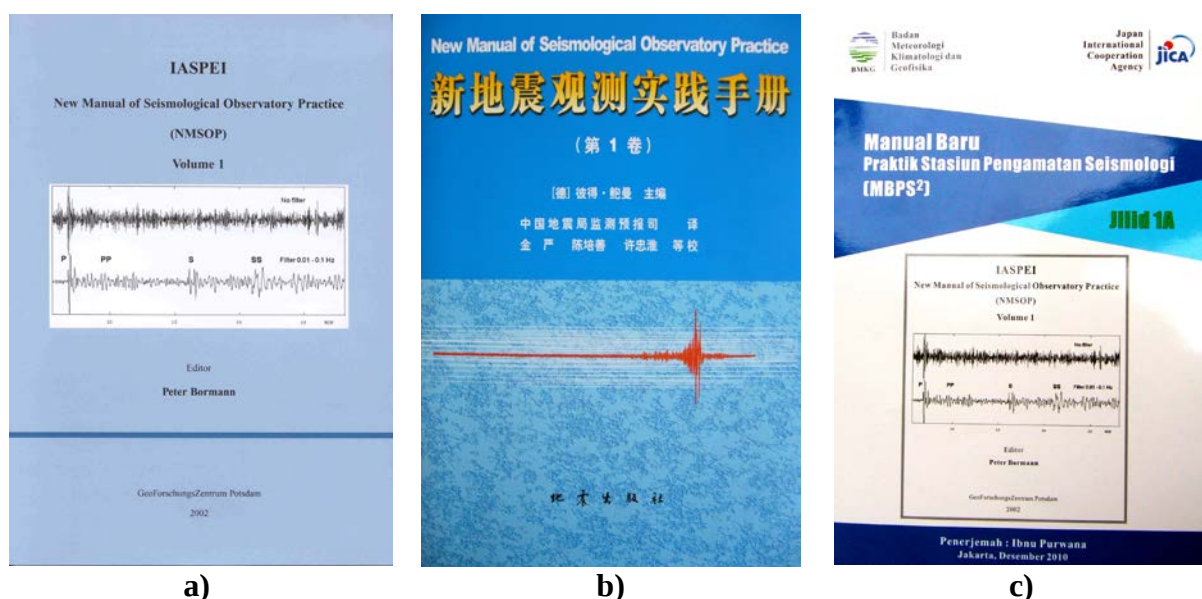
The first concept for the NMSOP was put forward at the XXIV General Assembly of the European Seismological Commission (ESC) in Athens, Greece, September 19-24, 1994 (Bormann, 1994) and in 1995 at the meeting of the IASPEI CoP on the occasion of the XXI General Assembly of the International Union of Geodesy and Geophysics (IUGG) in Boulder, Colorado. The concept was approved and both an IASPEI and an ESC Manual WG were formed. Most of the members met regularly at ESC and IASPEI Assemblies (ESC: 1996 in Reykjavík, 1998 in Tel Aviv and 2000 in Lisboa; IASPEI: 1997 in Thessaloniki, 1999 in Birmingham and 2001 in Hanoi) while others corresponded with the group and contributed to its work via the Internet. At these assemblies the Manual WG organized special workshop sessions, open to a broader public and well attended, with oral and poster presentations complemented by Internet demonstrations of the Manual web site under development. With a summary poster session at the IASPEI/IAGA meeting in Hanoi, 2001, the work of the IASPEI Manual WG was formally terminated and the WG chairman was entrusted with the final editorial work and the preparations for the publication of the Manual.



NMSOP was published in 2002 by the GeoForschungsZentrum (GFZ) Potsdam, Germany, with some financial support of IASPEI, as a hard cover loose-leaf collection of contributions in two volumes (Bormann, 2002; Fig. 1.1a). More than 2000 copies are meanwhile in use in more than 100 countries, several hundred copies alone bought and disseminated to seismological stations and member institutions by IASPEI, the Incorporated Research Institutions for Seismology (IRIS, USA) and the United Nations Comprehensive Test-Ban Treaty Organization (CTBTO). From feedback of users to the editor it is known that NMSOP has become not only a useful instruction material and guidance for the daily work of the personnel at seismological observatories and data analysis centers, but also for people working in field surveys taking along the NMSOP CD. Moreover, NMSOP Chapters and exercises are widely used by university lecturers as well as by trainers and participants in post-graduate international seismology courses.

In 2006 the Seismological Press Beijing published a Chinese translation of NMSOP as a two-volume book (Fig. 1.1b) and in 2010 the BMKG Indonesia, with support of the Japan International Co-operation Agency (JICA), issued a translation of the NMSOP Chapters 2-4 in Indonesian language (Fig. 1.1c), with more chapters still being in the process of translation.

Observatory practice relevant parts of NMSOP have been also translated into Russian by the Geophysical Survey of the Russian Academy of Sciences for use at its main seismological stations and networks and the first 4 Chapters have been translated also into Turkish by 2010.



**Fig. 1.1** First NMSOP editions in a) English, b) Chinese and c) Indonesian language.

Despite the undoubted success in the wide distribution and use of NMSOP there have been annoying hindrances in making this instructional and educational material available to those that mostly needed it, especially observatories and institutions in earthquake-prone developing countries. NMSOP had deliberately not been submitted for publication by one of the prestigious commercial publishers. To buy its 1250 pages on the international book market might then have cost a few hundred dollars and thus become unaffordable for many institutions, observatories and individuals in these countries. Therefore, the GFZ Potsdam decided to pay in advance for the pure printing cost and to assure free storing and handling of

the shipment to customers. However, it then turned out that the shipment cost to most countries for the two 5 kg volumes were twice as costly as their production, making NMSOP again not affordable for many.

It has been this sobering experience that let the Editor propose already in 2007 at the IUGG/IASPEI General Assembly in Perugia the elaboration of an amended second electronic edition, NMSOP-2, that was to be made freely available to any interested user. A detailed NMSOP-2 project plan, agreed with about two dozen potential old and new authors, was posted and adopted at the IASPEI General Assembly early 2009 in Cape Town, South Africa, and a special IASPEI symposium on NMSOP-2 was planned to be held in conjunction with the IUGG General Assembly in Melbourne, 2011. As a first step into the direction of an electronic NMSOP the GFZ library put a slightly revised version of the first NMSOP edition, NMSOP-1 (Bormann, 2009), with doi-numbers for each reviewed contribution, on the Internet. It is accessible via <http://www.iaspei.org/projects/NMSOP.html>.

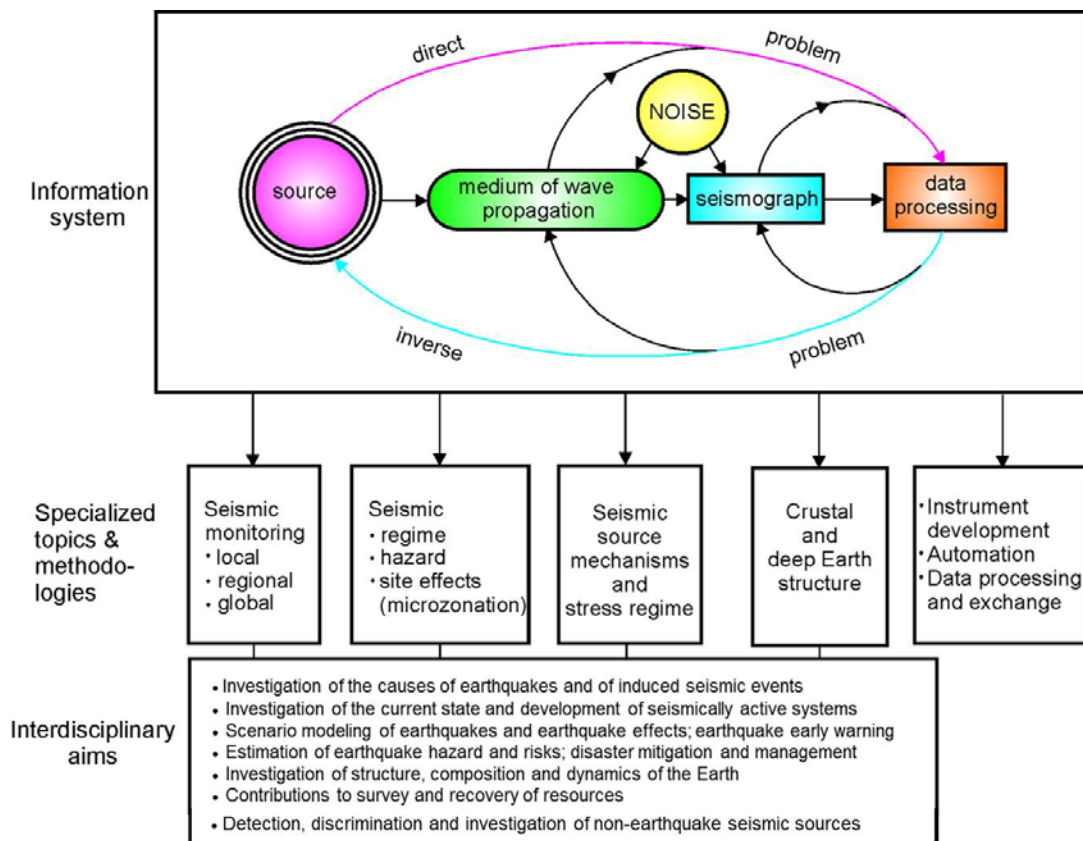
NMSOP-2 developed gradually. The first rigorously revised and largely amended Chapters, Information Sheets and Exercises of NMSOP-1 became available already in fall 2011, complemented by several new Chapters and rich auxiliary material by early 2012. On the other hand, quite a number of scheduled contributions could not be finalized in the planned time, due to both the serious chronic disease of the Editor and unforeseen priority obligations beyond the control of several committed authors. This led to the decision to open already in spring 2012 the NMSOP-2 website at <http://nmsop.gfz-potsdam.de> with the already available material. Even some contributions still under review, but earmarked differently, and amended by NMSOP-1 papers that remained either unchanged or were still under revision, were published. Thus, NMSOP-2 was “born” as – and will remain in future - a *dynamic publication*. It will develop and grow according to changing needs, new priorities, and available potentials in close collaboration between the IASPEI Commission on Seismic Observation and Interpretation (CoSOI) and the GFZ German Research Centre for Geosciences in Potsdam. The GFZ central library will assure the long-term competent maintenance of the NMSOP website. But at the same time the NMSOP is supposed to be also a mirror of the development of basic concepts, procedures and assumptions on which seismological observatory practice rests. Otherwise neither the reasons nor the appropriate ways for assuring long-term data continuity, stability and compatibility in Earth science will be understandable to the younger generation growing up in the computer age with its breathtakingly rapid changing possibilities and appealing new options.

By the end of 2013 more than 2000 Manual pages will be available for reading and downloading. They include, besides the 16 basic topical overview Chapters, plenty of complementary material such as specialized topical Information Sheets, Data Sheets, Tutorials, Exercises and Program Descriptions, Acronym explanations as well as the largest ever so far published Glossary of terms in seismology and related fields (see website cover page). Future NMSOP updates and complements will now be the main responsibility of the individual contributing authors and of IASPEI/CoSOI, in order to adapt to changing needs, newly developed procedures and gained insights. There is no more need to wait for an overall revision and new edition of the Manual before updates or complements can be made. This makes it easy to assure that with the Manual there will be in the field of seismological practice, which is usually not or only very marginally taught at universities, always a competent up-to-date educational and instructional material available. And many of its modules, animations and programs have proven to be even attractive at high-school level, suitable for promoting interdisciplinary problem understanding in general.

## 1.2 Scope of the NMSOP

### 1.2.1 Observatory seismology: Historically and regionally changing concepts, conditions and approaches

Emil Wiechert (1861-1928), professor of geophysics in Göttingen, Germany, and designer of the famous early mechanical seismographs named after him, had the following motto carved over the entrance to the seismometer house in Göttingen: “Ferne Kunde bringt Dir der schwankende Boden - deute die Zeichen.” (“The trembling rock bears tidings from afar – read the signs!”; see link to the dedication of this Manual to Emil Wiechert and Boris Galitzin). Wiechert also considered it as the supreme goal of seismology to “understand each wiggle” in a seismic record. Indeed, only then would we understand or at least have developed a reasonable model to explain the complicated system and “information chain” of seismology with its many interrelated sub-systems such as the seismic source, wave propagation through the Earth, the masking and distortion of “useful signals” by noise, as well as the influence of the seismic sensors, recorders and processing techniques on the seismogram (see Fig. 1.2).

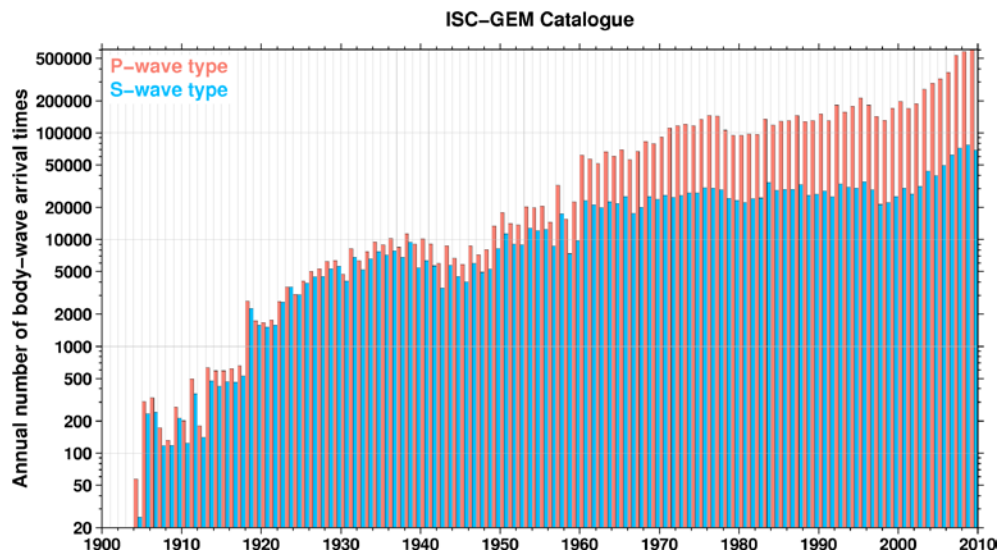


**Fig. 1.2** Scheme illustrating seismology as the analysis of a complex information system linked to a diversity of specialized and interdisciplinary tasks of research and applications.

During the early years of seismology analog seismic records have been precious unique documents for each station. Lending and shipping them to interested researches at other stations, even abroad, was a risky undertaking. Many valuable records got lost this way. Therefore, it has been a must to extract from each record as many as possible seismic phases

and other parameter data and publish them in bulletins which could be printed and distributed instead of shipping original seismograms.

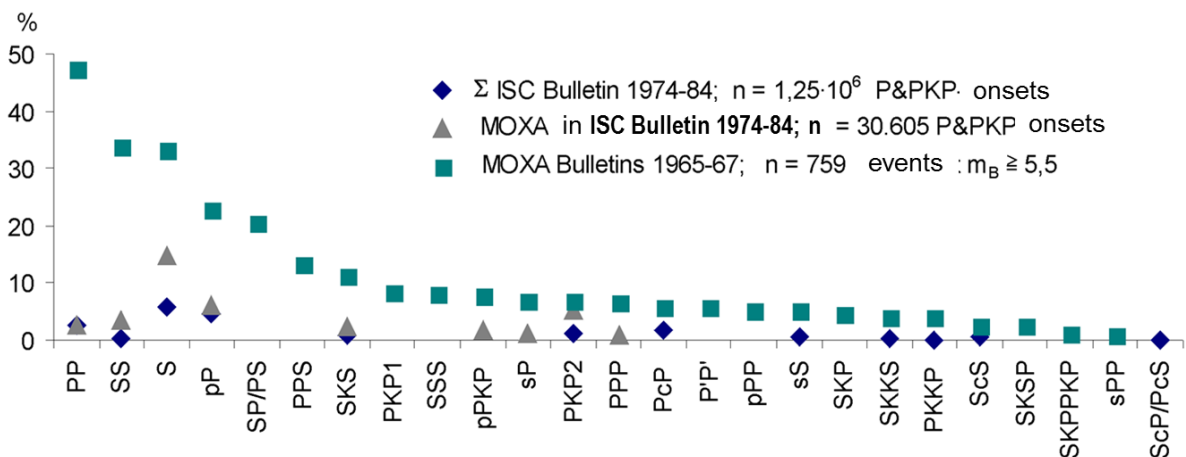
Yet, despite the tremendous progress made since Wiechert's time in understanding the most prominent features in seismic records, long-period ones in particular, we are still well short of reaching the goal he set. In fact, most operators and analysts at seismological observatories, even those who work with the most modern equipment, have not advanced much beyond the early 20<sup>th</sup> century with respect to their capability to understand and assign a proper "name" to each wiggle in a seismic record and to report their findings to international data centers. Even worse, some of the data centers are not even prepared to accept, publish and archive more detailed data reported to them if they consider them as not relevant for their duties. This has even discouraged many station and national data center operators to measure and/or report more phase data than those explicitly requested, e.g., by the NEIC of the USGS. Even pronounced seismic phases such as S shear-waves arrivals, which used to be almost as often measured and reported as P-wave arrivals before the 1960s, have often no longer been measured by many stations after the installation of the US World-Wide Standard Seismograph Network (WWSSN) in the 1960s and after the National Earthquake Information Center (NEIC) of the US Geological Survey and the International Seismological Center (ISC) in the United Kingdom (UK) resumed their operations (see Hwang and Clayton, 1991, and Figs. 1.3 -1.5). This happened irrespective of the unique importance of S-wave travel-time and amplitude readings for improving the shear-wave velocity and attenuation structure of the Earth and thus our understanding of the properties of the Earth matter.



**Fig. 1.3** Annual number of arrival times of P-waves (red) and S-waves (blue) used to produce the ISC-GEM (Global Earthquake Model project) catalogue (see Storchak et al., 2013; Bondár et al., 2013; Di Giacomo et al., 2013a,b). Different data sources were used: Before 1918 the body-wave arrival data was collected from Gutenberg's notepads (1904-1912), the International Seismological Association (ISA; 1904-1907), the British Association of the Advancement of Science (BAAS; 1913-1918) and seismological station bulletins (since 1904); between 1918 and 1963 from the International Seismological Summary (ISS) and seismological station bulletins; from 1964 until the end of 2009 from the ISC Bulletin. (Figure by courtesy of Domenico Di Giacomo, ISC, 2013).

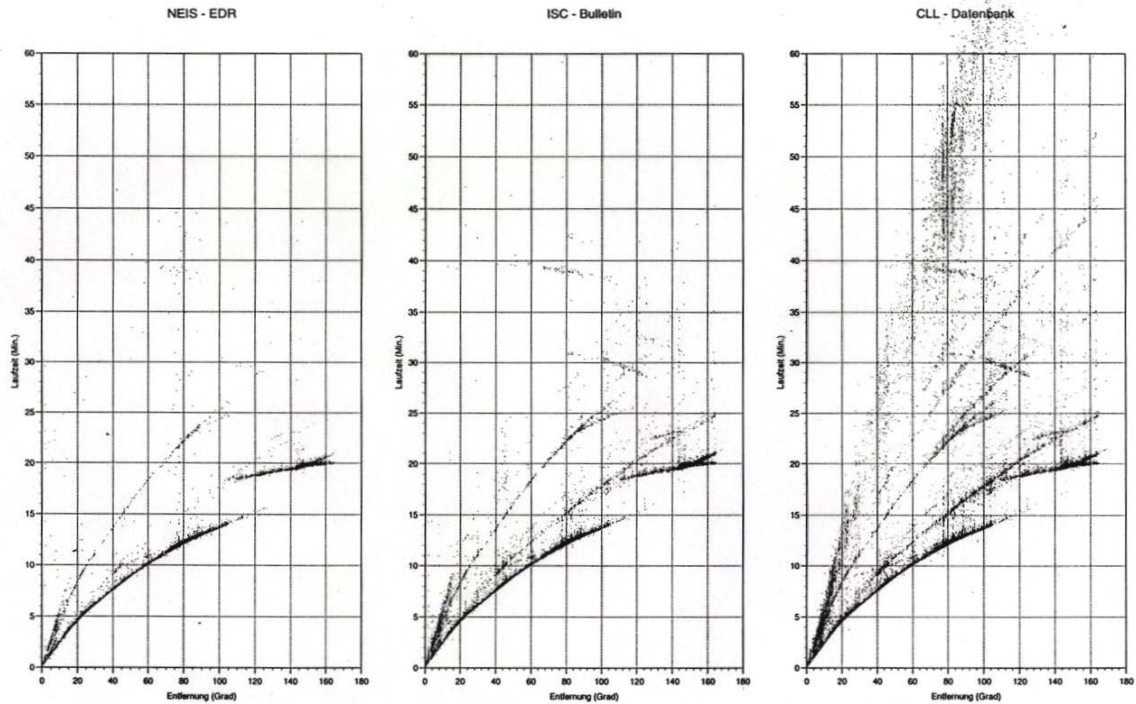
According to Fig. 1.3 the pre-1960 annual frequency of S-wave arrival times measured and documented in bulletins varied between about 55 to 100% of the P-wave readings. Since the 1960s this percentage has never been more than 40% in the ISC bulletins, dropped in recent years down to about 10%, and is even less in NEIC bulletins (PDE and EDR) (compare in Fig. 1.4 the CLL data in NEIC and ISC bulletins). There are two main reasons for this growing discrepancy. Firstly, for decades event location at both the NEIC and the ISC has been based solely on short-period P-wave first arrivals which also yielded the amplitude data for the short-period magnitude  $m_b$ , the exclusive classical body-wave magnitude calculated at the NEIC. So there seemed to be no explicit need for reporting and using of any other later body-wave onsets. Secondly, high-gain short-period narrowband seismographs introduced with the WWSSN into global standard monitoring practice allowed to increase the number of detected events by one order and even more. But for weaker events only P-waves have still a sufficient signal-to-noise ratio (SNR) in short-period filtered high-gain records, not, however, S waves located in the much noisier broadband or long-period records.

Fig. 1.4a compares the relative frequency of secondary phase onset readings to respective P and PKP readings in bulletins of the seismic station MOX in Germany. It dropped strongly from the first bulletins between 1965-67 (published by the author) to the last decade of printed bulletins (1974-84). But even from the latter data not all S, SS and PKP2 readings were included in the ISC bulletin. Even more striking is the difference between the number of secondary phase data analysed between 1990 and 1995 at the seismological observatory CLL, Germany, and the number of phase data accepted by the ISC and (much less) by the NEIC in their bulletins (Fig. 1.4b).



**Fig. 1.4a** Relative frequency of the measurement of identified secondary wave onsets as compared to P and PKP first arrivals by the author in early bulletins of station Moxa, Germany (squares), in later bulletins of station Moxa (triangles) and by the global network of seismic stations reporting to the ISC (diamonds).

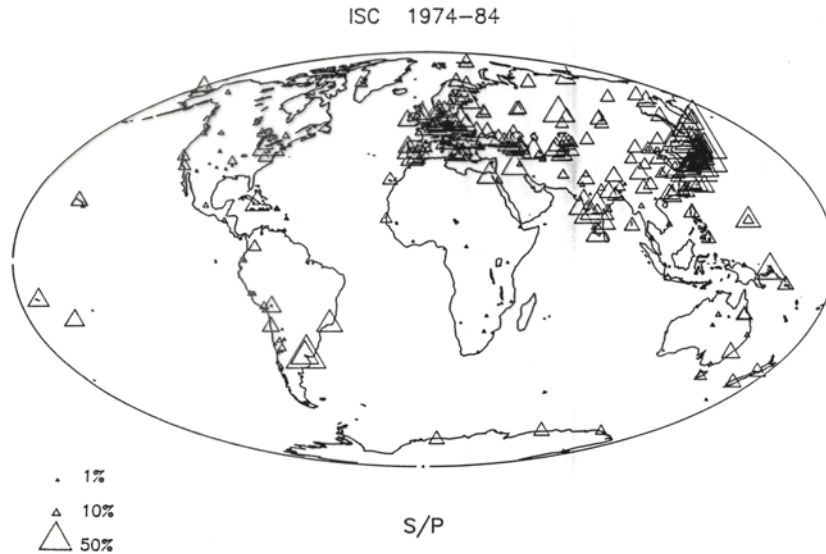




**Fig. 1.4b** Travel-time plots over epicentral distance of all seismic onsets analyzed at the observatory Collm (CLL), Germany, during the years 1990-95 (right-hand panel) and of the onsets reproduced for the same station during the same time span in the bulletins (EDR's) of the NEIC=NEIS (left-hand panel) and the ISC (central panel), respectively (Figure by courtesy of S. Wendt).

According to Bergman (1991), the first S-wave arrivals were reported on average to the ISC about twenty times less frequently than P, and other secondary phases were reported hundreds to thousands of times less often. These differences reflect operations practice at least as much as the usually reduced detectability of secondary phases. Because the NEIC did normally not use S phases in its routine processing, US station operators tended to interpret such readings as "wasted time". As a consequence, US stations reported between 1974 and 1984 also very seldom S-wave data to the ISC. Conversely, a heavy proportion of all S readings came from European and Asian stations, especially those in former Soviet Bloc countries, China and Japan, where standards of practice included an emphasis on complete reading of seismograms (Fig. 1.5).

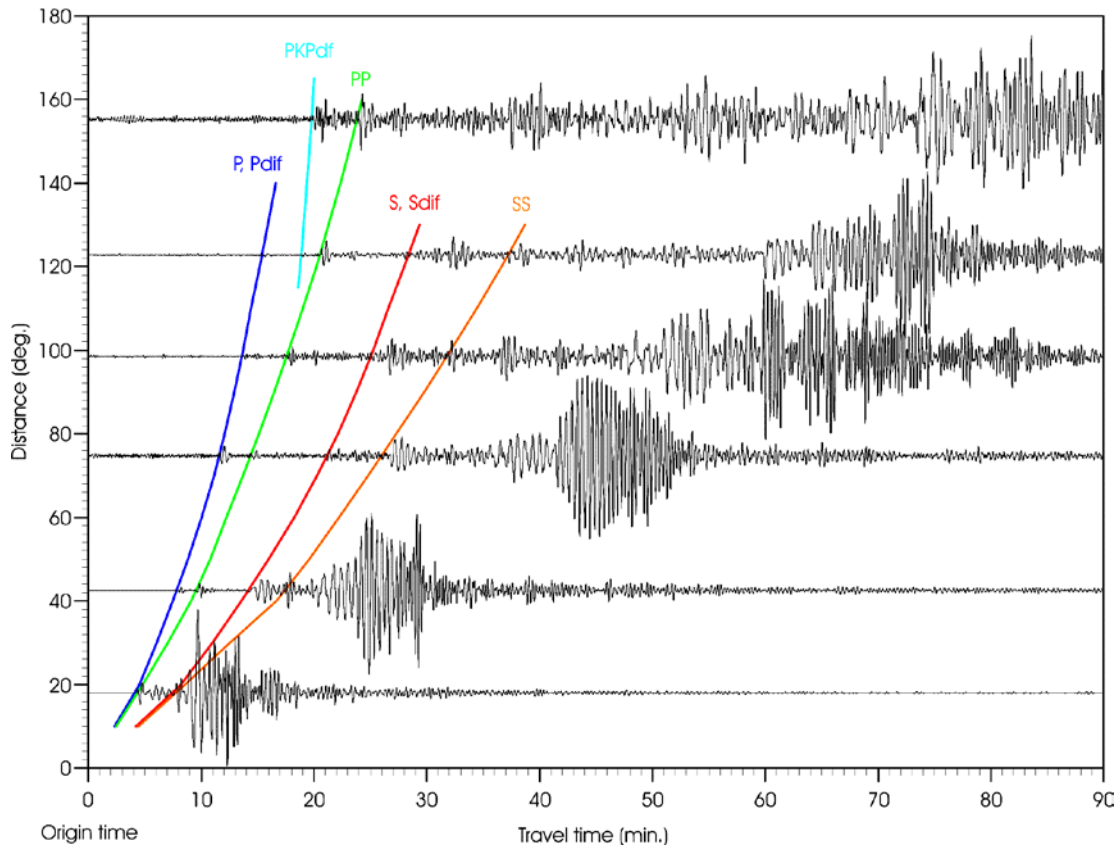




**Fig. 1.5** Relative frequency of reported S/P arrivals to the ISC by stations of the global seismic network during the decade 1974 to 1984. Note the embarrassing disregard of measuring S-wave arrivals, especially in the western Americas, Africa and Australia (Figure published by Doornbos et al., 1991).

Hwang and Clayton (1991) published a revealing analysis of the phase reports to the International Seismological Centre (ISC) by all the affiliated seismological stations of the global seismic network. Most of them, even those equipped with both short- and long-period or broadband seismographs, reported only the first P-wave onset even though later energy arrivals in teleseismic records of strong events are clearly discernible, even in fully automatic simple threshold detection procedures (see Chapter 2, Fig. 2.58), although much less frequently in short-period filtered records (Fig. 2.57). Even secondary phases with much larger amplitudes than P (e.g., Figs. 1.6 and 1.10, Fig. 2.25 in Chapter 2 and Figure 10c in DS 11.2) were usually not analyzed. And this situation has not generally changed to the better since then, rather often even worsened.

The scarcity of seismic body-wave phase readings, especially of depth phases, has strongly limited the accuracy of global seismic event locations (see Figure 7 in IS 11.1 and Engdahl et al., 1998). The same applies to the derived seismic velocity models, because the very limited number of frequently measured seismic body-wave phases is not able to sample the various lateral and depth ranges of the Earth homogeneously. This led IASPEI to launch the initiative for an International Seismological Observing Period (ISOP; see Doornbos et al., 1991). ISOP was scheduled to take place during the 1990s and aimed at training station operators in better recognizing, measuring (onset times, amplitudes and periods) and reporting relevant secondary seismic phase arrivals and providing suitable software tools for these tasks. The outcome of this initiative was supposed to assure a greatly improved and much more homogeneous future seismological database for the derivation of more detailed and better constrained structural, physical and compositional Earth models. Regrettably, after intensive preparations the project had to be cancelled even before proper take-off, because of drastic cuts in personnel at the planned ISOP co-ordination and analysis center at the NEIC. But this essential task still remains to be tackled in future. NMSOP offers a broad array of arguments, record data, seismological methods and suitable software tools to work in this direction.



**Fig. 1.6** Long-period filtered vertical-component broadband records of station CLL, Germany, of shallow earthquakes in the distance range  $18^\circ$  to  $157^\circ$ . Note the strong later longitudinal (PP) and transverse energy arrivals (S, SS) that are recognizable in the whole distance range, and the dispersed surface wave trains with large amplitudes. The record duration increases with distance (courtesy of S. Wendt, 2002).

There are several reasons for the lack of progress in the deeper understanding of seismogram analysis by station operators. Early seismic stations were mostly run or supervised by broadly educated scientists who pioneered both the technical and scientific development of these observatories. They took an immediate interest in the analysis of the data themselves and had the necessary background knowledge to do it. After World War II the installation of new seismic stations boomed and rapid technological advance required an increasing specialization. Station operators became more and more technically oriented, focusing on equipment maintenance and raw data production with a minimum of effort and interest in routine data analysis. Thus, they have tended to become separated from the more comprehensive scientific and application-oriented use of their data products in society. Also the seismological research community itself has become increasingly specialized, e.g., in conjunction with the monitoring and identification of underground nuclear tests. This trend has often caused changes in priorities and narrowed the view with respect to the kind of data and routine analysis required to better serve current scientific as well as public interest in earthquake seismology, improved hazard assessment and risk mitigation.

As a consequence, "classical" seismological observatories, as, e.g., Moxa (MOX) and Collm (CLL) in former East Germany (see Figs. 1.4a and b), belong now to an "endangered

species". They depended on a social and political system that was prepared to devote relatively large numbers of personnel and other resources to station operation and analysis, with the goal of extracting the maximum amount of information out of a limited number of recordings. One can think of this as the "observatory-centered" model for observational seismology. Beginning in the 1960s, seismology in the West favored deployment of global networks (e.g., the WWSSN - World-wide Standard Seismograph Network) with relatively less attention given to individual stations or records, making up in quantity what they gave away in quality. This "network model" of observational seismology now dominates global seismology. But a reasonable balance between quantity and quality has to be preserved. This Manual is explicitly intended to support the side of quality and completeness in the acquisition, processing, and analysis of seismic data. Currently some excesses in a pure network-centered model of observatory seismology are already questioned. In its report 2013 to CoSOI the IASPEI Working Group on Magnitude Measurements states:

*" The number of digital stations world-wide has increased dramatically, so that it is not unusual for the NEIC to have several thousand amplitude/period observations for a single, moderately strong, earthquake, .... At the NEIC, this situation has led to consideration of computing magnitudes only for a preferred subset of the overall station set, with preferred stations being selected on the basis of such criteria as geographic location, station sensitivity and reliability, or extent to which current observations would continue a long-running data set."*

Interestingly, this concept had already been proposed some 30 years ago at the time of analog seismology. It aimed at the creation of a homogeneous magnitude system (HMS) for Eurasia, based on a sub-set of first-rate stations, improved calibration functions, taking into account carefully determined station residuals for reducing drastically the data scatter in estimating event magnitudes, etc. But the proposal, although widely published (Christoskov et al, 1978; 1983; 1985; 1991), was not followed up further by the IASPEI magnitude WG and even discredited by the believers in the global network mass-data approach. Similarly, many other procedures discussed, examples presented and recommendations given in the NMSOP may appear for today's reader to be old-fashioned and off the current mainstream. This, however, should not prevent realizing their rationale and considering seriously how they might be taken into account in the modification/amendment of current procedures.

The accelerating advancement of computer capabilities during the last few decades is a strong incentive to automate more and more of the traditional tasks that need to be performed at seismological observatories. Most decision makers in seismology consider nowadays detailed seismogram parameter readings and event location nothing but time consuming and boring routine tasks, not worth paying qualified manpower for it with the exception of a clever programmer to automate these tasks. Such an assessment is apparently supported also by the facts that we can nowadays

- exchange world-wide full seismic waveform data/seismograms electronically, so that high-quality data for parameter extraction or specific research are now available (almost) everywhere (see IS 8.3);
- make use of algorithms and software for reliable and robust phase picking;
- synthesize the most important basic features in medium- to long-period seismic records;

- derive information on the source process, structure and properties of the medium from fitting synthetic waveforms to measured ones.

However, despite significant progress made in this direction, automated phase identification and parameter determination is still inferior to the results achievable by a well-trained analyst (see Chapter 3, section 3.2.3.2; Chapters 11 and 16). And for short-period records synthetics of both travel-times and waveforms look still rather different from records in real 3D media (see, e.g., Chapter 2, Fig. 2.85 and Figures 3 und 4 in IS 11.4.). Although it is likely that in future good documentations of progress made in detailed automatic seismogram interpretation can be added to the NMSOP, this second edition still heavily focuses on providing guidance, based on rich empirical experience, to: (i) the developers of software for interactive and automatic routines, (ii) to station operators and seismologists with less experience and (iii) to countries which lack specialists in the fields that should be covered by well-educated observatory personnel and application-oriented seismologists. But good knowledge of the complexity and diversity of actual body-wave waveforms is not only important for manual seismogram interpretation, but also for basic research and the development of appropriate methodologies and algorithms in general.

In any case, even in developing nations, observatory personnel has nowadays computers and software for seismogram analysis together with Internet connection for downloading the NMSOP and free shareware. During the classical analog days station analysts reading parametric data on records with fixed and limited time resolution and dynamic range, and performing their calculations with the aid of map projections, nomograms, travel-time curves, tables and logarithmic rulers, could not even dream of such comfortable tools.

Digital data, using such interactive analysis and processing tools in connection with real-time data links (see IS 8.3) and (virtual) network/array data processing (Chapters 8 and 9), now open the door for:

- Filtering standard and non-standard instrument responses;
- Phase identification using theoretical travel-times;
- Three-component processing and thus phase identification and event location based on polarization criteria;
- Hilbert transform to correct for phase distortions at caustics;
- Routine determination of spectral source parameters;
- Source parameter determination and phase identification by fitting synthetic seismograms to real seismic records;
- Frequency-wavenumber filtering and thus the application of array-processing for improved event detection, event location and phase identification;
- Automatic report generation.

Despite all these potential benefits, it is a fact that digital stations are increasingly less frequently read manually in a comprehensive way; if they are, then it happens usually only for strong events at first-order main analysis centers or at lucky stations that still have one or more well-trained analysts. Therefore, in the context of planning for ISOP, E.R. Engdahl and E. Bergmann asked already in a 1994 IASPEI talk the following question:

*“Why does detailed routine seismogram analysis discontinue when modern digital data acquisition, processing and analysis tools become available although they significantly ease and improve to achieve ISOP goals?”*

For an ISOP Pilot experiment between September 1993 and April 1994, 33 stations had been selected worldwide. The task was to identify and report secondary phase arrivals after P or PKP from records of some strong earthquakes with clear records of such secondary phases. 30% of the participating stations reported in the average less than one later onset. Only five stations, all from Eastern Europe, reported on the average four or more later onsets!

What are the reasons for such a disturbing development?

- Narrow-band high-frequency data, also from seismic arrays, lowered the detection threshold by at least 1 magnitude unit, i.e., the amount of data to be analyzed at WDCs increased by more than a factor of about 10.
- The tremendous increase of the number of stations in many seismically active countries decreased detection thresholds and increased the number of records even further.
- Nobody wants (or gets the money) to pay for more qualified manpower needed to carry out high-level routine analysis of these much improved modern data for general use. To the contrary, funds for so-called “routine tasks” are decreasing everywhere.
- Such a budget policy and lacking commitment of governments to assure sustained funding for high-quality long-term monitoring programs kill scientifically meaningful and necessary tasks.

When reading through the Manual, especially through Chapter 3 on seismic source parameters and through Chapter 11 on seismogram interpretation with their associated Data Sheets and Exercises, it becomes soon very clear that there is no simple stupid routine in measuring the relevant parameters. Even worse, the deviations from some simplified general rules are the most important ones to identify, measure and document. Rapid automatic default analysis programs can only work efficient on the basis of simplified general rules. Their development requires software specialists who usually possess, however, only a very basic understanding of the measurement parameters. On the other hand, the analysts trained over many years of analyzing tens of thousands of such parameters in all their diversity for earthquakes in very different distance, depth and magnitude ranges, do not usually have the top programming expertise needed for the development of highly efficient complex automatic data processing and analysis systems. Finally, an efficient communication and co-operation between these two types of experts as well as a comprehensive systematization and codification of complex empirical expertise, is also not a trivial task. Therefore, preserving and promoting rich empirical expertise, gained by daily tedious work with real data, preferably associated with a broad-enough seismological problem and with a good theoretical background understanding, is the indispensable precondition for any real progress in valuable automation efforts. In any event, the latter should be restricted to simplified rapid mass data handling and processing jobs and to those where rapidity counts most, and not to achieve data completeness, accuracy and novelty/peculiarity of the extracted information. The programmer and the empirically working seismologist have to be equally valued partners and the latter should not be discredited by considering him to be occupied with futile and boring routine work. NMSOP therefore focuses on the diversity and breathtaking complexity of the work with real seismological waveform data, which is usually only a minor part in most seismology text books.



In designing the Manual for a global audience, we have tried to take into account the widely varying circumstances of observatory operators worldwide. Whereas in developing countries proper education and full use of trained manpower for self-reliant development has (or should have) priority, highly advanced countries often push for the opposite, namely the advancement of automatic data acquisition and analysis. The main reasons for the latter tendency, besides the desire to limit personnel costs in high-wage countries, are:

- special requirements to assure a most rapid and objective data processing and reporting by the primary (mostly array) stations of the International Monitoring System (IMS) in the framework of the Comprehensive Nuclear-Test-Ban Treaty Organization (CTBTO) (see Chapter 15);
- coping with the huge data rates at dense digital seismic networks and arrays in areas of high seismicity;
- extracting within seconds or minutes only the most important first information from local, regional and/or global virtual network data on strong earthquakes with great risk potential (tsunami and earthquake early warning systems), which is required to assist risk mitigation, disaster management and relief operations.

Seismologists in highly industrialized countries can usually address their special concerns and requirements to national forums. Specialists in program development and automation algorithms in these countries, however, often lack the required background knowledge in seismology and/or the practical experience of operational applications in routine practice. A similar argument applies to young scientists, beginning their careers in seismological research, who often remain ignorant of the long history of operational seismology that produced the data now available for their research. A typical graduate program in seismology gives scant attention to the historical development of methods and measurement standards. This may lead either to neglect valuable older data and results, or to their incorrect interpretation and usage. In this sense, the NMSOP also aims at passing on the knowledge, experience and skills of previous generations of seismologists by addressing the educational needs of current highly specialized advanced user communities with a view to broaden both their historical and topical perspective, as well as their ability to contribute more efficiently to interdisciplinary basic research and methodological development.

## **1.2.2 Creation of awareness**

The subject of standards of practice at seismological observatories normally stays well below the active consciousness of most seismologists, yet it sometimes plays a central role in important research and policy debates. Also the awareness of the tremendous impact of recent technological advances, in both sensor developments and data communication, on the observatory and monitoring practice is not yet generally well developed. Even less developed is the understanding of the need for a much more detailed reading and reporting of secondary phases, not only of their onset times for improved event location but also of their amplitudes and periods for the determination of magnitudes and the derivation of improved attenuation models. However, both development and proper use of appropriate modern analysis tools require a profound understanding of seismic waveform data, their properties and variability. Moreover, accurate amplitude readings and reliable waveform-fitting inversion procedures require careful instrument calibration, i.e., knowledge of the transfer function and gain of the seismographs. However, there is no common standard practice yet of instrument calibration

and control. And finally, both the earthquake rupture process and the wave propagation are highly kinematic-dynamic phenomena, which are too often only explained with simplified static images and the 3D event patterns by their 2D projections. We believe that for their illustration animations are often much more suitable to create awareness of the complexity of these phenomena, easier to comprehend and much more appealing. Therefore, animations are a very valuable complementary educational tool to formal lectures, tutorials and exercises, suitable not only for academic and professional audience but also for public lectures and demonstrations at high schools. Therefore, the number and topics of animations has been significantly increased in NMSOP-2 (see IS 1.1 and IS 11.3). Yet, this is just a beginning. More such examples from other authors and on other subjects related to seismology and seismic effects are very much welcome. Authors are kindly invited to link their contributions to NMSOP-2. We will come back on some of these issues in the sections below.

### 1.2.2.1 The magnitude issue

Earthquake magnitude is one of the most widely used parameters in seismological practice, and one that is particularly subject to misunderstanding, even among seismologists. A striking example on how changing operational procedures have contaminated a valuable data set has been published by Hutton and Jones (1993). After re-examining the earthquake catalogue for southern California between 1932 and 1990 they concluded:

- ML magnitudes (in the following termed Ml with l for “local”) had not been consistently determined over that period;
- amplitudes of ground velocities recorded on Wood-Anderson instruments and thus Ml were systematically overestimated prior to 1944 compared to present reading procedures;
- in addition, changes from human to computerized estimation of Ml led to slightly lower magnitude estimates after 1975;
- these changes contributed to an *apparently higher rate of seismicity* in the 1930’s and 1940’s and a later decrease in seismicity rate which has been interpreted as being related to the subsequent 1952 Kern County ( $M_w = 7.5$ ) earthquake;
- variations in the rate of seismic activity have often been related to precursory activity prior to major earthquakes and therefore been considered suitable for earthquake prediction;
- the re-determination of Ml in the catalogue for southern California, however, *does not confirm any changes in seismicity* rate above the level of 90% significance for the time interval considered.

Similar experiences with other local and global catalogues led Habermann (1995) to state:

*“... the heterogeneity of these catalogues makes characterizing the long-term behavior of seismic regions extremely difficult and interpreting time-dependent changes in those regions hazardous at best. ... Several proposed precursory seismicity behaviors (activation and quiescence) can be caused by simple errors in the catalogues used to identify them. ... Such mistakes have the potential to undermine the relationship between the seismological community and the public we serve. They are, therefore, a serious threat to the well-being of our community.”*

Two striking examples of the consequences of disregarding procedural and nomenclature differences of teleseismic magnitude data, or even deliberately misusing them in the interest of political priorities, are given below:

Classical seismology was based on the recordings of medium-period instruments of relatively wide bandwidth such as Wiechert, Galitzin, Mainka, and Press-Ewing seismographs. Gutenberg's (1945 b and c; 1956) and Gutenberg and Richter's (1956 a and b) work on earthquake body-wave magnitude scales for teleseismic event scaling and energy determination was mainly based on records of such seismographs. Then, with the introduction of the WWSSN short-period instruments, body-wave magnitudes were determined routinely in the United States only from amplitude-measurements of these short-period narrowband records, which have better detection performance for weaker events than medium- and long-period seismographs and yield a better discrimination between earthquakes and underground nuclear explosions on the basis of the mb-Ms criterion (see 11.2.5.2 and Chapter 15). However, American seismologists calibrated their amplitude measurements with the Gutenberg-Richter Q-functions for medium-period body waves. This resulted in a systematic underestimation of the P-wave magnitudes (termed mb). In contrast, at Soviet "basic" stations, the standard instrument was the medium-period broadband Kirnos seismometer (displacement proportional between about 0.1 s to 10 s, later even 20 s. Accordingly, Russian medium-period body-wave magnitudes mB are more properly scaled to Gutenberg-Richters mB-Ms and logEs-Ms relations. Thus it happened that the corresponding global magnitude-frequency relationship logN-mB yielded a smaller number of annual  $m = 4$  events than the U.S. short-period mb data (Riznichenko, 1960). Accordingly, in the late 1950s at the Geneva talks to negotiate a nuclear test ban treaty, the US delegation assumed a much larger number of not reliably discriminated seismic events per year, since only teleseismic records were available to them. This prompted them to demand some 200 to 600 unmanned stations on Soviet territory at local and regional distances as well as on-site inspections in case of uncertain events (Gilpin, 1962). Thus, a biased magnitude-frequency assessment based on equalizing mb and mB played a significant role in the failure of these early negotiations aimed at achieving a Comprehensive Nuclear-Test-Ban Treaty (CTBT). The underground testing continued for several more decades.

In 1996 the CTBT was finally agreed upon, and signed by 71 States as of 2002. The United Nations CTBT Organization in Vienna runs an International Data Centre (IDC) that also determines body-wave magnitudes from records of the International Monitoring System (IMS). However, in the interest of the best possible discrimination between natural earthquakes and underground explosions by means of the body-wave/surface-wave magnitude ratio mb/Ms, they measure P-wave amplitudes after filtering the broadband records with a displacement frequency-response peaked around 4.5 Hz instead of around 1 Hz or 0.1 Hz. However, they calibrate their amplitude readings with a calibration function developed for 1 Hz data. Finally, they measure the maximum amplitudes for mb determination not, as recommended by IASPEI in 1977, within the whole P-wave train, up to 60 s after the first onset, but within the first 5.5 seconds after the P-wave onset. These differences in practice result in systematically smaller mb(IDC) values as compared to mb(NEIC). Although this difference is negligible for explosions, it is significant for earthquakes. The discrepancy grows with magnitude and may reach 0.5 to 1.5 magnitude units. Nonetheless, the IDC magnitudes are given the same name mb, although they sample different properties of the P-wave signal. Users who are not aware of the underlying causes and tricky procedural problems behind magnitude determination, may not realize this incompatibility of data and

come to completely different conclusions when using, e.g., the mb data of different data centers for seismic hazard assessment.

Such inconsistencies in the procedures to determine and name magnitude prompted IASPEI to set up again in 2002 a Working Group on magnitude measurements. It aimed at elaborating measurement standards for widely used magnitudes. The results are summarized in the Working Group recommendation (IASPEI, 2005 and 2013) and discussed in great detail in NMSOP-2, also with respect to the relationship between classical magnitudes and the newly proposed standards (see Chapter 3, sections 3.2.4, 3.2.5; IS 3.3 and 3.4).

But there are other important magnitudes, such as the moment magnitude  $M_w$  which is even considered to be an exclusively recommended *de facto* standard, that are determined with not yet standardized procedures using different wave types and period ranges. Accordingly, their magnitude values for identical events may differ by several tenths of magnitude units, such as for different versions of USGS moment magnitudes (see, e.g., Chapter 10, Tab. 10.5), or between authoritative Global Centroid Moment Tensor (GCMT) solutions and various rapid  $M_w$  proxy estimates that are available already within the first few minutes after origin times. Even magnitude estimates from the very first few seconds of a seismic record in earthquake early warning (EEW) schemes may underestimate the magnitude of strong to great ( $M > 8$ ) earthquakes by more than one magnitude unit. All these different types and procedures of magnitude estimation, the reasons for discrepancies and magnitude saturation, as well as the development of non-saturating near real-time magnitude procedures in the earthquake and tsunami early warning context, are discussed in detail in Chapter 3 and on specific issues also in IS 3.3 and IS 3.9 of this Manual.

### 1.2.2.2 Consequences of recent technical developments

When assembling the NMSOP we took into account that:

- modern seismic sensors (Chapter 5 and DS 5.1), in conjunction with advanced digital data acquisition (Chapter 6), transmission (Fig. 1.7, IS 8.2 and 8.6) and processing, allow event and magnitude determination (Fig. 1.8), as well as the analysis of seismic waves (Chapters 9 and 11), in a very broad frequency band with extremely high resolution within a much larger dynamic range (Fig. 1.9 and Fig. 4.7 in Chapter 4) and within a much shorter time than it was possible in the days of analog seismology;
- modern computer hardware and versatile interactive analysis software tremendously ease the task of comprehensive and accurate seismogram analysis (e.g., IS 11.6). This allows one to routinely determine also parameters which were far beyond the scope of seismogram analysis a few decades ago;
- precise time-keeping and reading is nowadays much less of a problem than it was in the pre-GPS (Global Positioning System) and pre-computer era;
- the rapid global spread of high-speed communication links largely eliminates any technical barrier to widespread data exchange of full waveform data in near-real time.

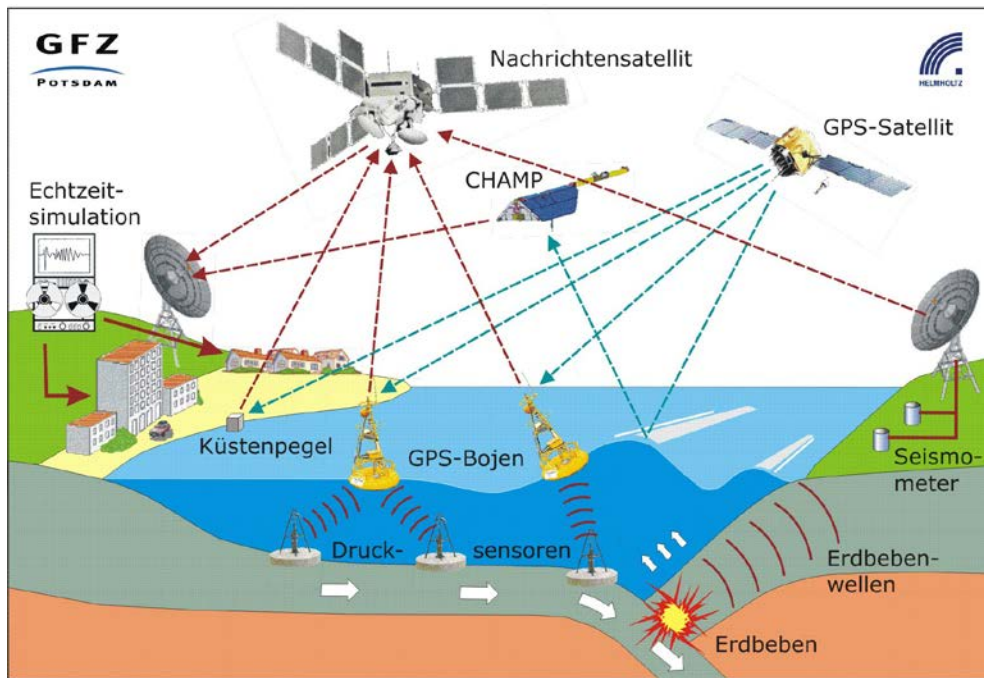


Fig. 1.7 Principal scheme of a modern ocean-spanning data acquisition and transmission system for seismological and complementary geodetic and oceanographic data as required for a tsunami early warning system (item names in German, courtesy of GFZ, Potsdam).

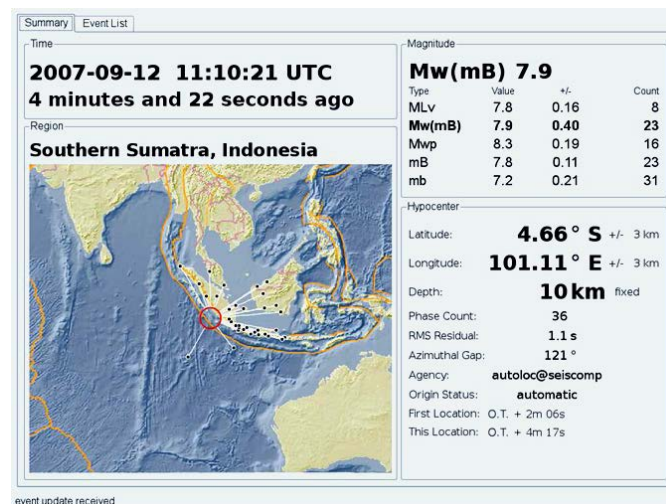
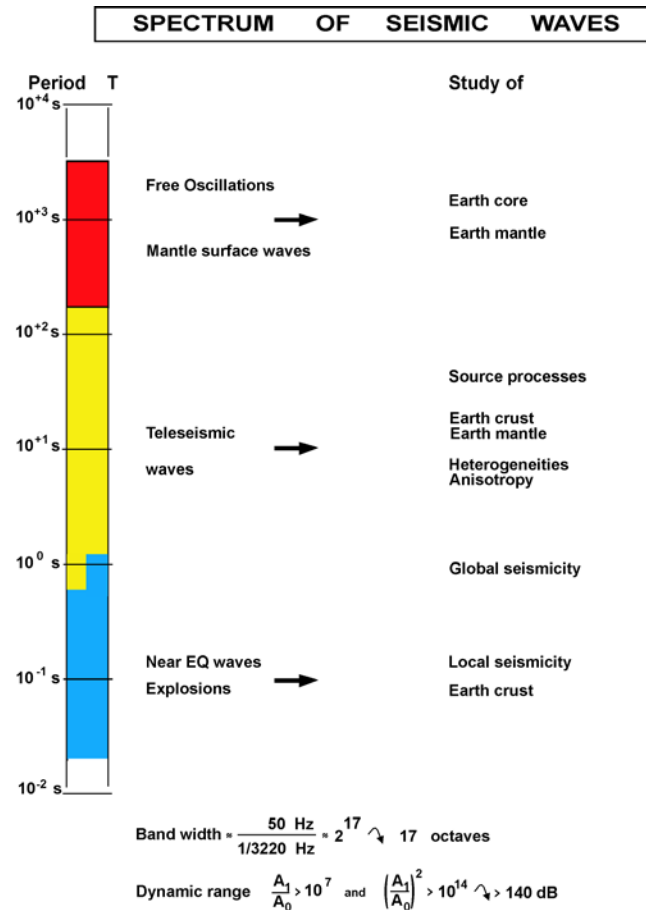


Fig. 1.8 Screen plot of the results of earthquake location and magnitude determination within the first few minutes after origin time using stations of the German-Indonesian Tsunami Early Warning System (GITEWS; <http://www.gitews.org>) and the GFZ developed SeisComp3 Software (<http://www.seiscomp3.org/>).



**Fig. 1.9** Frequency range, bandwidth and dynamic range covered by modern seismological records and related objects of research. The wavelengths of seismic waves vary, depending on their period and propagation velocity, between several meters (m) and more than 10,000 kilometers (km) and the spatial resolution of the investigated objects accordingly. The ground motion displacement amplitudes considered here range from nanometer (nm) to decimeter (dm). Continuous GPS may even measure displacements between cm and several m. Respective velocity amplitudes range between about 1 nm/s to several dm/s, however, strong-motion seismometers may even record accelerations up to about  $20 \text{ m/s}^2$  ( $\approx 2 \text{ g}$ ). When investigating with seismic methods even small-scale upper-crustal and near-surface structures or when recording induced seismic events in industrial seismology applications in conjunction with mining activities, liquid waste disposal, hydro-carbon extraction, tunnel drilling, or non-destructive material testing, even much higher frequencies than 50 Hz, up to MHz (ultrasound), may need to be recorded. This is, however, beyond the scope of the NMSOP.

At the same time, these new possibilities carry new risks:

- analysts who only use ready-made computer programs for solving a diversity of tasks, by feeding in the data and pressing the button, tend to lose a deeper understanding of the underlying model assumptions, inherent limitations and possible sources of error, and the quality of the results is sometimes judged by the attractiveness of the graphic user interface;
- Easily calculated and displayed standard deviations for all conceivable solutions often seem to indicate a reliability of the results, which is far from the truth. Therefore, an understanding of the difference between internal, computational and also model-dependent *precision* on the one hand, and *accuracy* of the solutions with



reference to reality on the other hand, has to be encouraged (see Glossary for the definition of terms);

- specialists are increasingly required to operate and properly maintain modern seismic equipment and software. They usually lack a broader geo-scientific background and thus an active interest in the use of the data themselves. This may result in a declining concern for long-term data continuity and reliability, which is the backbone for any geo-scientific observatory practice.

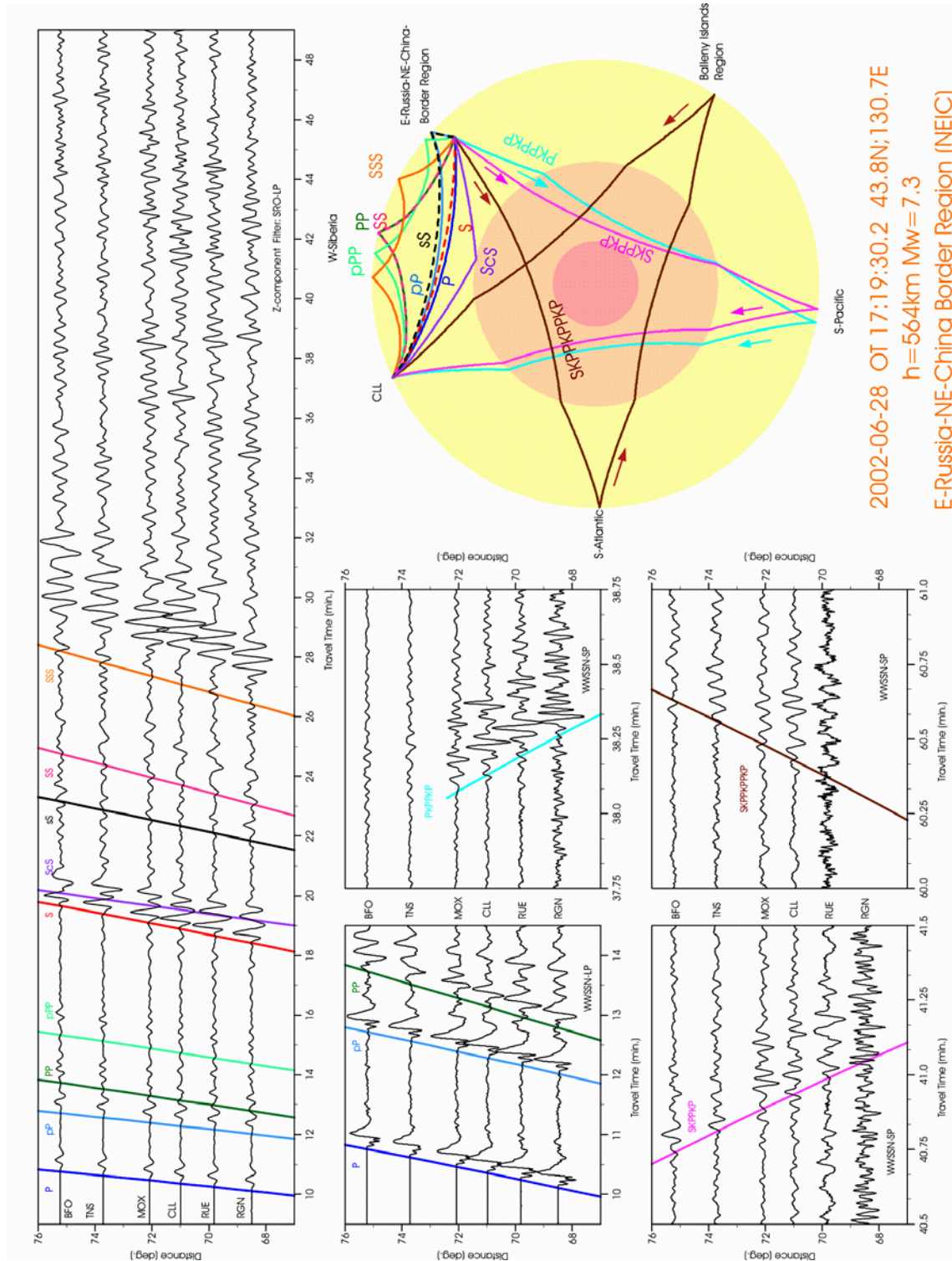
In consideration of these factors, the authors took as prime aims of the new Manual:

- to create an interdisciplinary problem understanding;
- to present a multitude of seismic waveforms depending on source type and size, epicentral distance, hypocentral depth, seismic component decomposition and axis rotation, transfer function of the seismic recordings, and with respect to the applied filters;
- to provide guidance for the selection of the most suitable station sites and the installation and shielding of modern seismic stations, vaults and sensors;
- to provide guidance for using modern tools and procedures for seismological data acquisition, processing and analysis;
- to motivate observatory personnel to overcome boring routines by developing curiosity and an active interest in the use of the data they produce both in science and society;
- to provide material for teaching, from high school to Bachelor and Master levels.

### **1.2.2.3 The need for secondary phase readings**

The currently still dominant practice of analyzing and reporting mainly first-arriving seismic phases results in conjunction with the inhomogeneous distribution of seismic sources and receivers over the globe results in a very incomplete and inhomogeneous sampling of the structural features and properties of the Earth's interior. The consequences are not only ill-constrained Earth models of low resolution, but also earthquake locations of insufficient accuracy (with mislocations up to several tens of km). This is reflected in the difficulty in understanding the seismotectonic origin of earthquakes and in the identification of the most likely places of their future occurrence. This prompted seismologists in the late 1980s (e.g., Doornbos et al., 1991) to conceive an International Seismological Observing Period (ISOP) aimed at:

- maximizing the reports of secondary phases from routine record readings aimed at improving source locations and sampling of the Earth (e.g., Fig. 1.10);
- taking best advantage, in the routine analysis, of the increasing availability of digital broadband records and easy-to-use data preprocessing and analysis software;
- improving the training of station operators and analysts;
- improving the communication, co-ordination and co-operation between the stations of the global and regional seismic networks.



**Fig. 1.10** Detailed interpretation of long-period (LP) and short-period (SP) filtered broadband records of the stations of the German Regional Seismic Network (GRSN). Note the clearly recognizable depth phases pP, pPP and sS, which are extremely important for a more accurate depth determination of the event (see Figure 11 in IS 11.1), as well as the rare but well developed multiple core phases PKPPKP, SKPPKP and SKPPKPPKP which sample very different parts of the deep Earth's interior than the direct mantle phases (courtesy of S. Wendt, 2002).

Ultimately, the ISOP plan for an international observational experiment focused on expanded reporting of secondary body wave phases collapsed in the face of entropy and inertia, but the issues raised in the ISOP project have remained important to many seismologists. The need for the NMSOP grew out of discussions within the ISOP project and has been developed in the spirit of ISOP. It is largely based on training material and practical exercises used in international training courses for station operators and analysts (see Bormann, 2000). Accordingly, Chapter 11 on Data Analysis and Seismogram Interpretation is, together with its extended annexes of seismogram examples (DS 11.1-11.4), information sheets and exercises on event location and phase identification together with related software the most extensive part of the NMSOP.

#### **1.2.2.4 New seismic sensors, data acquisition systems and sensor calibration**

NMSOP-2 presents an even more elaborated Chapter 5 on the basic theory of seismometry and the practice of instrument calibration and parameter determination. The theory is complemented by an extended Data Sheet 5.1 giving the essential parameters of two dozens of modern seismometer types, as well as by different exercises and links to freely available software for parameter determination and response calculations. NMSOP-2 introduces not only traditional inertial seismometers and strainmeters (IS 5.1), but for the first time also rotational sensors and rotation measurements (IS 5.3). In other chapters, the effects of different seismograph responses, post-record filtering or computational signal restitution on the appearance of seismograms and the reliability and reproducibility of parameter readings is demonstrated with many examples. Complementary to it, the updated Chapter 6 introduces the most recent developments and concepts of modern data acquisition systems.

Modern broadband seismographs record ground motions with a minimum of distortion and it is possible to restore true ground motion computationally with high accuracy. Seismic waveforms carry much more information about the seismic source and wave-propagation process than simple parameter readings of onset times, amplitudes and prevailing periods of seismic phases. Therefore, waveform modeling and fitting has now become a major tool both of advanced seismic research and increasingly also of routine processing and analysis. Seismic waveforms and amplitudes, however, strongly depend on the transfer function and gain of the seismograph, which must be known with high accuracy if the full potential of waveform analysis is to be exploited. Also reliable amplitude-based magnitude estimates, most of them still being determined from band-limited recordings, require accurate knowledge of the recording system's frequency-dependent magnification. Consequently, instrument parameters that control the instrument response must be known and kept stable with an accuracy of better than a few percent.

Unfortunately, at many seismic stations the seismographs have never been carefully calibrated, the actual gain and response shape is not precisely known and their stability with time is not regularly controlled. Some station operators rely on the parameters given in the data sheets of the manufacturers or those determined (possibly) by the primary installer of the stations. However, these parameters, instrumental gain in particular, are often not accurate enough. Therefore, station operators themselves should be able to carry out an independent, complete calibration of their instruments. For modern feedback-controlled broadband seismographs the basic parameters, eigenperiod and gain, are rather stable, provided that the seismometer mass is kept in the zero position. This, however, should be regularly controlled,

more frequently (e.g., every few weeks) in temporary installations and every few months in more stable permanent installations. Chapter 5 deals extensively with this issue and provides links to calibration software. Moreover, long-period seismographs are strongly influenced by the stability of the underground and installation platform. Variations in ambient temperature, air pressure fluctuations and other environmental disturbances may result in unwanted drifts and noise. Therefore, a new IS 5.4 deals specifically with the shielding and installation of broadband seismometers.

Although short-period instruments are generally considered to be much more robust and stable in their parameters, experience has shown that their eigenperiod and attenuation may change with time up to several tens percent, especially when these instruments are repeatedly deployed in temporary installations. Parameter changes of this order are not tolerable for quantitative analysis of waveform parameters. Therefore, more frequent control and absolute determination of these critical sensor parameters are strongly recommended after each re-installation.

#### **1.2.2.5 What has to be considered when installing new seismic networks?**

More and more countries now realize the importance of seismic monitoring of their territories for improved seismic hazard assessment and the development of appropriate risk-mitigation strategies. The installation and long-term operation of a self-reliant modern seismic network is quite a demanding and costly undertaking. Cost-efficiency largely depends on proper project definition, instrument and site selection based on a good knowledge of the actual seismotectonic and geographic-climatic situation, the availability of trained manpower and required infrastructure, and many other factors. Project-related funds are often available only within a limited time-window. Therefore, they are often spent quickly on high-tech hardware and *keys-in-hands* installations by foreign manufacturers without a careful site selection and proper allocation of funds for training and follow-up operation. If local people are not involved in these initial efforts and capable of using and maintaining these new facilities and data according to their potential, then the whole project might turn out to be a major investment with little or no meaningful return. These crucial practical and financial aspects are usually not discussed in any of the textbooks in seismology that mostly serve general academic education or research. Chapter 7, covering site selection, preparation and installation of seismic stations has therefore become, after adding 31 pages alone on seismic installations and observations in the marine environment, the second largest NMSOP-2 topic. Possible achievements using modern seismological networks, both physical and virtual ones, at local, regional and global scales, and their relation with respect to aperture, data processing and results to specialized seismic arrays, is extensively dealt with in Chapters 8, 9 and 14 of NMSOP-2.

### **1.3 Philosophy of the NMSOP**

The concept for the NMSOP was developed with consideration of the benefits and drawbacks of the old Willmore (1979) Manual, taking into account the technological developments and opportunities which have appeared during the last few decades, as well as the existing inequalities in scientific-technical conditions and availability of trained manpower world-wide (Bormann, 1994).

Seismological stations and observatories are currently operated by a great variety of agencies, whose staff consists of seismologists and technicians with widely varying training and interests, or with no staff at all and operated remotely from a seismological data or analysis center. They are equipped with hardware and software ranging from very traditional analog technology to highly versatile and sophisticated digital technology. While in industrialized countries the observatory personnel normally have easy access to up-to-date technologies, spare parts, infrastructure, know-how, consultancy and maintenance services, those working in developing countries are often required to do a reliable job with very modest means, without much outside assistance and usually lacking textbooks on the fundamentals of seismology or information about standard observatory procedures.

To ensure that data from observatories can be properly processed and interpreted under these diverse conditions, it is necessary to establish protocols for all aspects of observatory operation that may affect the seismological data itself. In addition, competent guidance is often required in the stages of planning, bidding, procurement, site-selection, installation of new seismic observatories and networks and the calibration of equipment so that they will later meet basic international standards for data exchange and processing in a cost-effective and efficient manner.

One drawback of the old Manual was that its chapters were organized purely according to components or tasks of observatory practice, namely:

- Organization of station networks;
- Instruments;
- Station operation;
- Record content;
- The determination of earthquake parameters;
- Reporting output;
- Macroseismic observations;
- International services.

A consequence of this structuring was that the seismological fundamentals required to understand the relevance and particulars of the various observatory tasks were sometimes referred to in various chapters and dealt with in a fragmented manner. This approach makes it difficult for observatory personnel to comprehend the interdisciplinary problems and aims behind observatory practice and to appreciate the related, often stringent requirements with respect to data quality, completeness, consistency of procedures etc. Further, this approach puts together in the same chapters basic scientific information, which is rather static, with technical aspects, which may change rapidly. This makes it difficult to keep the Manual up-to-date without frequent rewritings of entire chapters.

The IASPEI WG on the NMSOP agreed, therefore, to structure the new Manual differently:

- The body of the Manual should have a long-term character, outlining the scope, terms of reference, philosophy, basic procedures as well as the scientific-technical and social background of observatory practice. It should aim at creating the necessary awareness and sense of responsibility to meet the required standards in observatory work in the best interests of scientific progress and social service.
- The main body or backbone of the NMSOP should be structured in a didactically systematic way, introducing first the scientific-technical fundamentals underlying

each of the main components in the "information chain" (see Fig 1.2) before going on to major tasks of observatory work.

- This Manual core composed of thematic Chapters should be complemented by annexed material of complementary information, which can stand for itself. Some of these topics are too bulky or specific to be included in the body of the Manual, while others may require more frequent updating than the thematic Manual Chapters. Therefore, they should be kept separate and individualized. Some annexes give more detailed descriptions of special problems (e.g., event location or theory of source representation; seismic moment tensor and energy release calculations) others provide data about commonly-used Earth models, shareware for problem solving, seismic record examples, calibration functions for magnitude determination, widely-used sensors and their key parameters, or job-related exercises with solutions for specific observatory tasks such as phase identification, event location, magnitude estimation, fault-plane determination, complemented by a list of acronyms and a very detailed glossary of terms.
- While the printed first edition tried - for the sake of paper saving and cost reduction - to avoid any duplication throughout the Manual and therefore had an overall reference list and index, the electronic NMSOP-2 is more a compilation of stand-alone contributions. All are peer reviewed and have their own doi-number, overview listing of contents at the outset and a complete reference list at the end. They contain all essential information and illustrations required for understanding the problem or carrying out a specified task without the need to download or printout additional NMSOP material. Links and cross references to other NMSOP items or external sources of information mainly aim at stimulating further readings for the sake of justifying the statements made and/or deepening the understanding of the subject, as is the case of references in journal publications. This structuring also eases future independent upgrading of the MSOP contributions by the individual authors.

Thus, we hope to provide a new Manual which is a sufficiently complete, self-explanatory reference source ("cook and recipe book") aimed at providing awareness of the complexity of problems, basic background information, and specific instructions for the self-reliant execution of common "routine" or "pre-research" jobs by the technical and scientific staff at seismological stations, observatories, and network centers. This includes system planning, site investigation and preparation, instrument calibration, installation, shielding, data acquisition, processing and analysis, documentation and reporting to relevant national and international agencies, data centers or the public, and occasionally, also assessing and classifying earthquake damage.

The NMSOP will not cover the often highly automated procedures now in use at many international seismological data centers. These centers normally neither record nor analyze seismic records themselves, but rather use the parameters or waveforms reported to them by stations, networks or arrays. They usually have the expertise and the scientific-technical environment and international connections needed to carry out their duties effectively. Rather, the NMSOP mainly serves the needs of the majority of less experienced or too narrowly specialized operators and analysts in both developing and industrialized countries, so as to assure that all the necessary tasks within the scope and required standards for national and international data acquisition and exchange can be properly performed.



Worldwide there is no formal university education or professional training available for seismic station operators and data analysts. Observatory personnel usually acquire their training through “learning by doing”. The formal educational background of observatory personnel may be very different: Physicists, geologists, electronic or computer engineers, rarely geophysicists. Accordingly, the NMSOP tries to be comprehensible for people with different backgrounds, to stimulate their interest in interdisciplinary problems and to guide the development of their required practical skills. The method of instruction is mainly descriptive. Higher mathematics is only used where it is indispensable, e.g., in the seismometry chapter or in the information sheets about theoretical source representation (IS 3.1) and seismic moment tensor determination and decomposition (IS 3.9).

The NMSOP should, however, also be a contribution, at least in part, to public, high school and university education in the field of geosciences. Therefore, NMSOP-2 has been enriched by several new tutorials and quite many appealing animations on seismic rupture and ray propagation, event location, source radiation patterns, 3-D event-cluster presentation and its development in space and time. We hope that some of these components and practicals will be useful also for students of geophysics.

## 1.4 Contents of the NMSOP

The NMSOP has been made available in different forms:

- the first edition in 2002 as a loose-leaf collection of printed material in two volumes, complemented by a CD-ROM;
- the 1<sup>st</sup> slightly revised edition in 2009 (NMSOP-1) in electronic form as pdf-files on the Internet, accessible via <http://www.iaspei.org/projects/NMSOP.html> and <http://nmsop.gfz-potsdam.de>.
- the 2<sup>nd</sup> edition in electronic form only, accessible via the same websites and downloadable in all its components, as NMSOP-1 too. However, users with still insufficient or too slow internet connections may also request a NMSOP-2 DVD from the GFZ library ([bib@gfz-potsdam.de](mailto:bib@gfz-potsdam.de)).

### 1.4.1 The printed Manual (2002) and its electronic edition (NMSOP-1; 2009)

The IASPEI and ESC Working Groups for the NMSOP agreed on the following topical Manual chapters for the first edition (for details see List of Contents accessible via the websites <http://nmsop.gfz-potsdam.de> and <http://www.iaspei.org/projects/NMSOP.html>):

|            |                                                                                      |
|------------|--------------------------------------------------------------------------------------|
| Chapter 1: | Aim and scope of the IASPEI New Manual of Seismological Observatory Practice (NMSOP) |
| Chapter 2: | Seismic Wave Propagation and Earth Models                                            |
| Chapter 3: | Seismic Sources and Source Parameters                                                |
| Chapter 4: | Seismic Signals and Noise                                                            |
| Chapter 5: | Seismic Sensors and their Calibration                                                |
| Chapter 6: | Seismic Recording Systems                                                            |
| Chapter 7: | Site Selection, Preparation and Installation of Seismic Stations                     |
| Chapter 8: | Seismic Networks                                                                     |
| Chapter 9: | Seismic Arrays                                                                       |

- Chapter 10: Data Formats, Storage, and Exchange
- Chapter 11: Data Analysis and Seismogram Interpretation
- Chapter 12: Intensity and Intensity Scales
- Chapter 13: Volcano Seismology

These chapters form Volume 1 of the printed NMSOP and cover either the fundamental aspects of the main sub-systems of the "Information Chain of Seismology" as presented schematically in Fig. 1.2, or related specific tasks, technologies or methodologies of data acquisition, formatting, processing and application.

Volume 1 is complemented by Volume 2. The latter contains annexes in the following categories:

- **7 Datasheets (DS):** Lists of sensor parameters; record examples, travel-time curves, Earth models, calibration functions, etc.;
- **18 Information Sheets (IS):** They contain more detailed treatments of special topics or condensed summaries of special instructions/recommendations for quick orientation, present the standard nomenclature of seismic phase and magnitude names, give examples for parameter reports and bulletins, etc.;
- **14 Exercises (EX):** Practical exercises with solutions on basic observatory tasks such as event location, magnitude estimation, determination of fault-plane solutions and other source parameters, instrument calibration and response construction. For educational purposes, most of these exercises are carried out manually with very modest technical and computational means, however links are given to related software tools;
- **12 Program Descriptions (PD):** Short descriptions of essential features of freely available software for observatory practice and how to access it;
- **Miscellaneous:** A list of acronyms, an extensive index, the list of authors with complete addresses, an overall list of references for Volume 1 and a 28 page glossary of terms.

In total, NMSOP-1 comprises some 1250 pages.

### 1.4.2 The 2<sup>nd</sup> electronic edition (NMSOP-2; 2012-13)

All NMSOP-1 Chapters have been preserved in NMSOP-2, but significantly revised and/or amended. Four three Chapters have been added:

- Chapter 14: Investigation of Site Response in Urban Areas by using Earthquake Data and Seismic Noise
- Chapter 15: CTBTO: Goal, Networks, Data Analysis and Data Availability
- Chapter 16: Automated Event and Phase Identification

The 16 Chapters comprise already as many pages as the two volumes of NMSOP-1. An originally planned major new Chapter on "Seismological Contributions to Seismic Risk Mitigation", meant to become the crucial missing link to the engineering seismology, earthquake engineering, strong-motion recording, hazard and risk assessing and mitigating communities, could regrettably not be realized. It is hoped that this gap can at least partially be closed by related topical information sheets in future amendments to NMSOP-2 under a new editorship.

The number of information sheets (now 36), tutorials (5), exercises (16), and movie demonstrations (16 animations) has been greatly expanded in NMSOP-2, making for approximately another 850 pages, to which 200 pages of the currently most elaborate Glossary of terms in seismology, global and seismotectonics, engineering seismology and other related disciplines have been added. Thus, in total, NMSOP-2 will comprise about 2400 pages of material, complemented by many links to external sources.

## 1.5 Outreach of the NMSOP

The printed English version of the first edition of the NMSOP is meanwhile in use with some 2000 copies in more than 100 countries, complemented by another 2000 copies of a 2 volume book edition in Chinese and several 100 copies of partial translations into Russian, Indonesian and Turkish language, respectively. Accessibility and usage as well as the diversity of the user community will grow significantly with the two Internet editions now being available free of charge.

It is expected, therefore, that the user community of the NMSOP will not be limited to observatory personnel. Many chapters, sections, auxiliary materials and educational animations (for the latter see IS 1.1) will be of general interest to lecturers and students in seismology, geophysics or geosciences in general but also for increasingly popular courses of “seismology at schools” in many countries. Training institutions in the field of applied seismology may use any NMSOP-2 material, compiling training modules from it that are tailored to their specific requirements, provided that the data source and the individual authors of the related Manual contribution are properly cited (click on cover page: [Rights, Permissions, Acknowledgments and References to NMSOP-2](#), respectively NMSOP-1). Also, technical people from other disciplines might find (at least partially) useful pieces of information in the Manuals for their work and cooperation with seismologists such as disaster managers and risk mitigation planners. Thus NMSOP is hoped to be of long-term benefit to a rather diverse user community, including also those interested in historical approaches and results of observatory seismology.

Permanent maintenance and regular updating of NMSOP-2 will assure its wide and long-term outreach, in keeping with continuing developments and changing requirements. This will be a permanent duty of the IASPEI Commission on Seismological Observation and Interpretation (CoSOI) and its relevant Working Groups. The library of the GFZ German Research Centre of Geosciences will serve as the long-term host of the NMSOP website, assuring its presentation and maintenance according to international bibliographic and publishing standards, including the handling of all rights and permission affairs on behalf of IASPEI. Accordingly, the acting CoSOI chairman (see <http://www.iaspei.org/>) and the GFZ chief librarian, presently Mr. R. Bertelmann ([rab@gfz-potsdam.de](mailto:rab@gfz-potsdam.de)), are the two main focal points of contact in this grand international cooperative project. Related general questions and proposals should be addressed to them, in contrast to specific topical NMSOP-item inquiries and suggestions which should be addressed to the respective lead author.

## Acknowledgments

My thanks go to former members of the IASPEI Manual Working Group and all other authors who have actively contributed to the development of the Manual concept and the currently available drafts. Also acknowledged are the valuable comments and suggestions received on the original draft Chapter 1 from B. L. N. Kennett and S. A. Sipkin, which has now been amended and updated for the 2<sup>nd</sup> electronic edition NMSOP-2. The author is very grateful to P. Suhadolc and T. Meier for their very careful reviews of this new extended version of Chapter 1 and their valuable suggestions for improvement both content and language-wise.

## Recommended overview readings on seismology

Aki and Richards (1980, 2002)  
Båth (1979)  
Bolt (1982, 1993, 1999)  
Havskov and Alguacil (2006)  
Havskov and Ottemöller (2010)  
Kennett (2001 and 2002)  
Kulháněk (1990, 2002)  
Lay and Wallace (1995)  
Lilie (1998)  
Scherbaum (2007)  
Shearer (1999)  
Stein and Wysession (2003)  
Udias (1999, 2002)  
Willmore (1979)

## References

- Aki, K., and Richards, P. G. (1980). *Quantitative seismology – theory and methods*, Volume II, Freeman and Company, ISBN 0-7167-1059-5(v. 2), 609-625.
- Aki, K., and Richards, P. G. (2002). *Quantitative seismology*. 2<sup>nd</sup> ed., ISBN 0-935702-96-2, University Science Books, Sausalito, CA., xvii + 700 pp.
- Aki, K., and Richards, P. G. (2009). *Quantitative seismology*. 2<sup>nd</sup> ed., corr. Print., Sausalito, Calif., Univ. Science Books, 2009, XVIII + 700 pp., ISBN 978-1-891389-63-4 (see [http://www.ldeo.columbia.edu/~richards/Aki\\_Richards.html](http://www.ldeo.columbia.edu/~richards/Aki_Richards.html)).
- Båth, M. (1979). Introduction to seismology. *Birkhauser Verlag*, Basel, 428 pp.
- Bergman, E. A. (1991). International Seismological Observing Period: Preliminary Science Plan. Appendix A: Global Network Phase Data Reported to the ISC. *USGS Working Document*, 13 figs., 3 pp.
- Bergman, E. A. (1994). *ISOP circular letter* with attachments, April 10, 1994, 14 pp.
- Bergmann, E. A., and Sipkin, S. A. (1994). Measurement protocols for routine analysis of digital seismic data. January 11, 1994, 8 pp.
- Bolt, B. A. (1982). Inside the Earth: Evidence from earthquakes. *W. H. Freeman and Comp.*
- Bolt, B. A. (1993). Earthquakes and geological discovery. The Scientific American Library, *W. H. Freeman and Company*, New York, 220 pp.

- Bolt, B. A. (1999). Earthquakes. 4<sup>th</sup> edition, W. H. Freeman NY, 366 pp.
- Bormann, P. (1994). Concept for a new “Manual of Seismological Observatory Practice”. European Seismological Commission, XXIV General Assembly 1994, Sept. 19-24, *Proceedings and Activity Report 1992-1994*, Vol. II, Athens, Greece, Univ. of Athens, Faculty of Sciences, Dept. of Geophysics and Geothermy, Athens (1994), 698-707.
- Bormann, P. (2000). International Training Courses on Seismology and Seismic Risk Assessment. *Seism. Res. Lett.*, **71**, 5, 499-509.
- Bormann, P. (ed.) (2002). *IASPEI New Manual of Seismological Observatory Practice*. ISBN 3-9808780-0-7, GeoForschungsZentrum Potsdam, Vol. 1 and 2, 1250 pp.
- Bormann, P. (ed.) (2009 ). *IASPEI New Manual of Seismological Observatory Practice (NMSOP-1; electronic edition)*. GeoForschungsZentrum Potsdam, Vol. 1 and 2, 1250 pp.; <http://www.iaspei.org/projects/NMSOP.html>.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1978). Homogeneous magnitude system of the Eurasian continent. *Tectonophysics*, **49**, 131-138.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1983). Homogeneous magnitude system of the Eurasian continent: S and L waves. *World Data Center A for Solid Earth*, Boulder Report SE-34.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1985). Magnitude calibration functions for a multidimensional homogeneous system of reference stations. *Tectonophysics*, **118**, 213-226.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1991). Homogeneous magnitude system with unified level for usage in seismological practice. *Studia geoph. et geod.*, **35**, 221-233.
- Doornbos, D. J., Engdahl, E. R., Jordan, T. H., and Bergman, E. A. (1991). International Seismological Observing Period; Preliminary Science Plan. Revised version. *USGS Working Document*, 11 figs., 21 pp.
- Engdahl, E. R., Van der Hilst, R. D., and Buland, R. P. (1998). Global teleseismic earthquake relocation with improved travel times and procedures for depth determination. *Bull. Seism. Soc. Am.*, **88**, 722-743.
- Gilpin, R. (1962). American scientists and nuclear weapons policy. *Princeton Univ. Press*.
- Gutenberg, B. (1945b). Amplitudes of P, PP, and S and magnitude of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 57-69.
- Gutenberg, B. (1945c). Magnitude determination of deep-focus earthquakes. *Bull. Seism. Soc. Am.*, **35**, 117-130.
- Gutenberg, B. (1956). The energy of earthquakes. *Q. J. Geol. Soc. London*, **112**, 1-14.
- Gutenberg, B., and Richter, C. F. (1956a). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Gutenberg, B., and Richter, C. F. (1956b). Earthquake magnitude, intensity, energy and acceleration. *Bull. Seism. Soc. Am.*, **46**, 105-145.
- Habermann, R. E. (1995). Opinion. *Seism. Res. Lett.*, **66**, 5, 3.
- Havskov, J., and Alguacil, G. (2006). Instrumentation in earthquake seismology. *Springer*, Berlin, 2<sup>nd</sup> edition, 358 pp. plus CD.
- Havskov, J., and Ottemöller, L. (2010). *Routine data processing in earthquake seismology*. Springer, 347 pp.

- Hutton, L. K., and Jones, L. M. (1993). Local magnitudes and apparent variations in seismicity rates in Southern California. *Bull. Seism. Soc. Am.*, **83**, 2, 313-329.
- Hwang, L. J., and Clayton, R. W. (1991). A station catalog of ISC arrivals: Seismic station histories and station residuals. *U.S. Geological Service Open-File Report* 91-295.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf)
- IASPEI (2013). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)
- Kennett, B. L. N. (2001). The seismic wavefield. Volume I: Introduction and theoretical development. Cambridge University Press, Cambridge, x + 370 pp.
- Kennett, B. L. N. (2002). The seismic wavefield. Volume II: Interpretation of seismograms on regional and global scales. *Cambridge University Press*, Cambridge, x + 534 pp, ISBN 0-521-00665-1, 163-170.
- Kulhánek, O. (1990). Anatomy of seismograms. *Developments in Solid Earth Geophysics* **18**, Elsevier, Amsterdam, 78 pp.
- Kulhánek, O. (2002). The structure and interpretation of seismograms. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). International Handbook of Earthquake and Engineering Seismology, Part A. *Academic Press, Amsterdam*, 333-348.
- Lay, T., and Wallace, T. C. (1995). Modern global seismology. ISBN 0-12-732870-X, *Academic Press*, 521 pp.
- Lillie, R. J. (1999). Whole Earth Geophysics. *Prentice-Hall Inc.*, New Jersey, ISBN 0-13-4990517-2, 361 pp.
- Riznichenko, Yu. V. (1960). About the magnitude of underground nuclear explosions (in Russian). *Trudy Inst. Fiziki Zemli*, **182**, 15, 53-87.
- Scherbaum, F. (2007). *Of poles and zeros; Fundamentals of Digital Seismometry*. 3<sup>rd</sup> revised edition, Springer Verlag, Berlin und Heidelberg.
- Shearer, P. M. (1999). Introduction to seismology. Cambridge University Press, 260 pp.
- Stein, S., and Wysession, M. (2003). *An introduction to seismology, earthquakes and Earth structure*. Blackwell Publishing Ltd., Oxford (UK), 498 pp.
- Udías, A. (1999). Principles of Seismology. *Cambridge University Press*, United Kingdom, 475 pp.
- Udías, A. (2002). Theoretical seismology: An introduction. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). International Handbook of Earthquake and Engineering Seismology, Part A. *Academic Press, Amsterdam*, 81-102.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp., <http://www.iaspei.org/projects/NMSOP.html>.



| Topic   | Animations as educational complements to NMSOP-2                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|---------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p><b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> <p><b>Siegfried Wendt</b>, Geophysical Observatory Collm, University of Leipzig, D-04779 Wermsdorf, Germany, E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a></p> <p><b>Ute Starke</b>, Computer Center, University of Leipzig, Augustusplatz 10-11, 04109 Leipzig; E-mail: <a href="mailto:starke@rz.uni-leipzig.de">starke@rz.uni-leipzig.de</a></p> |
| Version | November 2012; DOI: 10.2312/GFZ.NMSOP-2_IS_1.1                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |

|                                                                                    | page      |
|------------------------------------------------------------------------------------|-----------|
| <b>1 Introduction</b>                                                              | <b>1</b>  |
| <b>2 Examples</b>                                                                  | <b>2</b>  |
| 2.1 Propagation of seismic rays through the Earth and formation of seismic records | 2         |
| 2.2 Wave propagation, travel-time curves and location of seismic events            | 6         |
| 2.3 Presentation of earthquake sequences in their space-time-magnitude development | 9         |
| 2.4 Radiation patterns of earthquake fault mechanisms                              | 12        |
| 2.5 Rupture tracking of the great 2004 Mw9.3 Sumatra-Andaman earthquake            | 18        |
| 2.6 Rupture tracking of the great 2010 Mw8.8 Chile earthquake                      | 19        |
| <b>Acknowledgments</b>                                                             | <b>20</b> |
| <b>References</b>                                                                  | <b>20</b> |

## 1 Introduction

NMSOP-2 has been amended by many new materials, including demonstrations and animations to topical chapters, exercises, information sheets etc. Animations visualize in an appealing and easy comprehensible way key aspects outlined in the main texts. They are made freely available for viewing and downloading. These complementary materials are suitable for being used independently of NMSOP as a whole as educational material at different levels, e.g., for “seismology at schools”, which enjoys currently growing interest and support in many countries, in undergraduate university or on-the job training courses. Animation have proven to be useful and well received also by the audience in public lectures about Earth science including topics such as the nature and location of earthquakes by means of seismic recordings, the propagation of seismic waves through the Earth and their use for investigating the Earth structure. This is illustrated below by way of a several examples. IS 1.1 itself can be used as an introductory tutorial to the animations. With a view to a broad user spectrum we aim at a largely non-technical language and the introduction of all basic technical terms so that both this text and the animations can be understood also by interested pupil and layman.

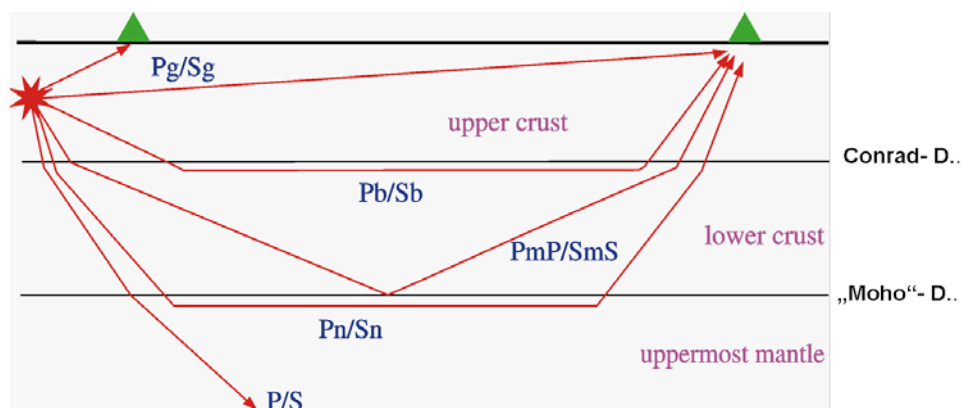
The tutorials, demos and animations currently offered in NMSOP-2 are just a beginning. We sincerely hope that they will be further upgraded and significantly amended in future by additional ones on other topics, satisfying both specialized as well broader interests. Therefore, the Editor (first author) invites potential contributors to contact him and to make their related products available to the broad NMSOP user community. He will in turn arrange that these new contributions will be highlighted in and properly linked with the topically related NMSOP main texts, thus assuring their widest possible propagation and use.

## 2 Examples

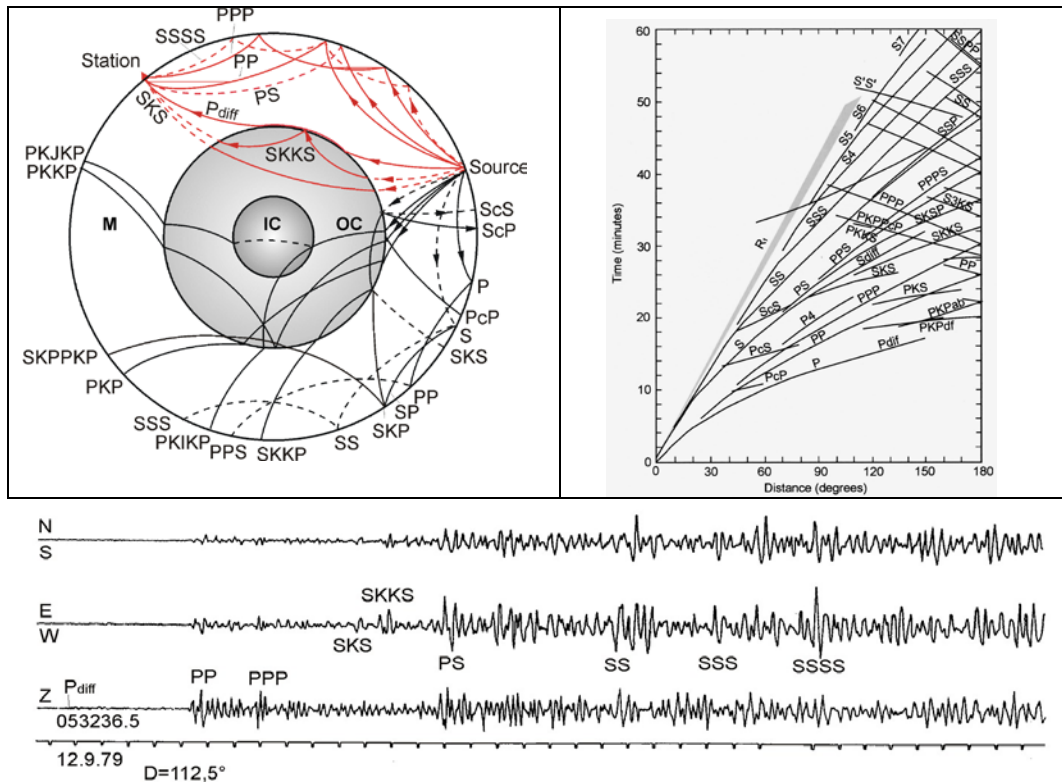
### 2.1 Propagation of seismic rays through the Earth and the formation of seismic records

NMSOP-2 offers 10 animations which show the propagation of seismic rays of different seismic phases through the Earth as well as the formation of seismic records of these waves from earthquakes at different depth (from 10 to 600 km) at seismic stations of the German Regional Seismic Network (GRSN) or local networks at distances ranging from  $0.1^\circ$  to  $167^\circ$  (i.e., from about 10 km to over 18,000 km). These phases relate to both direct longitudinal (P) and transverse (S) body waves (see Figure 13) as well their (often multiply) reflected and/or converted versions. Rays are refracted, reflected and/or converted when interacting with velocity and density discontinuities or gradients in the Earth crust (e.g., the Conrad and Mohorovičić-discontinuity), or in the deeper Earth, such as the core-mantle boundary (see Figures 1 and 2).

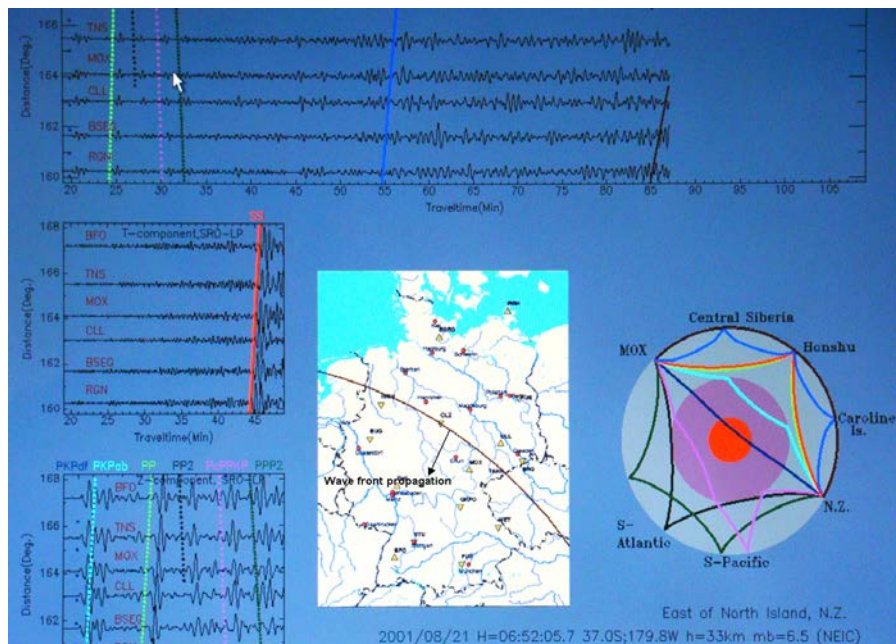
Figure 3 shows a screen-shot of one of the seismic ray tracing and record animations for an earthquake in New Zealand, recorded at station MOX and other stations of the German Regional Seismic Network (GRSN) at epicentre distances between  $160^\circ$  and  $167^\circ$ . On the lower right a simplified Earth model with the mantle in turquoise, the outer core in lilac and the inner core in red. The generally gradual increase of seismic wave velocity with depth in the mantle causes the bending of the seismic rays and the strong decrease of wave velocity at the core-mantle boundary (CMB) from about 13 km/s in the lowermost mantle to 8 km/s in the uppermost core a pronounced refraction of the P waves into the core (see the turquoise ray). However, for rays that incident on the CMB at near vertical angles the refraction is negligible (see the dark blue ray through the inner core). Thus, the figure illustrates the dependence of the refraction angle on both the gradual (in the mantle) or discontinuous (CMB) change in velocity as well as on the incidence angle of seismic rays with respect to the velocity change with depth. And Figure 4 shows the currently best 1-D model AK 135 of velocity, density and attenuation related changes in the Earth according to Kennett et al. (1995) and Montagner and Kennett (1996). Such models have been derived by analyzing empirically determined travel-time curves of P, S and other seismic body waves as well as dispersion curves of surface waves.



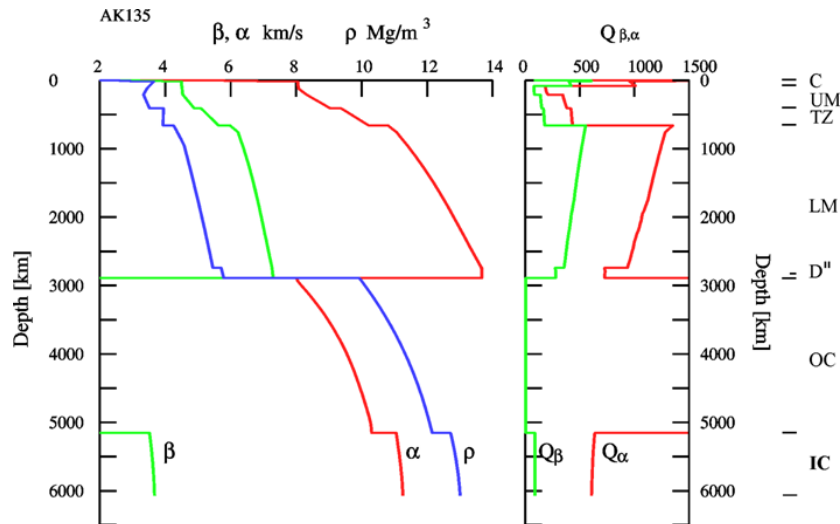
**Figure 1** Simplified model of the crust showing ray traces of the main “crustal phases” observed in records of earthquakes at distances up to about 1000 km. The short term “Moho” stands for Mohorovičić-discontinuity. It is the boundary between the Earth crust and the Earth mantle, characterized by a strong increase in P-wave velocity from about 6.5 km/s to 8 km/s).



**Figure 2** **Top left:** Seismic rays through the Earth mantle (M), outer core (OC) and inner core (IC) with their respective phase symbols according to international nomenclature. Full lines: P-wave rays; broken lines: S-wave rays. **Top-right:** Related travel-time curves of these phases. **Red rays** relate to the seismic phases identified in a 3-component broadband seismogram recorded at station MOX, Germany (**bottom**), from an earthquake at a distance of 112.5° (compiled from various Figures in Chapter 2).



**Figure 3** Screenshot of a moment towards the end of the first animation example, based on seismic recordings of the Mw = 7.1 New Zealand earthquake of 21 August 2001 at stations of the German Regional Seismic Network (GRSN). For more explanation see text.

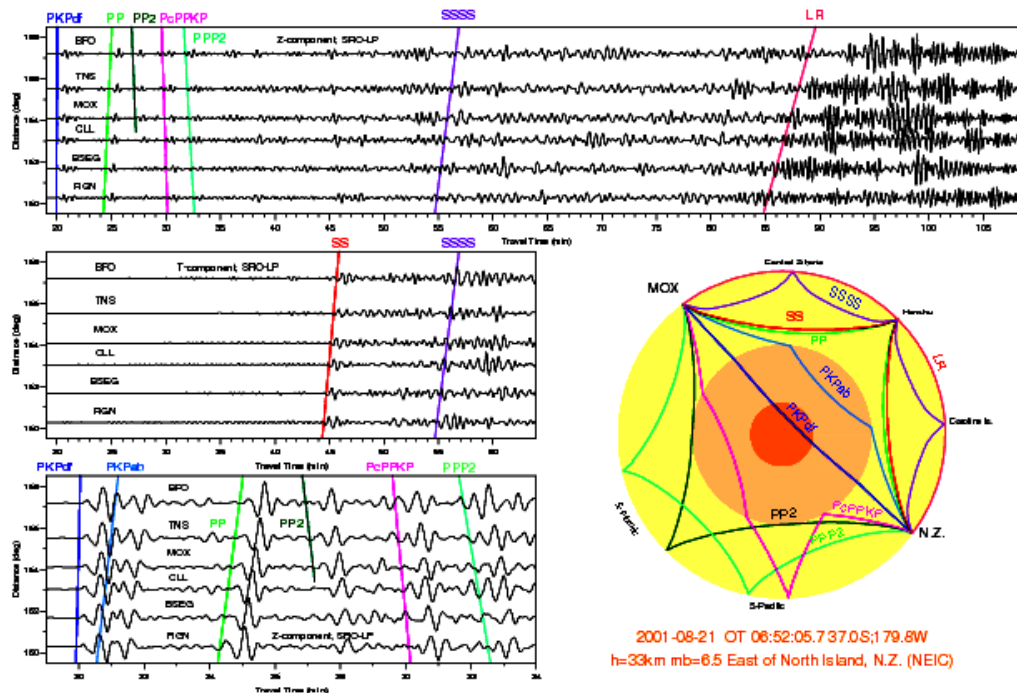


**Figure 4** The 1-D Earth model AK135. Seismic wave speeds according to Kennett et al., 1995), quality parameter  $Q$  and density according to Montagner and Kennett (1996).  $\alpha = v_p$  and  $\beta = v_s$  are the P- and S-wave velocity, respectively;  $\rho$  - density,  $Q_\alpha$  and  $Q_\beta = Q_\mu$  the “quality factor”  $Q$  for P and S waves. Note that wave attenuation is proportional to  $1/Q$ . The abbreviations on the outermost right stand, within the marked depth ranges, for: C – crust, UM – upper mantle, TZ – transition zone, LM – lower mantle, D'' - gradient layer, OC – outer core, IC – inner core. (Copy of part of Fig. 2.79 in Chapter 2).

The ray-propagation cartoon in Figure 3, and with better resolution Figure 5, shows that seismic rays, the deeper they penetrated into the Earth, the steeper they return back to the Earth surface. And the steeper they come to the surface the shorter are the time differences at which the respective seismic waves will arrive at different seismic stations on the surface. But the shorter the difference between the recorded phase arrivals the faster this wave - **apparently** - seems to travel along the surface. This apparent horizontal velocity  $v_{app}$  of the approaching wave front (see the bended curve in the middle insert of Figure 3 with the arrow in the direction of propagation) should not be mistaken as being the true wave velocity.  $v_{app}$  will become infinite for vertical incidence although the real velocity of P and S near to the Earth surface is not larger than about 5 km/s and 3 km/s, respectively. The inverse of  $v_{app}$  is called *slowness*  $= 1/v_{app}$ . The animation will show that surface-waves, travelling along the Earth surface (bended brown ray) with less than 3-4 km/s, are indeed the slowest waves of all. Also note that seismic rays do not travel only the shortest way from the source towards the recording stations but also the longer one. Accordingly, for the same type of wave, the respective long-path wavefront arrives later and from the opposite direction with *negative slowness*, i.e., they arrive first at the most distant stations and later at the nearer ones. This results in travel-time branches of different tilt (see the colored lines on the record plots of Figure 5 which connect the arrival times of the respective seismic phases.)

In the animation three kinds of seismic records are “written” as soon as the seismic waves of different phases, represented by their differently colored seismic rays, arrive at the stations. All three have been long-period filtered with the SRO-LP response which emphasizes periods around 20 s. The uppermost record section in Figure 3 (and more clearly in Figure 5) shows vertical-component records, the middle section transverse-component records and the lowermost

record section shows enlarged in amplitude and with better time resolution the first 15 min of the uppermost record. The T-component records only shear-wave motions in the horizontal plane.



**Figure 5** A modified version of Figure 3 with cleaner record and ray sections. For explanations see text.

By comparing these different records one can conclude:

- The first arriving longitudinal PKPdf waves (dark blue and steeply incident ray travelling through the inner Earth core), which oscillate in the direction of ray propagation, are best recorded on short-period vertical component records of high magnification. The same applies to the surface reflections of P (PP for the shorter way and PP2 for the longer opposite way of travel) and to PcPPKP.
- In contrast to P waves, shear waves, which oscillate perpendicular to the direction of propagation, can travel only through the Earth mantle and crust but not through the liquid outer core. They are best recorded in long-period horizontal component records. And if they contain only horizontally oscillating SH wave energy then they can be seen only in T-component records and not at all in the vertical component records. This applies to our record example to the SS phase (see the red ray and travel-time branch).
- If these arrival times increase with distance of the station to the earthquake than the seismic wave fronts arrive in our case from NNE (as for PKPdf, PP and SS), but if they decrease with distance then they arrive from the SSW (as for PP2, PcPPKP and PPP2).

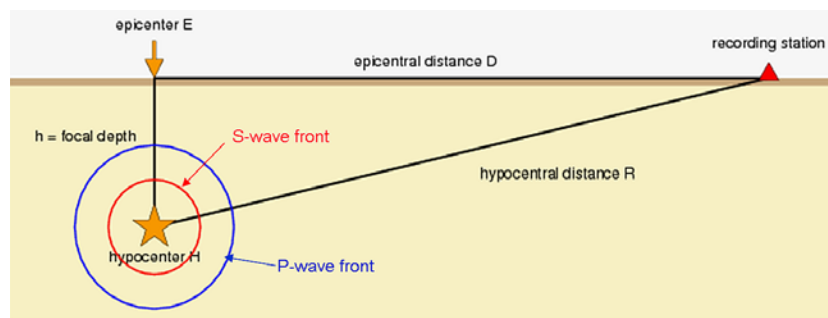
Having understood these basic features of wave propagation, the formation of different seismic phases due to reflection and conversion, and their appearance in seismic records you may now enjoy the first animation by left-mouse click on the bold-blue file name [wendt.nz.qt](#) either here or by downloading it from the Directory [Download Programs and Files](#) (see link on the NMSOP-2 front page). Note that the file is very large and the download may take quite some time. Moreover, if you feel that for your purpose or comprehension the movie speed is either too fast or too slow try to catch the cursor running below the movie with a left-button mouse



click. Hold this button pressed and control the movie speed according to the time you need to read the texts and to comprehend the demonstration, or to speed up too slow sections. More explanations for all 10 movies are given in IS 11.3.

## 2.2 Wave front propagation, travel-time curves and location of near seismic events

Figure 6 is a simplified plot of the propagation of seismic wave fronts (blue circles for P waves and red circles for S waves) in a homogeneous Earth crust, i.e., a crust with constant velocity of wave propagation in all directions away from the seismic source (star). The source position at focal depth  $h$  below the Earth surface is termed the hypocenter  $H$  and its projection on the surface the epicenter  $E$ . Accordingly, the distance on the Earth surface between the epicenter  $E$  and the recording station is termed epicenter distance  $D$  and the straight line distance between the hypocenter and the station the hypocenter distance  $R$ .



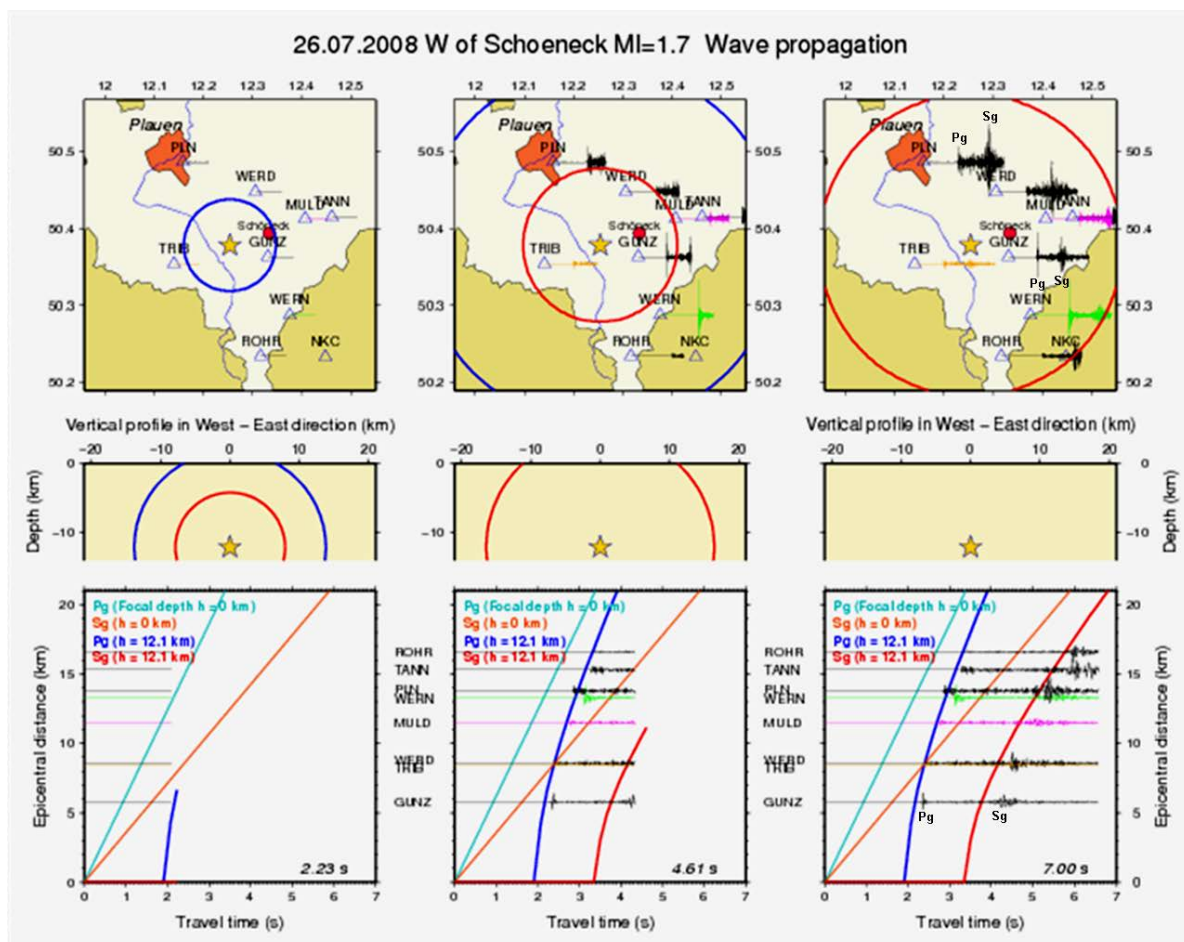
**Figure 6** Simplified plot of wave propagation from a seismic source in a homogeneous Earth crust and explanation of terms for characterizing source position and station distance.

The animation shows:

- that in the case of such simplified assumptions about the crustal velocity model the wavefronts of P and S propagate as expanding circles, both in the vertical plane through the source (star) and in the horizontal plane around the epicenter;
- how the seismic records are written on several seismic station sites as soon as the wave fronts arrive,
- how different the amplitudes and amplitude ratios between the first arriving P waves and the later following S waves may look like in the seismic records, and
- how the travel-time curves of P and S as a function of epicenter distance  $D$  are formed according to the arrival times of the P and S waves.

This is illustrated in Figure 7 by three screen shots of the movie, taken at different times. The rather smoothly bended travel-time curves shown in the lower row of Figure 7 for P and S waves are average curves for the area assuming a homogeneous crustal velocity model and a source depth of 12.1 km. In contrast, the light colored straight-line curves relate to a source depth of zero kilometers. The onset times for P and S in real records at the different stations may differ up to several tenths of seconds from the theoretical ones due to velocity inhomogeneities in the different directions of wave propagation. When drawing the travel-time curves through the really observed onset times of P and S, these curves would look rather “bumpy”, and accordingly the circular wave fronts in the middle and upper row of the screen shots as well. Event location algorithms are based on the least-square minimization of

the difference between observed arrival times of seismic phases and calculated arrival times based on average velocity models. However, without prior detailed investigations we usually do not know much better the real velocity distribution. Then one has to work with the simplified model of circular wavefront propagation. But we should be aware of the possible range of related uncertainties of the location results. They also depend on the number and distance of available recording stations with respect to the seismic source. If, however, we have already a rather detailed knowledge about the 3-D distribution of velocity inhomogeneities by tomographic or other investigations then the number of data to be taken into account in calculating the source location would increase with the third power. Handling such data requires huge data storage and super high-speed data processing facilities. But our tutorial and related animation demonstrates that even with simplified assumption reasonably good seismic source locations can already be made by applying a **circle and chord method**, which can be performed even by hand (for details see EX 11.1).



**Figure 7** Screen shots of the movie about P- and S-wave propagation from a small earthquake in Germany with magnitude  $M_l = 1.7$ , recorded in the local distance range up to 20 km. **Upper row:** Horizontal circular expansion of the of P (blue) and S (red) wavefronts and the related short-period seismic records at stations of a local seismic network. Note the strong variation of the amplitude ratio between P and S waves at different stations. **Middle row:** Vertical circular expansion of the P and S wavefronts. **Lower row.** Light blue and red linear curves: Theoretical travel-time curves for P and S waves from a source at depth  $h = 0$  km; dark blue and red: theoretically expected travel-time curves for P and S waves radiated by a source at 12.1 km depth, superimposed by short-period records at seismic stations as a function of epicenter distance. Download and activate the movie by left-mouse click on the file name [3D-wave-prop travel-time qt.](#)



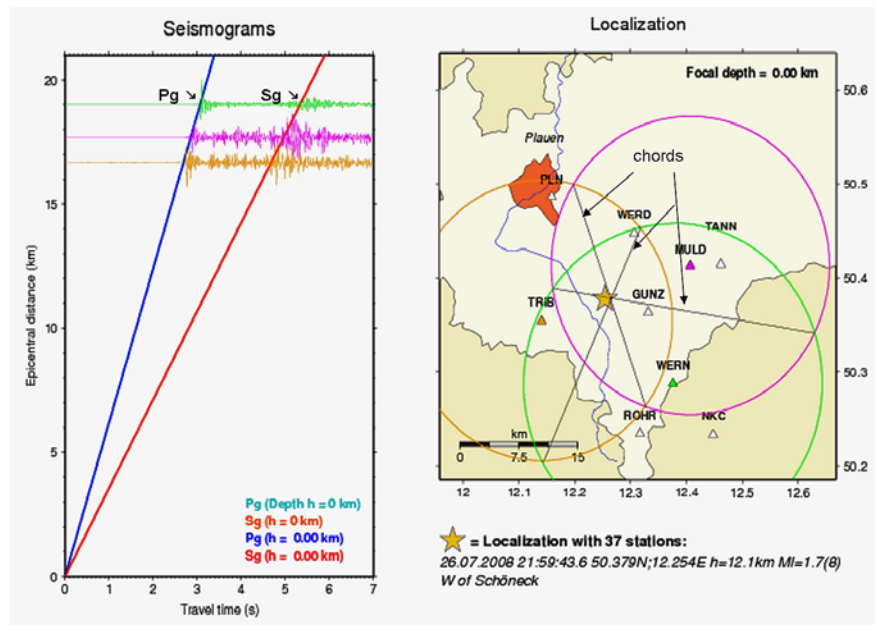
For carrying out the circle and chord exercise one has to know that longitudinal P waves travel roughly  $\sqrt{3}$  times faster than transverse S waves. Further that in continental areas the crustal thickness is on average about 30-35 km thick. For shallow crustal earthquakes Pg is then the first arriving longitudinal wave and Sg the first arriving transverse wave up to distances of about 5 times the crustal thickness (see Figure 1). The difference of arrival times,  $t(Sg - Pg)$  measured in seconds, increases more or less linear with hypocenter distance R in kilometers according to the relationship

$$R = t(Sg - Pg)[(v_p \cdot v_s) / (v_p - v_s)].$$

R can thus roughly be estimated when  $v_p$  and  $v_s$  are known. Assuming for Pg in a homogeneous crust a constant average velocity  $v_p = 5.9$  km/s and thus for Sg of  $v_s \approx v_p / \sqrt{3} \approx 3.4$  km/s then  **$R \approx 8 \times t(Sg - Pg)$** . This is the commonly assumed “rule-of-thumb”. However, for events in the region considered in Figure 7  $v_p = 6.2$  km/s yields better locations. Accordingly,  $R = 8.3 \times t(Sg - Pg)$  would then be more appropriate.

For  $h > 0$  R is generally larger than the epicenter distance.  $R = D$  holds only for either  $h = 0$  or  $D \gg h$  ( $\rightarrow$  asymptotic approximation).

Measuring  $t(Sg - Pg)$  on records of at least 3 stations allows a rough source location by drawing circles with radii  $R_i$  around the stations. However, since for  $h > 0$  km  $D < R$  the circles do not intersect at the epicenter but overshoot. Only their chords, i.e., the straight lines connecting the two overcrossings between different pairs of circles (Figure 8 right) intersect close to the epicenter.



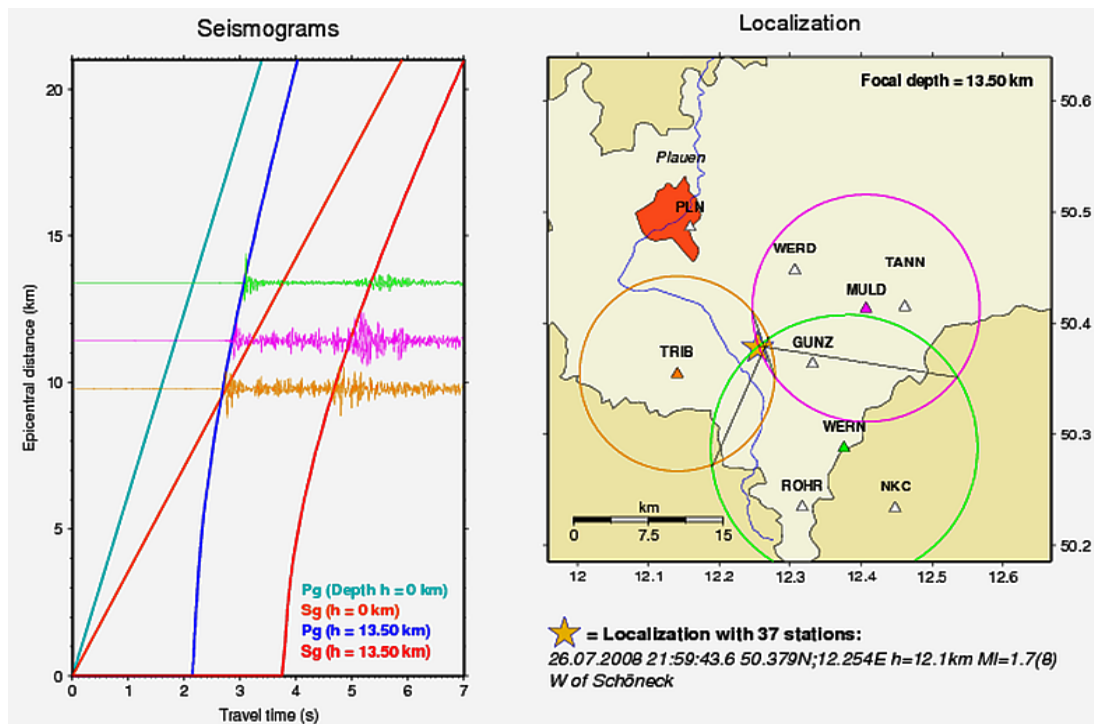
**Figure 8 Left:** Average linear travel-time curves for P and S from a surface source ( $h = 0$  km) in the Vogtland swarm earthquake region in the South of Eastern Germany and their best match with 3 seismic records at stations of the local Vogtland network at epicenter distances of 16.7, 17.7 and 19.0 km, respectively. **Right:** Circles drawn around the stations with the radii according to the epicenter distances estimated from the left-hand diagram, together with the three chords drawn between the crossing points of overshooting circle. The chords intersect close to the epicenter.

When matching the onset times of the Pg and Sg wave groups recorded at three stations in the near source area with the average linear travel-time curves for P and S waves from a surface source one would estimate epicenter distances of 16.7, 17.7 and 19.0 km, respectively (Figure 8 left). When drawing circles with these radii around the respective stations they strongly overshoot. The circle overshoot, however, is a strong indicator for a finite source depth, which is the deeper the larger the overshoot. By step-wise increase of  $h$ , the calculated epicenter distance radii  $d_i$  decrease (see Figure 9) until for  $d_{i \min}$  the “true” (better: the most likely) source depth  $h_{\text{true}}$  can approximately be estimated via the relationship

$$d_{i \min} = (R_i^2 - h_{\text{true}}^2)^{-2}.$$

The circles drawn around the stations  $i$  with  $d_{i \min}$  for “ $h_{\text{true}}$ ” are expected to have minimum overshoot (see Figure 9), provided that the wave propagation conditions are sufficiently homogeneous and well represented by the assumed velocity model, and further, that the reading errors of phase onset times on the station records are negligible.

For seeing and downloading the whole movie click with the left mouse button on the file name [wendt waveprop circles](#) either here, or in the list of [Download programs and files](#).

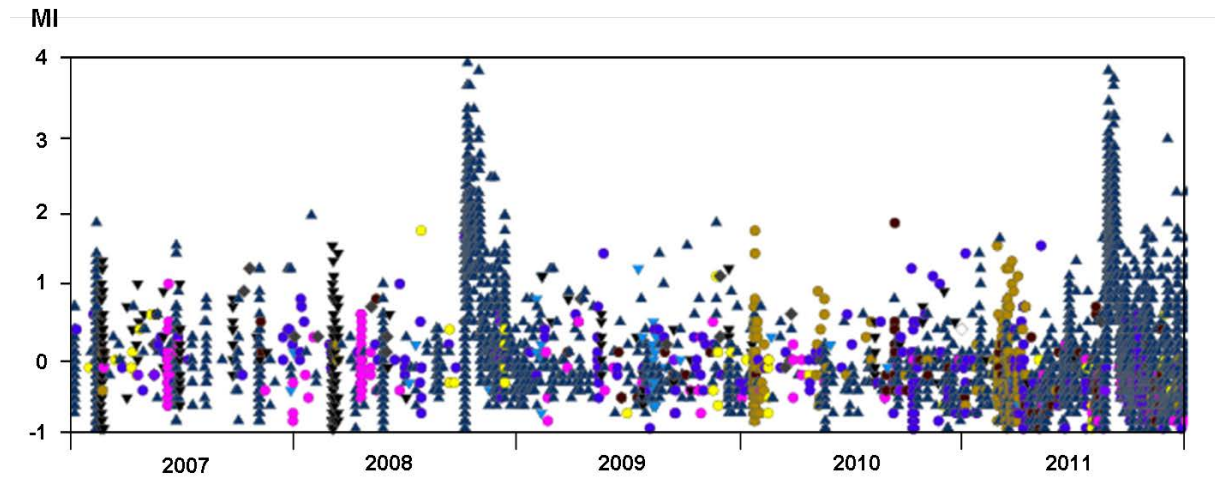


**Figure 9** The minimum circle overshoot is achieved for “ $h_{\text{true}}$ ” = 13.5 km, corresponding to circle radii of epicenter distances of 9.8, 11.4 and 13.4 km, respectively.

### 2.3 Presentation of earthquake sequences in their space-time-magnitude development

Earthquake swarms and aftershock sequences of large earthquakes may consist of hundreds to thousands of events within a limited area, occurring within a relatively short time span with more or less clear spatial relationship to the activated regional seismotectonic features. Moreover, sometimes such sequences are characterized by specific event-number-magnitude-

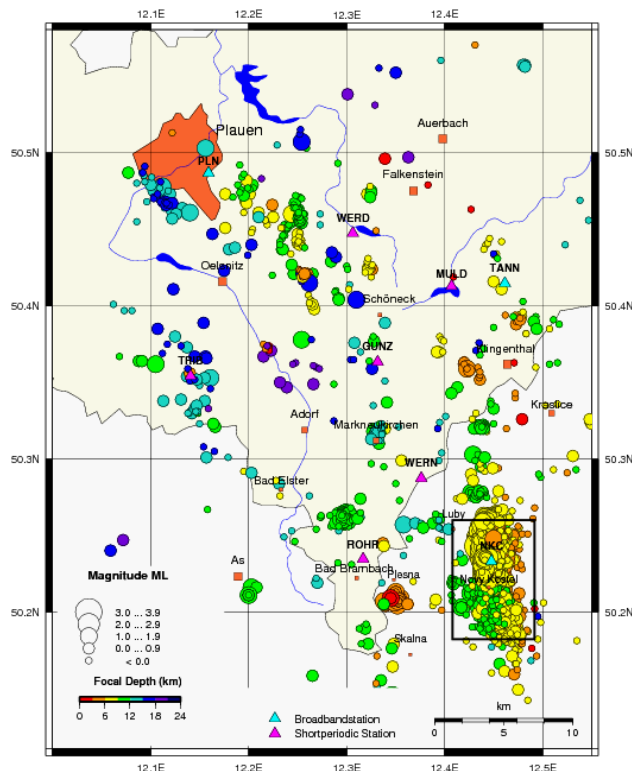
frequency time plots (Figure 10) and may also show pronounced spatial migration. In order to facilitate a well structured comprehensive analysis of such a multitude of features suitable forms of data presentation are highly desirable, both in the form of 2-D diagrams or projections of 3-D hypocenter clouds as well as their viewing from different perspectives.



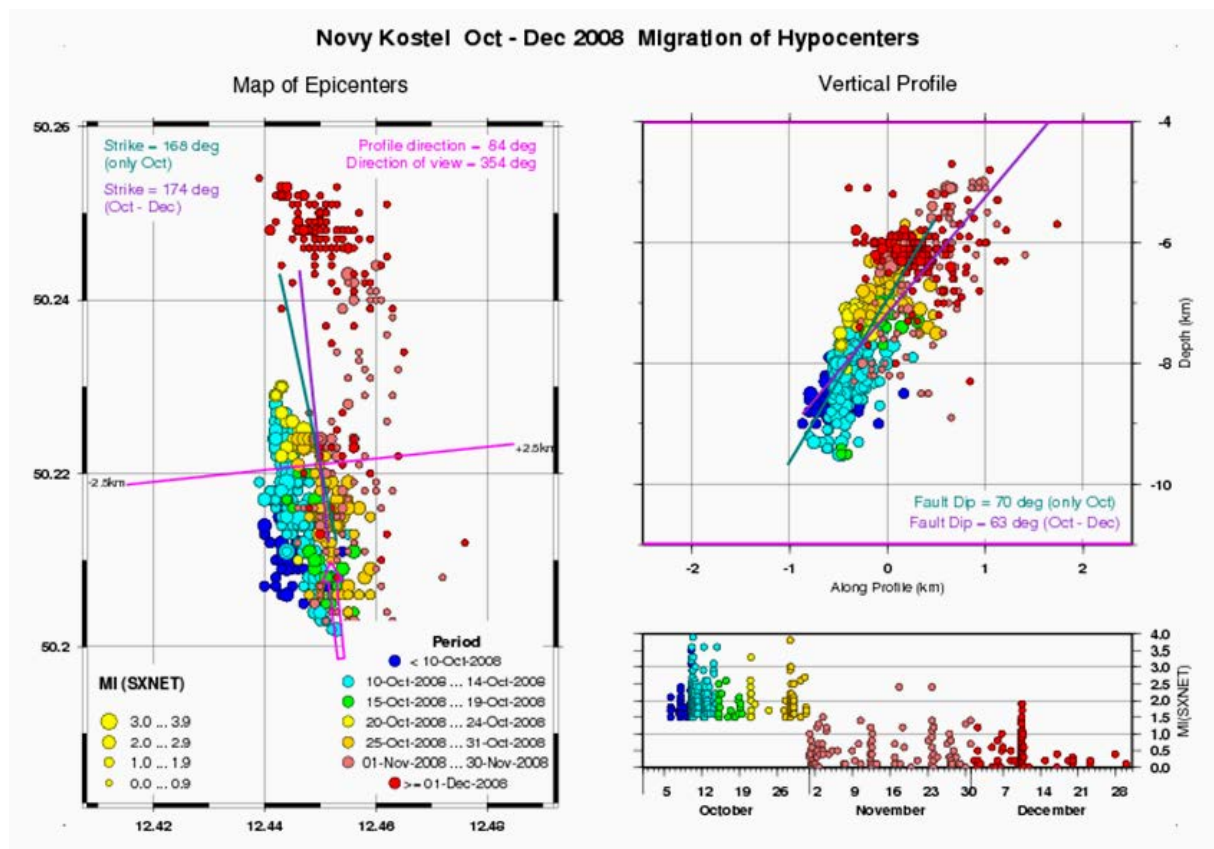
**Figure 10** Local magnitude MI frequency of occurrence of swarm and individual earthquakes in the Vogtland/Germany (dots) and Western Bohemia/Czechia (triangles) region (see Figure 11) between January 1, 2007, and Dezember 31, 2011. Different colors relate to different subregions. Total number of earthquakes 5392.

The animation illustrates this by way of example of an earthquake swarm that occurred between October and December 2008 in Western Bohemia/Czechia. Figure 11 shows the wider area of this source region and the inserted box frame delimitates the area of the specific swarm dealt with in our animation. The total number of detected events in this swarm was 8500 in the magnitude range  $-1 < MI < 4$ . The epicenters fall within an area of only some 6 km in length and 1.5-2.5 km in width. During this swarm a south-to-northward migration of the foci has been observed with gradually decreasing focal depth from initially about 9.5 km to 4.5 km and slight changes in the dominating strike and dip direction of the activated local fault system as inferred from the cumulative clustering of hypocenters (see Figure 12).

To ease the comprehension of these developments in time, the epi- and hypocenters have been assigned different color, ranging from dark blue in early October to bright red in late December 2008 and then been projected to vertical cut planes through the 3-D event cloud along different profiles. The size of the event circles differs with magnitude. The dominating strike direction of the activated fault system can be inferred from the main axis through the epicentre cloud and the related fault dip from the main axis of the best aligned hypocenter projection plot. The latter can best be found by “walking around” the source area and looking at vertical profile projections from different view angles (see open arrow in the animation).. The animation illustrates this. Activate and/or download the movie by left-button-mouse click on the bold-blue file name [\*\*wendt source migra around\*\*](#) either here or in the listing of [Download programs and files](#), which can be opened via *overview* on the front page of NMSOP-2.



**Figure 11** Epicenter map of the earthquake region Vogtland/Germany and Western Bohemia/Czechia. The inserted box in the lower right part delimitates the area of the earthquake swarm in Czechia between October and December 2008 in Figure 10.

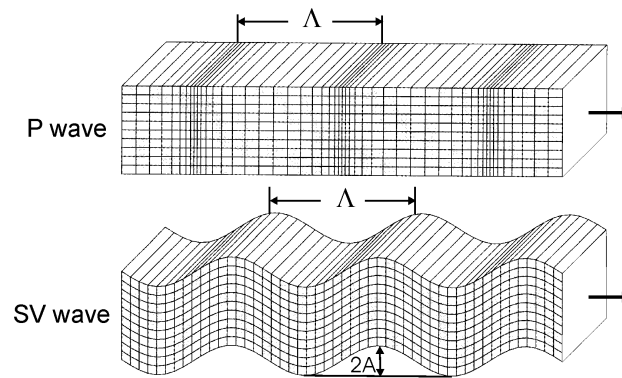


**Figure 12** **Left:** Epicenter plot of the earthquake swarm. **Upper right:** hypocenters projected to the vertical plane of the near E-W profile in the epicentre plot. **Lower right:** Event-number-magnitude development with time for MI > 1.5 (during October), respectively MI > 0 during November and December 2008.

In summary, the respective data plots yield a kind of regional cumulative fault plane solution for the earthquake swarm as a whole or of its sub-phases in time. For the main phase in October 2008 the inferred strike direction of the activated fault system is about  $168^\circ$  from North and the fault dip is about  $70^\circ$  towards west. The average strike direction for the total swarm was about  $174^\circ$  with a dip of  $63^\circ$ .

## 2.4 Radiation patterns of earthquake fault mechanisms

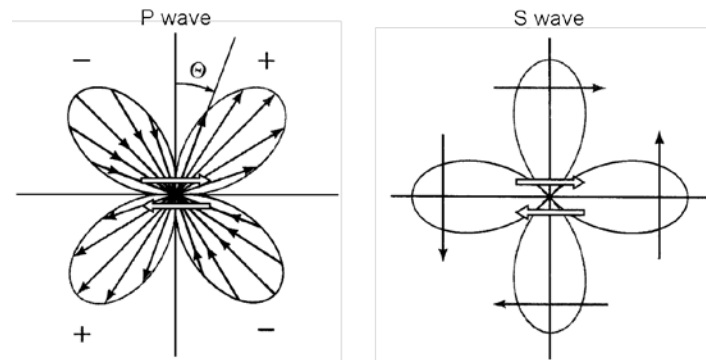
In the event of an earthquake the two adjacent rock formations on both sides of the earthquake fault move relatively to each in opposite direction (see Figure 14, the open arrows). The released elastic strain energy generates both near-source kinematic energy, and the kinematic energy in the source region that propagates as seismic waves to the far field. The seismic waves consist of longitudinal motions in the direction of wave propagation (P waves) and transverse motions perpendicular to the direction of wave propagation (S waves). In the case of S waves the transverse motion may be separated into two components, namely motion in the vertical plane, as in Figure 13, termed SV, and those in the horizontal plane, termed SH.



**Figure 13** Displacement patterns of a harmonic plane P wave (top) and SV wave (bottom) propagating in a homogeneous isotropic medium.  $\Lambda$  is the wavelength and  $2A$  means double amplitude. The white surface on the right is a segment of the propagating plane wavefront where all particles undergo the same motion at a given instant in time, i.e., they oscillate *in phase*. The arrows indicate the seismic rays, defined as the *normal* to the wavefront, which points in the direction of propagation. (Copied from Fig. 2.6 in Chapter 2, modified according to Shearer, Introduction to Seismology, 1999; with permission from Cambridge Univ. Press).

Earthquake shear ruptures show a typical radiation pattern, which is different for P and S waves. These patterns are characterized by so-called amplitude lobes, i.e., by azimuth-dependent radiation of amplitudes, respectively energy. In the case of a pure shear rupture in a homogeneous isotropic medium P-waves have the smallest (idealized zero) amplitudes in the direction of fault rupture and the largest amplitudes (see black arrows in Figure 11, left side) at an angle  $45^\circ$  off the fault plane whereas shear (S) waves have the largest amplitudes in the direction of the shear rupture and perpendicular to it. Joint analysis of P and S wave amplitudes, and especially of the amplitude ratios of P/SV and P/SH, is – although a quite complicated procedure – a powerful method to identify the type of earthquake rupture, to reliably constrain its orientation in space and to quantify its defining parameters with an even

smaller number of observations points and less complete azimuthal coverage than required when only P-wave first motion polarities are used.



**Figure 14** Diagrams of the radiation pattern of the radial displacement component of P-waves (left) and of the transverse displacement component of S waves (right) from a double couple shear rupture (Modified combination of Figures 4.5 and 4.6 of Aki and Richards, 2002; with permission of P. Richards).

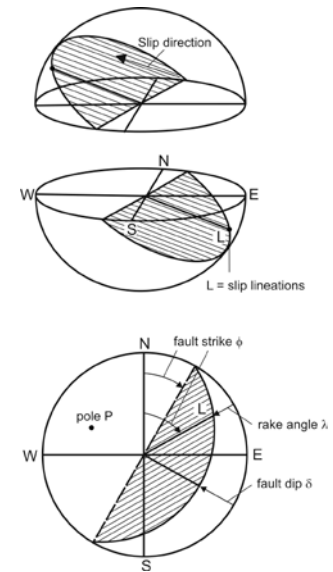
One recognizes from Figure 14 (left-hand panel) that the longitudinal source displacement is outward directed in the upper right and lower left quadrant. Accordingly, the related P waves arrive at seismic stations as a compression signal which produces in vertical component records an upward first motion (+). In contrast, in the two other quadrants, 90° away from the + quadrants, the longitudinal displacement is directed towards the source (dilatation). Accordingly, the P-wave first motion in vertical component records is downward (-). Therefore, the most simple and most frequently used method of determining the seismic source mechanism is to investigate the azimuthal distribution of first motion polarities observed at seismic stations surrounding the seismic source in all directions (see EX 3.2). However, small earthquakes are commonly recorded in narrowband filtered seismic records of high magnification. Due to the rather pronounced transient response of such records the first motion amplitudes may be significantly smaller than the largest amplitudes in the P waveforms and be distorted or even buried in the seismic noise (see Chapter 4, Figs. 4.9 and 4.10 and Chapter 16, Fig. 16.25). Therefore, one requires many more and azimuthally well distributed P-wave polarity readings than P/S wave amplitude ratios in order to produce a well constrained fault plane solution. Also, maximum amplitudes in the P- and S-waveforms can still reliably be read if the P-wave first motion is already buried in the noise.

According to Figure 14 a fault which ruptures perpendicular to the horizontally plotted fault rupture would produce the same radiation pattern. Therefore, analyzing only P-wave polarity distributions will not allow to say, which of the two possible faults has really ruptured. Therefore, in this kind of an idealized point source rupture model according to a double-couple shear force system (see black arrows in the right S-wave panel of Figure 14) the amplitude radiation lobes have been symmetrically mirrored. Yet, this ambiguity which of the two potential fault planes is really the acting one is a principle one, independent on whether one analyzes the P- or S-wave radiation pattern or that of the P/S-wave amplitude ratios. In practice it can only be resolved by assessing either the directivity effects (see Figure 19 and related comments), or the distribution of aftershocks, or by invoking geological information (tectonic setting, mapped faults; see solutions of EX 3.2).



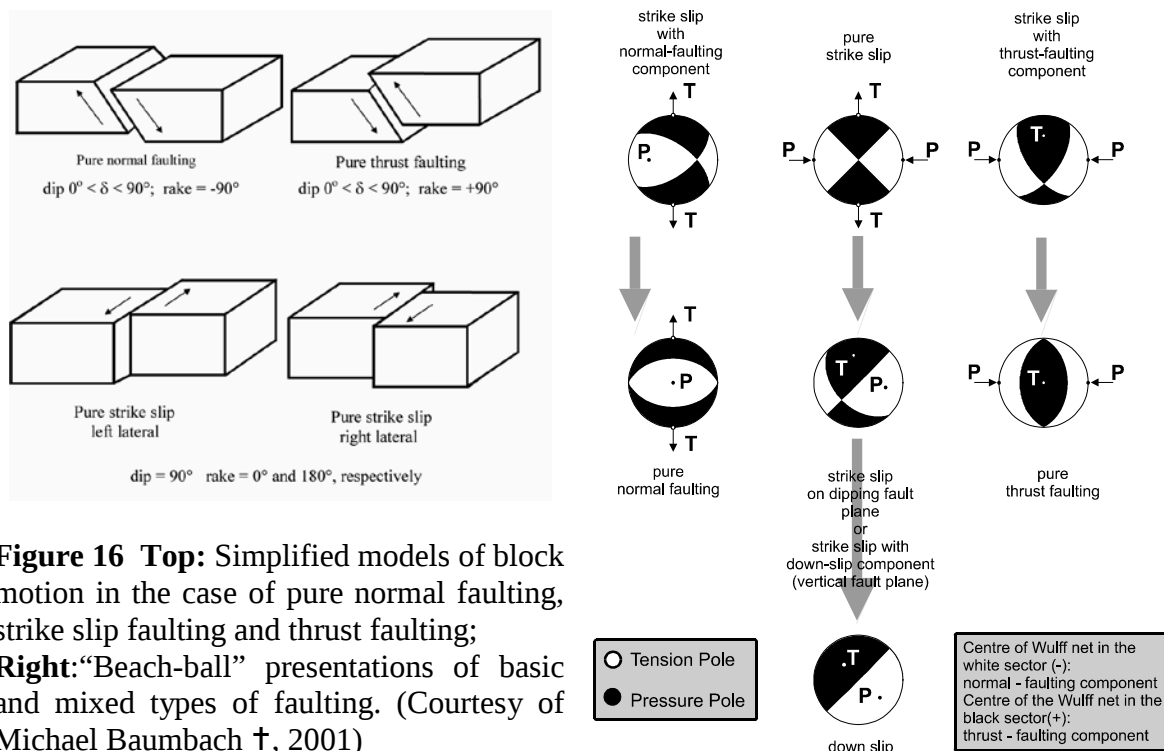
The key parameters which control the orientation and slip of a fault in space and thus the observed radiation pattern of P- and S-wave amplitudes are (see Figure 15):

- strike angle  $\phi$  (measured clockwise from north);
- dip angle  $\delta$  (measured downwards from the horizontal);
- rake angle  $\lambda$ , which describes the slip direction (values between  $0^\circ$  and  $180^\circ$  for upward motion or between  $0^\circ$  and  $-180^\circ$  for downward motion of the hanging/overlying wall).



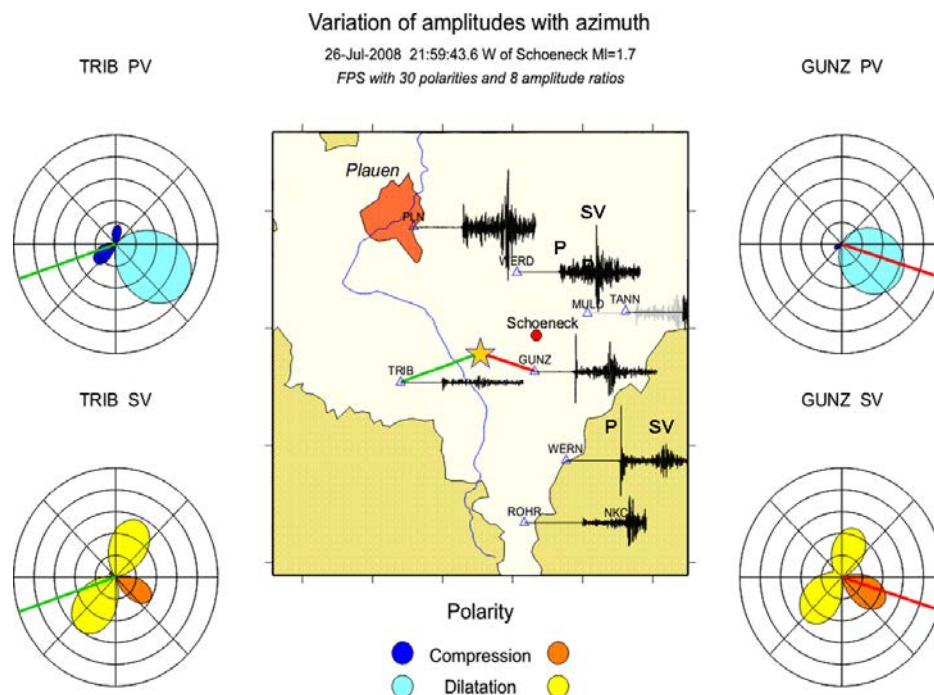
**Figure 15** Idealized rupture plane with slip lineations surrounded by an imagined so-called “focal sphere” centered in the origin. **Top:** upper hemisphere, **middle:** lower hemisphere; **bottom:** projection of the lower focal hemisphere fault plane onto the horizontal plane between the two hemispheres. The block of the Earth medium above the fault plane is termed the “hanging wall”, that below the “foot wall”.

According to the relationship between these three fault-plane parameters one classifies earthquake ruptures into 3 basic types and related mixed types. In seismological publications they are usually presented as the projections of the fault plane cut traces through the focal sphere onto the horizontal plane between the upper and lower focal hemisphere and by coloring or shading differently the quadrants with compressional and dilatational first P-wave motions. Such presentations of fault plane solutions are also nick-named as “beach-ball solutions” (see Figure 16).



**Figure 16** **Top:** Simplified models of block motion in the case of pure normal faulting, strike slip faulting and thrust faulting; **Right:** “Beach-ball” presentations of basic and mixed types of faulting. (Courtesy of Michael Baumbach †, 2001)

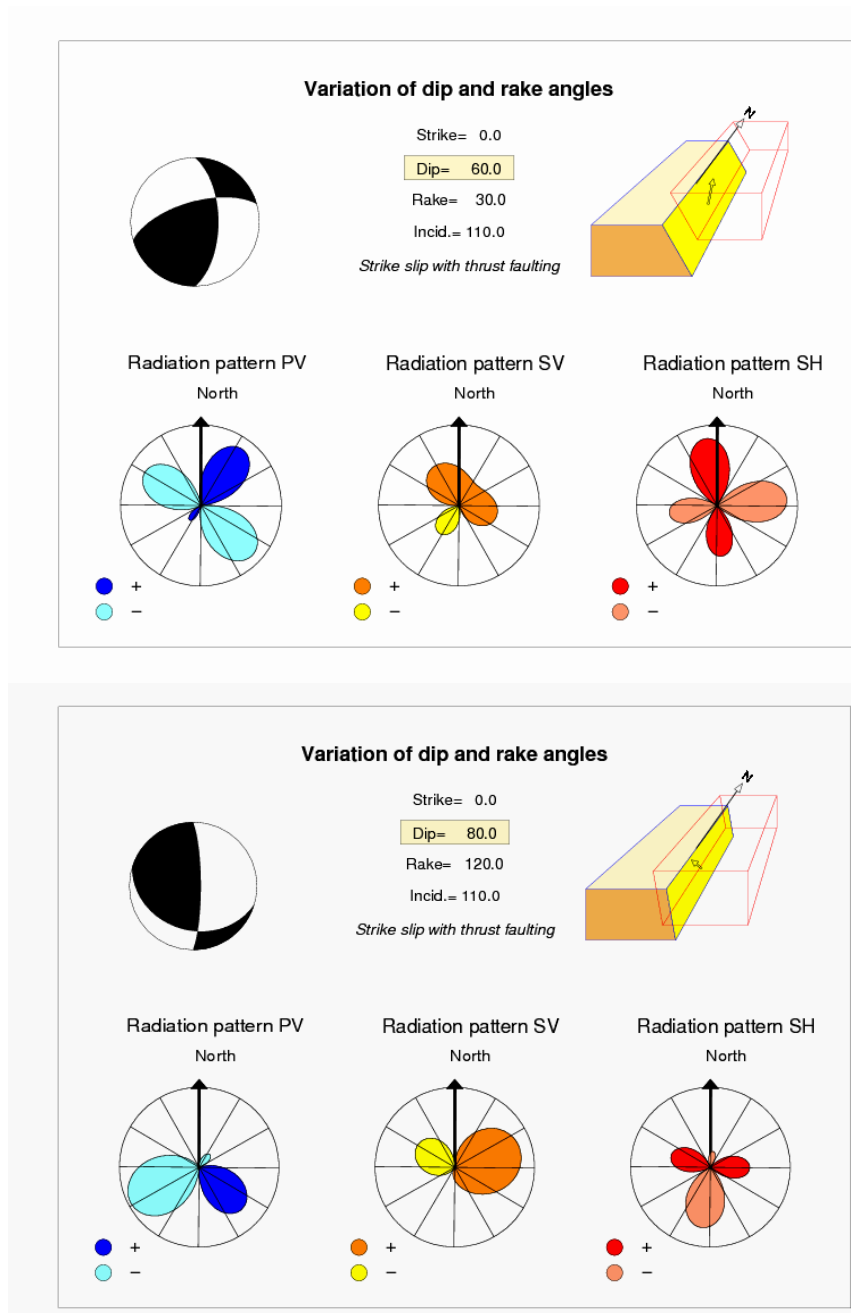
As compared to Figure 14, the amplitude radiation patterns for P and S waves observed on the Earth surface from earthquake ruptures with different fault-plane and slip orientation in space are rather asymmetric. This could be shown for a well constrained fault plane solution of a Vogtland earthquake in Germany with known strike, dip and rake. The radiation lobes (also for the animation examples) have been calculated by using the formulas (4.89) to (4.91) published by Aki and Richards (2002). No assumptions with respect to *source directivity*, rupture velocity or type of slip (e.g., unilateral, bi-lateral, circular) have been made (see final discussion below). The agreement with the actually recorded amplitude ratios measured at stations of a local network is remarkably good. E.g., for the station TRIB the calculations predict for PV a very small amplitude but a slightly larger SV amplitude. In contrast, as observed, PV should be larger than SV for station GUNZ (Figure 17). This concerns the expected amplitude ratios, not the absolute amplitudes. The latter additionally depend on epicenter distance, local site amplification effects and other factors.



**Figure 17** Comparison of theoretically calculated radiation patterns for an earthquake with well constrained fault plane solution and slip direction with the really observed amplitude ratios between vertical component P and S amplitudes at stations of a local seismic network in Germany. The source depth of the earthquake has been 12.1 km and the epicenter (respectively hypocenter) distances of stations GUNZ and TRIB were 5.8 km (13.4 km) and 8.5 km (14.7 km).

The azimuth-dependent PV, SV, SH amplitude lobe patterns depend on the **strike, dip and rake** and thus on the type of source mechanism of the acting fault plane. But with respect to the seismic stations which record these amplitudes the lobe size and patterns also depend on the incidence angle of the seismic rays. This is shown with the first animation, while in the following animations, which demonstrate the effect of variations in dip and rake, the incidence angle at the recording stations has been kept fixed. The movie with several of such

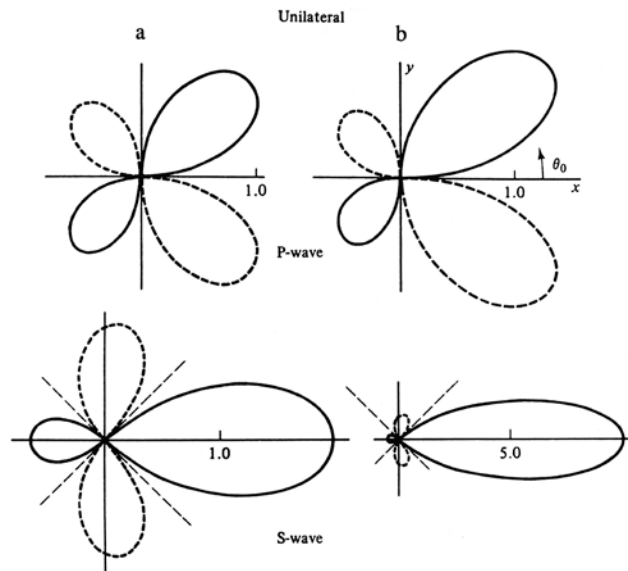
animations can be viewed and/or downloaded by left mouse-button click on the file name [wendt\\_source\\_rad\\_pat](#) either here or in the listing [Download programs and files](#). After downloading Linux users should activate the movie by typing after the command **xanim** the file name **wendt\_source\_rad\_pat**. Figure 18 presents two snap shots from this movie.



**Figure 18** Cut-out scenes of the movie [wendt\\_source\\_rad\\_pat](#) for dip and rake variations.

In all these calculations, source directivity has been neglected. According to the NMSOP-2 Glossary, the term **directivity** is defined as: “An effect of a propagating fault rupture whereby the amplitudes and frequency of the generated ground motions depend on the direction of wave propagation with respect to fault orientation, slip direction(s) and rupture velocity  $v_r$ . The directivity and thus the radiation pattern is different for P and S waves.” It becomes more pronounced when  $v_r$  approaches the respective wave velocity. Accordingly, it is larger

for S than for P waves. This is illustrated by Figure 19.  $v_R$  commonly ranges between about 0.2 and 0.9 times of the shear-wave velocity  $v_s$  (Venkataraman and Kanamori, 2004b). Yet, in rare cases, it may even exceed  $v_s$  if the hypothesis and still debated occasional observational evidence of supershear slip pulses with rupture velocities up to 5 km/s really holds (e.g., Bouchon and Vallée, 2003).



**Figure 19** Variability of P- and SH-wave amplitude for a fault rupture propagating unilaterally from left to right with  $v_R/v_s = 0.5$  (left column) and  $v_R/v_s = 0.9$  (right column), respectively.

(From Kasahara, 1981; © Cambridge University Press).

In the directions of enlarged amplitudes the recorded broadband pulse width of P and S (or source time function) is compressed, in the directions of reduced amplitudes, however, stretched, i.e., the apparent rupture duration is azimuth dependent. Yet the area underneath the pulses, which is proportional to the seismic moment of the fault rupture, remains constant, i.e., it is independent on the azimuth of observation.

Pronounced directivity effects are observed for extended (finite linear in contrast to point source) ruptures of strong earthquakes, primarily in the case of unilateral rupture propagation. The azimuthal effects, however, are much less in the case of bilateral ruptures and disappear in the case of circular ruptures. Only in rare cases the effects of vertical directivity have been observed. (See related discussions and Figures in Lay and Wallace, 1995). The frequency dependence is due to the Doppler-effect. It can be quantified by investigating the azimuth-dependence of the frequency spectra of the recorded P and S waves. Thus, in the case of distinct directivity effects one can unambiguously identify, which of the two potential rupture planes has really moved.

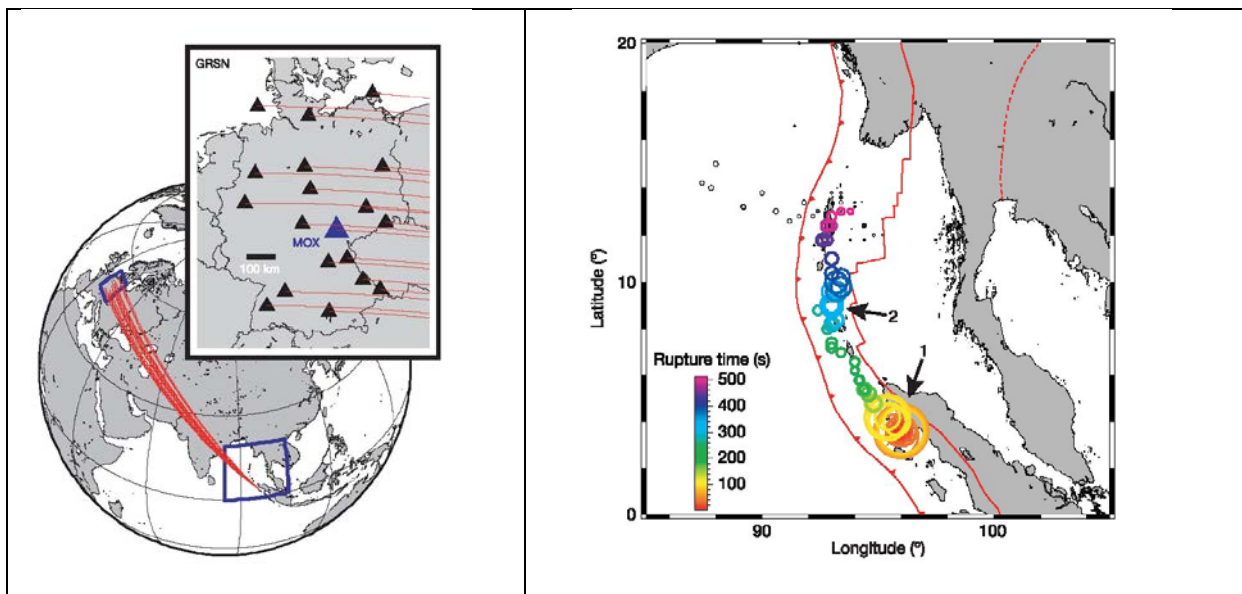
For small local events, however, recorded at rather high frequencies, directivity effects can usually be neglected. E.g., in the case of the earthquake in Figure 17 with  $M_l = 1.7$  the source radius has been estimated to be about 0.1 km only, when assuming a circular rupture according to the Brune (1970) or Madariaga (1976) models (see EX 3.4). Moreover, according to Cho et al. (2010), observations suggest that directivity effects are generally present only at frequencies less than about 1 Hz, whereas the records presented in Figure 17 consist only of frequencies above 6 Hz. At high frequencies, the directivity effect is generally reduced and the radiation pattern observed in the far-field (i.e., at distances of several wavelengths, as in our case) changes from the usual double-couple pattern as in Figure 14 to a more isotropic one. Such differences in the radiation patterns between high and low frequency

ground motion have already earlier been explained as either a source effect (with real sources losing coherence at short wavelengths associated with source roughness/irregularities that are not considered in planar fault models) or as a multi-pathing effect (with high frequency waves being preferentially scattered by crustal heterogeneities (e.g., Newmann and Okal, 1998) or a combination of both (Bormann and Di Giacomo, 2011)).

In summary, if spatially extended earthquakes do not, as in Figure 19, propagate unilaterally but bilateral or in an even more complex manner (as, e.g., the Chile earthquake 2010 in section 2.6), radiation patterns may look different from those presented here. They may even vary with time during great ruptures that last for several minutes.

## 2.5 Rupture tracking of the great 2004 Mw9.3 Sumatra-Andaman earthquake

This is a pure movie without accompanying text. It illustrates the variability of energy release in space and time by great earthquake ruptures which can not be described by a point source or otherwise simplified rupture models. The movie has kindly been made available by Matthias Ohrnberger. It relates to the publication in *Nature* by Krüger and Ohrnberger (2005) about rupture tracking of the great 26 December 2004 Sumatra-Andaman tsunamigenic earthquake by using the German Regional Seismic Network as a large-scale teleseismic array (see Figure 20). The movie illustrates that such a great earthquake with variable rupture trend, rupture velocity and energy release in different rupture segments can really not be described reasonably well by a common single source fault-plane solution. Tsai et al. (2005) have later approximated this earthquake by a more realistic multi source rupture with 5 major sub-sources.



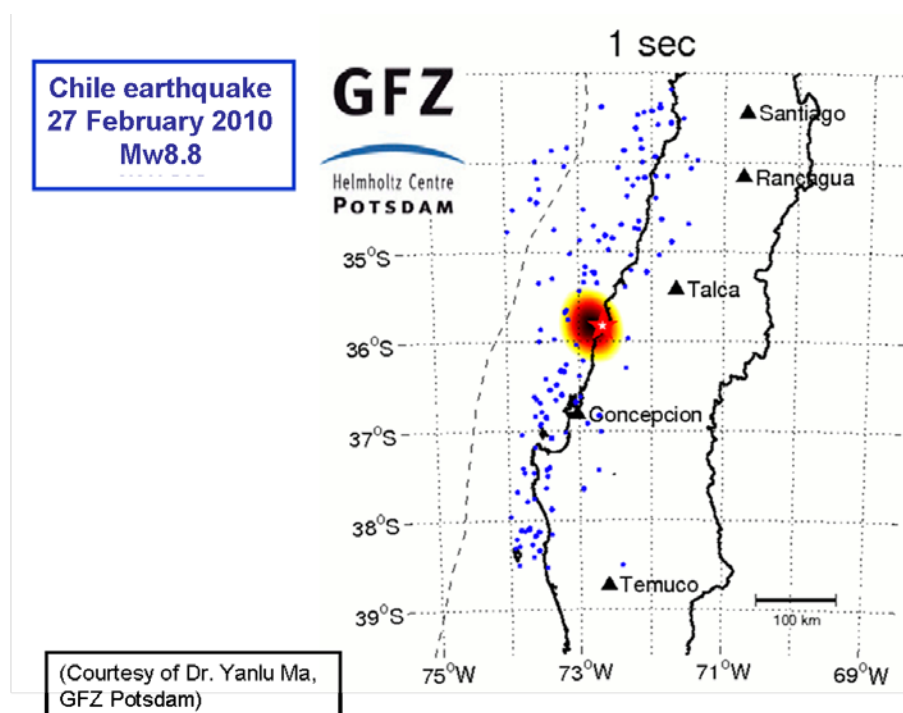
**Figure 20** **Left:** Ray paths from the source area of the Sumatra-Andaman earthquake to Germany and, in the insert, the outlay of stations of the German Regional Seismic Network (GRSN) that has been used as a large-scale teleseismic array for tracking the rupture. **Right:** Rupture propagation as a function of time with the radius of the circles being proportional to the seismic energy released. Combined Figures 1 and 3 of Krüger and Ohrnberger (2005), *Nature*, 435, p.937 and 938; doi: 10.1038/nature0396 (© Copyright Clearing Centre of Nature).

For activating and/or downloading of the movie click with the left mouse button on the file name [ohrnberger sumatra2004 rupture](#) proceed as described above.

## 2.6 Rupture tracking of the great 2010 Mw8.8 Chile earthquake

A similarly strong damaging earthquake occurred in Chile on 27 February 2010. Its rupture process has been tracked by Yanlu Ma of the GFZ German Research Centre for Geosciences by applying a similar procedure described by Krüger and Ohrnberger (2005). He used travel-time delays observed at North American seismic stations, mainly of the USArray, yet including some other permanent stations, e.g., of the California network, as well.

In contrast to the Sumatra earthquake, which can be described as a unilateral rupture, the centers of large seismic energy release during this Chile earthquake jump forth and back, making it to a multilateral rupture. This animation illustrates well how irrelevant in such a case parameters such as epicenter location, average displacement, average stress drop, average rupture velocity or similar are with respect to assessing the hazard posed by such an extended earthquake rupture.



**Figure 20** Map of area of the great Chile earthquake of 2010. Blue dots: epicentres of aftershock within the first few days. Red star: epicentre of the main shock. The colored areas cover the 95% confidence region of seismic energy release of major sub-events of this complex rupture.

Activate/download this movie with the left-button mouse click on the file name [gfz ma chile2010 rupture](#) here or in the directory *Download programs and files* (see overview on the NMSOP-2 front page).



## Acknowledgments

We acknowledge with gratitude the animations and figures made available by Matthias Ohrnberger, University of Potsdam, and of Yanlu Ma, GFZ Potsdam, Germany, for sections 2.5 and 2.6. as well as the use of Figures produced by late Michael Baumbach †, GFZ Potsdam, for Chapter 3 of NMSOP-1 (2002) which enabled us to compile Figure 16. We also thank Paul Richards for extensive clarifying discussions and helpful comments on matters discussed in section 2.4. and for permitting the compilation of Figure 14 based on Figures 4.5 and 4.6 in Aki and Richards (2002).

## References

- Aki, K., and Richards, P. G. (2002). *Quantitative seismology*. 2nd edition, University Science Books, Sausalito, CA., xvii + 700 pp.
- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bouchon, M., and Vallée, M. (2003). Observation of long supershear rupture during the magnitude 8.1 Kunlunshan earthquake. *Science*, **301**, 824-826.
- Brune, J. N. (1970). Tectonic stress and the spectra of shear waves from earthquakes. *J. Geophys. Res.*, **75**, 4997-5009.
- Cho, H., Hu, J., Klinger, Y., and Dunham, E. M. (2010). Frequency dependence of radiation patterns and directivity effects in ground motion from earthquakes on rough faults. Am. Geophys. Union, Fall Meeting 2010, abstract #S51A-1914.
- Kasahara, K. (1981). *Earthquake mechanics*. Cambridge University Press, Cambridge, UK.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, **122**, 108-124.
- Lay, T., and Wallace, T. C. (1995). *Modern global seismology*. ISBN 0-12-732870-X, Academic Press, 521 pp.
- Madariaga, R. (1976). Dynamics of an expanding circular fault. *Bull. Seism. Soc. Am.*, **66**, 639-666.
- Montagner, J.-P., and Kennett, B. L. N. (1996). How to reconcile body-wave and normal-mode reference Earth models? *Geophys. J. Int.*, **125**, 229-248.
- Newman, A. V., and Okal, E. A. (1998). Teleseismic estimates of radiated seismic energy: The  $E/M_0$  discriminant for tsunami earthquakes. *J. Geophys. Res.*, **103**, 26,885-26,897.
- Krüger F., and Ohrnberger, M. (2005). Tracking the rupture of the  $M_w = 9.3$  Sumatra earthquake over 1,150 km at teleseismic distances. *Nature*, **435**, 937-939; doi: 10.1038/nature03696.
- Tsai, V. C., Nettles, M., Ekström, G., and Dziewonski, A. (2005). Multiple CMT source analysis of the 2004 Sumatra earthquake. *Geophysical Research Letters*, **32**; L17304, doi: 10.1029/2005GL023813.
- Venkataraman, A., and H. Kanamori (2004a). Observational constraints on the fracture energy of subduction zone earthquakes, *J. Geophys. Res.* **109**, B04301, doi: 10.1029/2004JB002549.

# Chapter 2

## Seismic Wave Propagation and Earth models

(Version December 2012; DOI: 10.2312/GFZ.NMSOP-2\_ch2)

Peter Bormann<sup>1)</sup>, E. Robert Engdahl<sup>2)</sup>, and Rainer Kind<sup>1)</sup>

<sup>1)</sup> Formerly GFZ German Research Center for Geosciences, Department 2: Physics of the Earth, Telegrafenberg, 14473 Potsdam, Germany; E-mail: [pb65@gmx.net](mailto:pb65@gmx.net) and [kind@gfz-potsdam.de](mailto:kind@gfz-potsdam.de);

<sup>2)</sup> University of Colorado, Department of Physics, Boulder, CO 80309, USA; E-mail: [Bob.Engdahl@Colorado.EDU](mailto:Bob.Engdahl@Colorado.EDU)

|                                                                                                            | <b>Page</b> |
|------------------------------------------------------------------------------------------------------------|-------------|
| <b>2.1 Introduction (P. Bormann)</b>                                                                       | 2           |
| <b>2.2 Elasticity modules and body waves (P. Bormann)</b>                                                  | 3           |
| 2.2.1 Elastic modules                                                                                      | 3           |
| 2.2.2 Stress-strain relationships and equation of motion                                                   | 5           |
| 2.2.3 P- and S-wave velocities, waveforms and polarization                                                 | 7           |
| <b>2.3 Surface waves (P. Bormann)</b>                                                                      | 15          |
| 2.3.1 Origin                                                                                               | 15          |
| 2.3.2 Dispersion and polarization                                                                          | 17          |
| 2.3.3 Crustal surface waves and guided waves                                                               | 23          |
| 2.3.4 Mantle surface waves                                                                                 | 25          |
| <b>2.4 Normal modes (P. Bormann)</b>                                                                       | 27          |
| <b>2.5 Seismic rays, travel times, amplitudes and phase shifts (P. Bormann)</b>                            | 31          |
| 2.5.1 Introduction                                                                                         | 31          |
| 2.5.2 Huygens's and Fermat's Principle and Snell's Law                                                     | 32          |
| 2.5.2.1 Snell's Law for a flat Earth                                                                       | 33          |
| 2.5.2.2 Snell's Law for the spherical Earth                                                                | 34          |
| 2.5.3 Rays and travel times in laterally homogeneous (1-D) media                                           | 34          |
| 2.5.3.1 Velocity gradient                                                                                  | 34          |
| 2.5.3.2 Effect of a sharp velocity increase                                                                | 35          |
| 2.5.3.3 Effect of a low-velocity zone                                                                      | 38          |
| 2.5.3.4 Refraction, reflection, and conversion of waves at a boundary and formation of inhomogeneous waves | 39          |
| 2.5.3.5 Seismic rays and travel times in homogeneous models with horizontal and tilted layers              | 41          |
| 2.5.3.6 Wiechert-Herglotz inversion                                                                        | 43          |
| 2.5.4 Amplitudes and phase distortions                                                                     | 44          |
| 2.5.4.1 Energy and amplitudes of seismic waves                                                             | 44          |
| 2.5.4.2 Wave attenuation                                                                                   | 46          |
| 2.5.4.3 Phase distortions and Hilbert transform                                                            | 50          |
| 2.5.4.4 Effects not explained by ray theory                                                                | 52          |
| <b>2.6 Seismic phases and travel times in real Earth (P. Bormann)</b>                                      | 54          |
| 2.6.1 Seismic waves and travel times from local and regional events                                        | 55          |

|            |                                                                          |               |
|------------|--------------------------------------------------------------------------|---------------|
| 2.6.2      | Shallow source body waves and travel times at teleseismic distances      | 59            |
| 2.6.3      | Depth phases                                                             | 70            |
| 2.6.4      | Pz±P; Sz±S, etc.                                                         | 73            |
| 2.6.5      | T phases                                                                 | 76            |
| 2.6.6      | Surface-wave fundamental and higher modes and the W phase                | 78            |
| <b>2.7</b> | <b>Global Earth models (E. R. Engdahl)</b>                               | <b>81</b>     |
| 2.7.1      | 1-D models                                                               | 81            |
| 2.7.2      | 3-D models and their use in event location                               | 87            |
| <b>2.8</b> | <b>Synthetic seismograms and waveform modeling (R. Kind, P. Bormann)</b> | <b>91</b>     |
|            | <b>Acknowledgments</b>                                                   | <b>98</b>     |
|            | <b>Recommended overview readings</b>                                     | <b>99</b>     |
|            | <b>References</b>                                                        | <b>98-105</b> |

## 2.1 Introduction (P. Bormann)

The key data to be recorded by means of *seismic sensors* (Chapter 5) and *recorders* (Chapter 6) at seismological observatories (*stations* – Chapter 7, *networks* – Chapter 8, *arrays* – Chapter 9) are *seismic waves*, radiated by *seismic sources* (Chapter 3). Weak signals may be masked or significantly distorted by *seismic noise* (Chapter 4), which is usually considered disturbing and unwanted. Only in some special engineering-seismological applications is seismic noise also appreciated as a useful signal, from which some information on the structure, velocity and fundamental resonance frequency of the uppermost sedimentary layers can be derived (Chapter 14). But most of what we know today of the structure and physical properties of our planet Earth, from its uppermost crust down to its center, results from the analysis of seismic waves generated by more or less localized natural or man-made sources such as earthquakes or explosions (Chapter 3, Figs. 3.1 to 3.4) by either (repeatedly) solving the so-called forward (direct) or the inverse problem of data analysis (Chapter 1, Fig. 1.1).

It is not the task of the New Manual of Seismological Observatory Practice (NMSOP), to provide an in-depth understanding of the theoretical tools for this kind of analysis. There exist quite a number of good introductory (Lillie, 1999; Shearer, 1999) and more advanced textbooks (e.g., Aki and Richards, 1980, 2002 and 2009; Ben-Menahem and Singh, 2000; Bullen and Bolt, 1985; Dahlen and Tromp, 1998; Lay and Wallace, 1995; Kennett, 2001), and a variety of monographs, overview and special papers related to specific methods (e.g. Fuchs and Müller, 1971; Červený et al., 1977; Kennett, 1983; Müller, 1985; Červený, 2001; Chapman, 2002), types of seismic waves (e.g., Malischewsky, 1987; Lapwood and Usami, 1981; Kanamori, 1993; Lognonné and Clévéde, 2002; Romanovicz, 2002; Kennett, 2002) or applications (e.g., Gilbert and Dziewonski, 1975; Sherif and Geldart, 1995; Cara, 2002; Curtis and Snieder, 2002; Minshull, 2002; Mooney et al., 2002; Nolet, 1987 and 2008; Song, 2002; Kanamori and Rivera, 2008; Hayes et al., 2009). Rather, we will take here a more phenomenological approach and refer to related fundamentals in physics and mathematical theory only as far as they are indispensable for understanding the most essential features of seismic waves and their appearance in seismic records and as far as they are required for:

- Identifying and discriminating the various types of seismic waves;
- Understanding how the onset-times of these phases, as observed at different distances from the source, form so-called travel-time curves;

- Understanding how these curves and some of their characteristic features are related to the velocity-structure of the Earth and to the observed (relative) amplitudes of these phases in seismic records;
- Using travel-time and amplitude-distance curves for seismic source location and magnitude estimation;
- Understanding how much these source-parameter estimates depend on the precision and accuracy of the commonly used 1-D Earth models (see IS 11.1);
- Appreciating how these source parameter estimates may be improved by using more realistic (2-D, 3-D) Earth models as well as later (secondary) phase onsets in the processing routines; and
- Being aware of the common assumptions and simplifications made in synthetic seismogram calculations that are increasingly used nowadays in seismological routine practice (see 2.5.4.4; 2.8; 3.5.3).

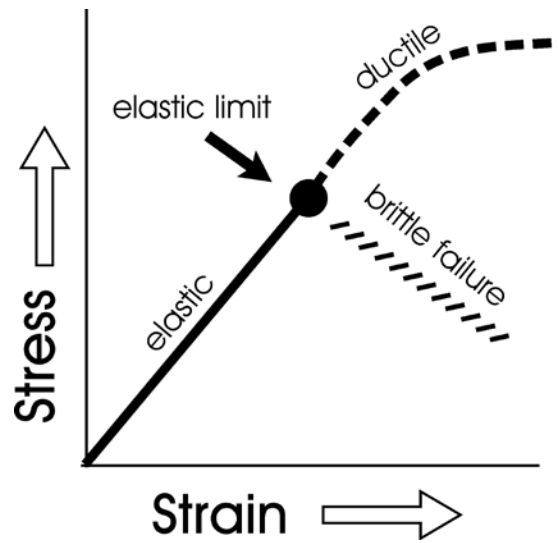
## 2.2 Elastic modules and body waves (P. Bormann)

### 2.2.1 Elastic modules

Seismic waves are elastic waves. Earth material must behave elastically to transmit them. The degree of elasticity determines how well they are transmitted. By the pressure front expanding from an underground explosion, or by an earthquake shear rupture, the surrounding Earth material is subjected to *stress* (compression, tension and/or shearing). As a consequence, it undergoes *strain*, i.e., it changes volume and/or distorts shape. In an inelastic (plastic, ductile) material this deformation remains while elastic behavior means that the material returns to its original volume and shape when the stress load is over. There exist also materials with elastic-plastic behavior.

The degree of elasticity/plasticity of real Earth material depends mainly on the *strain rate*, i.e., on the length of time it takes to achieve a certain amount of distortion. At very low strain rates, such as movements in the order of mm or cm/year, it may behave ductile. Examples are the formation of geologic folds or the slow plastic convective currents of the hot material in the Earth's mantle with velocity on the order of several cm per year. On the other hand, the Earth reacts elastically to the small but rapid deformations caused by a transient seismic source pulse. Only for very large amplitude seismic deformations in soft soil (e.g., from earthquake strong-motions in the order of 40% or more of the gravity acceleration of the Earth) or for extremely long-period free-oscillation modes (see 2.4) one has to take the inelastic non-linear response of the otherwise more or less solid Earth's material into account.

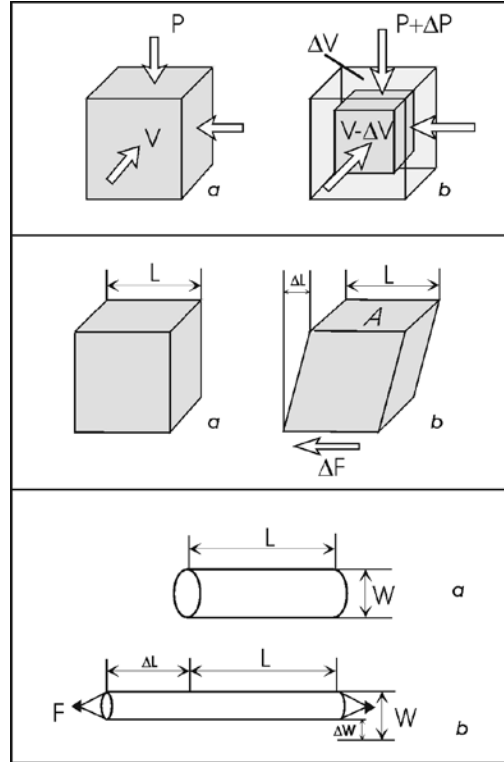
Within its elastic range the behavior of the Earth material can be described by *Hooke's Law* that states that the amount of strain is linearly proportional to the amount of stress. Beyond its elastic limit the material may either respond with brittle fracturing (e.g., earthquake faulting, see Chapter 3) or non-linear ductile behavior/plastic flow (Fig. 2.1).



**Fig. 2.1** Schematic presentation of the relationship between stress and strain.

Elastic material resists differently to stress depending on the type of deformation. It can be quantified by various elastic modules:

- The *bulk module*  $\kappa$  is defined as the ratio of the hydrostatic (homogeneous all-sides) pressure change to the resulting relative volume change, i.e.,  $\kappa = \Delta P / (\Delta V/V)$ , which is a measure of the *incompressibility* of the material (see Fig. 2.2 top);
- The *shear module*  $\mu$  (or “*rigidity*”) is a measure of the resistance of the material to shearing, i.e., to changing the shape and not the volume of the material. Its value is given by half of the ratio between the applied shear stress  $\tau_{xy}$  (or tangential force  $\Delta F$  divided by the area  $A$  over which the force is applied) and the resulting shear strain  $e_{xy}$  (or the shear displacement  $\Delta L$  divided by the length  $L$  of the area acted upon by  $\Delta F$ ), that is  $\mu = \tau_{xy}/2 e_{xy}$  or  $\mu = (\Delta F/A) / (\Delta L/L)$  (Fig. 2.2 middle). For fluids  $\mu = 0$ , and for material of very strong resistance (i.e.  $\Delta L \rightarrow 0$ )  $\mu \rightarrow \infty$ ;
- The *Young’s module*  $E$  (or “*stretch module*”) describes the behavior of a cylinder of length  $L$  that is pulled on both ends. Its value is given by the ratio between the extensional stress to the resulting extensional strain of the cylinder, i.e.,  $E = (F/A) / (\Delta L/L)$  (Fig. 2.2 bottom);
- The *Poisson’s ratio*  $\sigma$  is the ratio between the lateral contraction (relative change of width  $W$ ) of a cylinder being pulled on its ends to its relative longitudinal extension, i.e.,  $\sigma = (\Delta W/W) / (\Delta L/L)$  (Fig. 2.2 bottom).



**Fig. 2.2** Deformation of material samples for determining elastic modules. Top: bulk module  $\kappa$ ; middle: shear module  $\mu$ ; bottom: Young's module  $E$  and Poisson's ratio  $\sigma$ . a – original shape of the volume to be deformed; b – volume and/or shape after adding pressure  $\Delta P$  to the volume  $V$  (top), shear force  $\Delta F$  over the area  $A$  (middle) or stretching force  $F$  in the direction of the long axis of the bar (bottom).

Young's module, the bulk module and the shear module all have the same physical units as pressure and stress, namely (in international standard (SI) units):

$$1 \text{ Pa} = 1 \text{ N m}^{-2} = 1 \text{ kg m}^{-1} \text{ s}^{-2} \quad (\text{with } 1 \text{ N} = 1 \text{ Newton} = 1 \text{ kg m s}^{-2}). \quad (2.1)$$

## 2.2.2 Stress-strain relationships and equation of motion

The most general linear relationship between stress and strain of an elastic medium is governed in the generalized *Hook's law* (see Equation (10) in the IS 3.1) by a fourth order parameter tensor. It contains 21 independent modules. The properties of such a solid may vary with direction. Then the medium is called *anisotropic*. Otherwise, if the properties are the same in all directions, a medium is termed *isotropic*. Although in some parts of the Earth's interior anisotropy on the order of a few percent exists, isotropy has proven to be a reasonable first-order approximation for the Earth material as a whole. The most common models, on which data processing in routine observatory practice is based, assume isotropy and changes of properties only with depth (1-D models; see 2.7.1). Yet, advanced tomographic procedures of seismic travel-time and waveform analysis allow nowadays deriving also 3D-models of the Earth material from local to global scale (see, e.g., 2.7.2; Nolet, 1987; Curties and Snieder, 2002).



In the case of isotropy the number of independent parameters in the elastic tensor reduces to just two. They are called after the French physicist *Lamé* (1795-1870) the *Lamé parameters*  $\lambda$  and  $\mu$ . The latter is identical with the *shear* module.  $\lambda$  does not have a straightforward physical explanation but it can be expressed in terms of the above mentioned elastic modules and Poisson's ratio, namely

$$\lambda = \kappa - 2\mu/3 = \frac{\sigma E}{(1 + \sigma)(1 - 2\sigma)} \quad (2.2a)$$

while for the shear module holds

$$\mu = \frac{E}{2(1 + \sigma)} \quad (2.2b)$$

Accordingly, the other elastic parameters can be expressed as functions of  $\mu$ ,  $\lambda$  and/or  $\kappa$ :

$$E = \frac{(3\lambda + 2\mu)\mu}{(\lambda + \mu)} \quad (2.3)$$

and

$$\sigma = \frac{\lambda}{2(\lambda + \mu)} = \frac{3\kappa - 2\mu}{2(3\kappa + \mu)} \quad (2.4)$$

For a *Poisson solid*  $\lambda = \mu$  and thus, according to (2.4),  $\sigma = 0.25$ . Most crustal rocks have a Poisson's ratio between about 0.2 and 0.3. But  $\sigma$  may reach values of almost 0.5, e.g., for unconsolidated, water-saturated sediments, and even negative values of  $\sigma$  are possible for synthetic materials (see Tab. 2.1 in section 2.2.3).

The elastic parameters govern the velocity with which seismic waves propagate. The *equation of motion* for a continuum can be written as

$$\rho \frac{\partial^2 u_i}{\partial t^2} = \partial_j \tau_{ij} + \mathbf{f}_i, \quad (2.5)$$

with  $\rho$  - density of the material,  $u_i$  - displacement,  $\tau_{ij}$  - stress tensor and  $\mathbf{f}_i$  - the body force term that generally consists of a *gravity term* and a *source term*. The gravity term is important at low frequencies in *normal mode seismology* (see 2.4), but it can be neglected for calculations of body- and surface-wave propagation at typically observed wavelengths. Solutions of Eq. (2.5) which predict the ground motion at locations some distance away from the source are called *synthetic seismograms* (see Figs. 2.79, 2.80 and 2.82).

In the case of an inhomogeneous medium, which involves gradients in the Lamé parameters, Eq. (2.5) takes a rather complicated form that is difficult to solve efficiently. Moreover, in the case of strong inhomogeneities, transverse and longitudinal waves (see below) are not decoupled. This results in complicated particle motions. Therefore, most methods for synthetic seismogram computations ignore gradient terms of  $\lambda$  and  $\mu$  in the equation of motion by modeling the material either as a series of homogeneous layers (which also allows to approximate gradient zones; see *reflectivity method* by Fuchs and Müller, 1971; Kennett,

1983; Müller, 1985) or by assuming that variations in the Lamé parameters are negligible over a wavelength  $\Lambda$  and thus these terms tend to zero at high frequencies (*ray theoretical* approach; e.g., Červený et al., 1977; Červený, 2001; Chapman, 2002). In homogeneous media and for small deformations, however, the *equation of motion* for seismic waves outside the source region (i.e., without the source term  $\mathbf{f}_s$  and neglecting the gravity term  $\mathbf{f}_g$ ) takes the following simple form:

$$\rho \ddot{\mathbf{u}} = (\lambda + 2\mu)\nabla\nabla\cdot\mathbf{u} - \mu\nabla\times\nabla\times\mathbf{u} \quad (2.6)$$

with  $\mathbf{u}$ -displacement vector and  $\ddot{\mathbf{u}}$ , its second time derivative, the acceleration vector.

The first term on the right side of Eq. (2.6) contains the divergence of vector  $\mathbf{u}$  which is the scalar product  $\text{div } \mathbf{u} = \nabla \cdot \mathbf{u}$  of  $\mathbf{u}$  with the Nabla vector operator

$$\nabla = (\partial/\partial x)\mathbf{i} + (\partial/\partial y)\mathbf{j} + (\partial/\partial z)\mathbf{k}$$

and  $\mathbf{i}$ ,  $\mathbf{j}$ , and  $\mathbf{k}$  the components of the unity vector. and describes a volume change (compression and dilatation) which always contains some (rotation free) shearing too, unless the medium is compressed hydrostatically (as in Fig. 2.2 top)..

Further holds:  $\nabla \nabla = \nabla^2 = \Delta = \partial^2/\partial x^2 + \partial^2/\partial y^2 + \partial^2/\partial z^2$ .

In contrast, the second term on the right side of (2.6) is a vector product of the Nabla operator with a rotation or curl of  $\mathbf{u}$ .  $\text{rot } \mathbf{u} = \nabla \times \mathbf{u}$ , with the vector components

$$\begin{aligned} \text{rot } u_x &= \partial u_z/\partial y - \partial u_y/\partial z \\ \text{rot } u_y &= \partial u_x/\partial z - \partial u_z/\partial x \\ \text{rot } u_z &= \partial u_y/\partial x - \partial u_x/\partial y \end{aligned}$$

describes a change of shape without volume change (pure shearing).

Eq. (2.6) provides the basis for most body-wave synthetic seismogram calculations. Although it describes rather well most basic features in a seismic record we have to be aware that it is an approximation only for an isotropic homogeneous linearly elastic medium.

### 2.2.3 P- and S-wave velocities, waveforms and polarization

Generally, every vector field, such as the displacement field  $\mathbf{u}$ , can be decomposed into a rotation-free ( $\mathbf{u}^r$ ) and a divergence-free ( $\mathbf{u}^d$ ) part, i.e., we can write  $\mathbf{u} = \mathbf{u}^r + \mathbf{u}^d$ . Since the divergence of a curl and the rotation of a divergence are zero, we get accordingly two independent solutions for Eq. (2.6) when forming its scalar product  $\nabla\cdot\mathbf{u}$  and vector product  $\nabla\times\mathbf{u}$ , respectively:

$$\frac{\partial^2(\nabla\cdot\mathbf{u})}{\partial^2 t} = \frac{\lambda + 2\mu}{\rho} \nabla^2(\nabla\cdot\mathbf{u}^r) \quad (2.7)$$

and

$$\frac{\partial^2 (\nabla \times \mathbf{u})}{\partial^2 t} = \frac{\mu}{\rho} \nabla^2 (\nabla \times \mathbf{u}^d). \quad (2.8)$$

Eqs. (2.7) and (2.8) are solutions of the wave equation for the propagation of two independent types of seismic *body waves*, namely *longitudinal (compressional - dilatational)* P waves and *transverse (shear)* S waves. Their velocities are

$$v_p = \sqrt{\frac{\lambda + 2\mu}{\rho}} = \sqrt{\frac{\kappa + 4\mu/3}{\rho}} \quad (2.9)$$

and

$$v_s = \sqrt{\frac{\mu}{\rho}}. \quad (2.10)$$

Accordingly, for a Poisson solid with  $\lambda = \mu$   $v_p/v_s = \sqrt{3}$ . This comes close to the  $v_p/v_s$  ratio of consolidated sedimentary and igneous rocks in the Earth's crust (see Tab. 2.1). Eqs. (2.9) and (2.10) also mean that P (*primary*) waves travel significantly faster than S (*secondary*) waves and thus arrive ahead of S in seismic records (see Fig. 2.4).

Values of  $\mu$  and  $\kappa$  vary significantly depending on the highly variable composition, density (degree of compaction) and fabric of rock material both in the upper crust and in a rather systematic manner with depth in the deeper Earth interior. There, these parameters are largely pressure and temperature controlled. Accordingly, values for  $v_p$  range between an approximate average of 5.8 km/s in the upper crust and 13.7 km in the lowermost mantle and for  $v_s$  between 0 km/s in the liquid outer core, about 3.4 for the upper crust and 7.3 km/s in the lowermost mantle. For details as a function of depth see the 1-D Earth models in DS 2.1 and Figure 2.76 in section 2.7.1 of this Chapter. Of interest may be, although seismologically irrelevant, that diamonds have the highest P-wave velocity of 18.0 km/s of all natural materials.

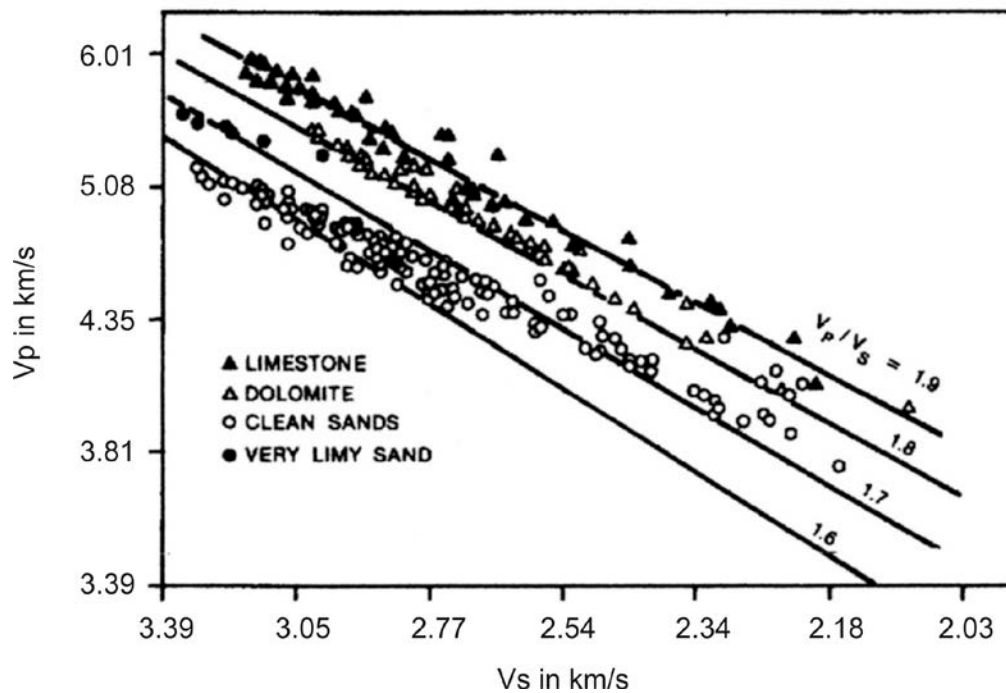
For near surface materials, ranging from water over muddy or unconsolidated clastic rocks via porous consolidated to almost pore-free outcropping ultramafic hard rocks, the range of  $v_p$  and  $v_s$  variations is comparably large yet with significantly smaller absolute values of  $v_p$  and  $v_s$  than for crustal averages (see Tab. 2.1 and Fig. 2.3). According to Eqs. (2.5), (2.9) and (2.10) the Poisson's ratio  $\sigma$  depends on  $v_p$  and  $v_s$ . When knowing the ratio  $v_p/v_s$ ,  $\sigma$  can be derived from the relationship

$$\sigma = (v_p^2/v_s^2 - 2) / 2(v_p^2/v_s^2 - 1) \quad (2.11)$$

or, when knowing the Poisson's ratio of rock material, the related  $v_p/v_s$  can be inferred. Tab. 2.1 gives approximate average values for  $\sigma$  for some Earth, natural and synthetic materials. Where available, they have been complemented by the related elastic modules  $\kappa$  and  $\mu$ , the density  $\rho$  and/or the seismic velocities  $v_p$  and  $v_s$ . Additionally, Fig. 2.3 presents some of the pioneering early results by Pickett (1963) which reveal distinct differences of the ratio  $v_p/v_s$  for limestone, dolomite and clean sandstone, thus demonstrating the usefulness of  $v_p/v_s$  as a lithological indicator. For shallow seismic surveys in conjunction with engineering seismological and microzonation/site effect investigations this may be of great help.

**Tab. 2.1** Values (averages and/or approximate ranges) of elastic constants, density, Poisson's ratio and seismic wave velocities for some selected materials, unconsolidated sediments, sedimentary rocks of different geologic age and igneous/plutonic rocks. Values for granite relate to 200 MPa confining pressure, corresponding to about 8 km depth, for basalt to 600 MPa (about 20 km depth), and for Peridotite, Dunite and Pyroxenite to 1000 MPa (about 30 km depth) (compiled from Hellwege, 1982; Castagna et al., 1985; Lillie, 1999; and Wikipedia). Ranges are given in brackets.

| <b>Material or Geologic Formation</b>   | <b>Bulk Module<br/>in <math>10^9</math> Pa</b> | <b>Shear Module<br/>in <math>10^9</math> Pa</b> | <b>Density<br/>in <math>\text{kg m}^{-3}</math></b> | <b>Poisson Ratio</b>                       | <b><math>v_p</math><br/>in <math>\text{km s}^{-1}</math></b> | <b><math>v_s</math><br/>in <math>\text{km s}^{-1}</math></b> | <b><math>v_p/v_s</math></b> |
|-----------------------------------------|------------------------------------------------|-------------------------------------------------|-----------------------------------------------------|--------------------------------------------|--------------------------------------------------------------|--------------------------------------------------------------|-----------------------------|
| <b>Air</b>                              | 0.0001                                         | 0                                               | 1.0                                                 | 0.5                                        | 0.32                                                         | 0                                                            | $\infty$                    |
| <b>Water</b>                            | 2.2                                            | 0                                               | 1000                                                | 0.5                                        | 1.5                                                          | 0                                                            | $\infty$                    |
| <b>Ice</b>                              |                                                |                                                 |                                                     | (0.3-0.47)                                 | 3.25 at $-4^\circ$                                           |                                                              |                             |
| <b>Diamond</b>                          |                                                |                                                 |                                                     |                                            | <b>18.0</b>                                                  |                                                              |                             |
| <b>Rubber</b>                           |                                                |                                                 |                                                     | <b>0.50</b>                                |                                                              |                                                              |                             |
| <b>Foam material</b>                    |                                                |                                                 |                                                     | (0.10-0.40)                                |                                                              |                                                              |                             |
| <b>Cork</b>                             |                                                |                                                 |                                                     | <b><math>\approx 0.00</math></b>           |                                                              |                                                              |                             |
| <b>Fiber</b> (compound material)        |                                                |                                                 |                                                     | <b>(0.05 to 0.55)</b>                      |                                                              |                                                              |                             |
| <b>Metallic raw material.</b>           |                                                |                                                 |                                                     | <b>(<math>\approx 0.25</math> to 0.35)</b> |                                                              |                                                              |                             |
| <b>Sedimentary rock (clastic)</b>       |                                                |                                                 |                                                     |                                            | (1.4-5.3)                                                    |                                                              |                             |
| <b>Sand &amp; Sandstone (dry)</b>       | 24                                             | 17                                              | 2500                                                | 0.21<br>(0.20-0.45)                        | 4.3<br>(0.3-5.5)                                             | 2.6<br>(0.2-3.5)                                             | 1.65                        |
| <b>Shaly rock</b>                       |                                                |                                                 |                                                     |                                            | (3.0-5.2)                                                    | (1.5-3.5)                                                    |                             |
| <b>Clay</b>                             |                                                |                                                 |                                                     | (0.30-0.45)                                | $\approx 3.3$                                                | $\approx 1.6$                                                | $\approx 2.1$               |
| <b>Salt</b>                             | 24                                             | 18                                              | 2200                                                | 0.17                                       | 4.6<br>(3.8-5.2)                                             | 2.9                                                          | 1.59                        |
| <b>Limestone <math>\emptyset</math></b> | 38                                             | 22                                              | 2700                                                | 0.19                                       | 4.7<br>(2.9-5.6)                                             | 2.9                                                          | 1.62                        |
| <b>Granite</b>                          | 56<br>(47-69)                                  | 34<br>(30-37)                                   | 2610<br>(2340-2670)                                 | 0.25<br>(0.20-0.31)                        | 6.2<br>(5.8-6.4)                                             | 3.6<br>(3.4-3.7)                                             | 1.73<br>(1.65-1.91)         |
| <b>Basalt</b>                           | 71<br>(64-80)                                  | 38<br>(33-41)                                   | 2940<br>(2850-3050)                                 | 0.28<br>(0.26-0.29)                        | 6.4<br>(6.1-6.7)                                             | 3.6<br>(3.4-3.7)                                             | 1.80<br>(1.76-1.82)         |
| <b>Peridotite, Dunite, Pyroxenite</b>   | 128<br>(113-141)                               | 63<br>(52-72)                                   | 3300<br>(3190-3365)                                 | 0.29<br>(0.26-0.29)                        | 8.0<br>(7.5-8.4)                                             | 4.4<br>(4.0-4.7)                                             | 1.8<br>(1.76-1.91)          |
| <b>Metamorphic/igneous rock</b>         |                                                |                                                 |                                                     |                                            | (3.8-6.4)                                                    |                                                              |                             |
| <b>Ultramafic rock</b>                  |                                                |                                                 |                                                     |                                            | (7.2-8.7)                                                    |                                                              |                             |
| <b>Cenozoic</b>                         |                                                |                                                 | (1500-2100)                                         | (0.38-<0.5)                                | (0.2-1.9)                                                    |                                                              | (2.3 – 8)                   |
| <b>Cenozoic (saturated)</b>             |                                                |                                                 | 1950                                                | 0.48                                       | 1.7                                                          | 0.34                                                         | 5                           |
| <b>Cretaceous &amp; Jurassic</b>        |                                                |                                                 | (2400-2500)                                         | (0.28-0.43)                                |                                                              |                                                              | (1.8 – 2.8)                 |
| <b>Triassic</b>                         |                                                |                                                 | (2500-2700)                                         | (0.28-0.40)                                |                                                              |                                                              | (1.8 – 2.5)                 |
| <b>Upper Permian</b>                    |                                                |                                                 | (2000-2900)                                         | (0.23-0.31)                                |                                                              |                                                              | (1.7 – 1.9)                 |
| <b>Carboniferous</b>                    |                                                |                                                 |                                                     | (0.31-0.35)                                |                                                              |                                                              | (1.9 – 2.1)                 |



**Fig. 2.3** Relationship between  $v_p$  and  $v_s$  of limestone, dolomite, and sandstone according to data published by Pickett (1993), rescaled to SI units and velocity instead of slowness. Note the non-linearity in the velocity scales due to the rescaling.

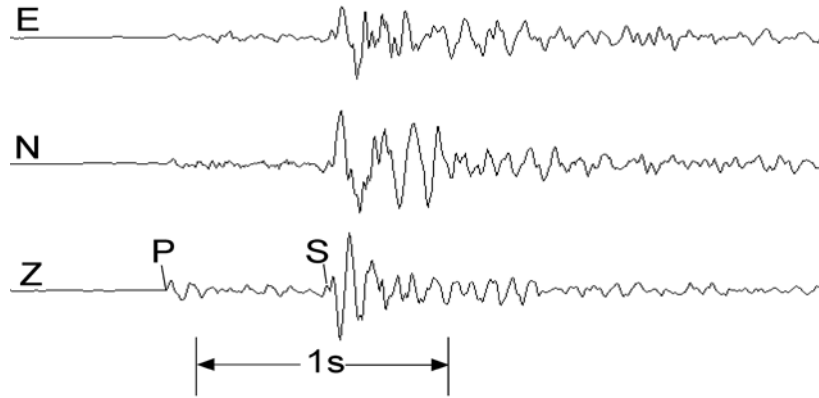
Comments and conclusions drawn from Tab. 2.1 and Fig. 2.5 are:

- The Poisson's ratio varies for isotropic materials between 0 and 0.5;
- If  $\sigma = 0.5$  then the volume of an elastic body remains constant under deformation;
- If  $\sigma < 0.5$  then the volume increases under tensional load and shrinks under pressure;
- If  $\sigma > 0.5$  then the volume shrinks under tensional load. This may be the case, e.g., for some porous materials such as polymer foams or carbon fiber compounds depending on the fiber orientation;
- Negative values of  $\sigma$  are possible only in some rare synthetic materials, resulting, in the case of length stretching (as in Fig. 2.2 below) in a cross-extension instead of a length extension (so-called auxetic behavior) ;
- For the same material, shear waves travel always slower than compressional waves;
- The higher the rigidity of the material, the higher the P- and S-wave velocities;
- The rigidity usually increases with density  $\rho$ , but more rapidly than  $\rho$ . This explains why denser rocks have normally faster wave propagation velocities although  $v^2 \sim 1/\rho$  ;
- Fluids (liquids or gasses) have no shear strength ( $\mu = 0$ ) and thus do not propagate shear waves;
- For the same material, compressional waves travel slower through its liquid state than through its solid state (e.g., water and ice, or, in the Earth's core, through the liquid outer and solid inner iron core, respectively).

More specifically, Castagna et al. (1985) documented for clastic silicate rocks a very large range of velocity variations depending on composition, grain size, porosity, fluid saturation etc. E.g.,  $v_p$  and  $v_s$  vary for geopressed shaly rocks between 3.0-5.2 km/s and 1.5-3.5 km/s and for dry sand and sandstones, depending on the degree of compaction, cementation

material, porosity and fluid saturation in an even wider range ( $v_p = 0.3 - 5.5$  km/s and  $v_s = 0.2 - 3.5$  km/s). When normalizing bulk and shear module to density and comparing dry and liquid saturated porous rocks for equal bulk module then the shear module of dry rock is significantly higher than for water saturated rocks, whereas for equal P-wave velocity the bulk module of dry rock is usually significantly lower than for saturated rock. The latter is an important indicator in exploration seismology for saturated reservoir rocks. For more details we refer to the most recent textbook on physical properties of rocks by Schön (2011).

Fig. 2.4 shows a record of a local earthquake which clearly shows the first arriving P and the later arriving S onset (with larger amplitudes and longer periods).

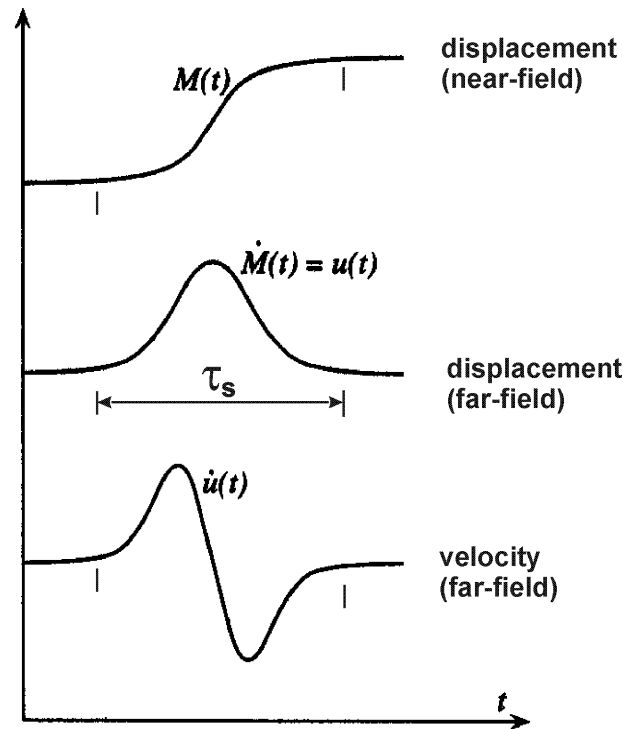


**Fig. 2.4** The three components of ground-velocity proportional digital records of the P and S waves from a local event, an aftershock of the Killari-Latur earthquake, India (18.10.1993), at a hypocenter distance of about 5.3 km.

Note the simple *transient waveform* (*wavelet*) of P in the Z-component of Fig. 2.3. The waveform and duration of the primary body wave is related to the shape and duration of the source-time function. It is for an earthquake shear rupture usually a more or less complex displacement step (Fig. 2.5) which can be described by the moment-release function  $M(t)$ . In the *far-field*, i.e., at distances larger than the source dimension and several wavelengths of the considered signal, the related displacement  $u(t)$  looks, in the idealized case, bell-shaped and identical with the moment-rate  $\dot{M}(t)$  (or velocity source-time) function (Fig. 2.5 middle). The far-field pulse shape is given by eq. (10.13) in Aki and Richards (2002), provided fault length  $L$ , wavelength  $\lambda$  and source-receiver distance  $r_0$  satisfy the constraint  $L^2 \ll 0.5 \lambda r_0$ . The base-width of this *far-field displacement source pulse*  $u(t)$  corresponds to the duration of displacement at the source (see moment rate source-time functions in Fig. 3.9 of Chapter 3). However, usually modern broadband seismometers record ground velocity  $\dot{u}(t)$  instead of ground displacement. The recorded waveform then looks similar to the ones seen in Fig. 2.4 and Fig. 2.5 bottom. The period of the wavelet  $\dot{u}(t)$  then corresponds to the duration of the displacement of the source,  $\tau_s$ .

This waveform of primary body waves will be slightly changed due to frequency-dependent attenuation and other wave-propagation effects, e.g., those that cause phase shifts (see section 2.5.4.3). But the duration of the body-wave ground-motion wavelet (or wave-group) will remain essentially that of the source process, independent of the observational distance, unless

it is significantly prolonged and distorted by the coda of scattered waves (e.g., Figs. 2.42 and 2.43) and/or by narrowband seismic recordings (see Chapter 4.2). We have made this point in order to better appreciate one of the principal differences in the appearance in seismic records of transient non-dispersed body waves on the one hand and of dispersed surface waves (see 2.3 and Figs. 2.12 and 2.13) on the other hand.

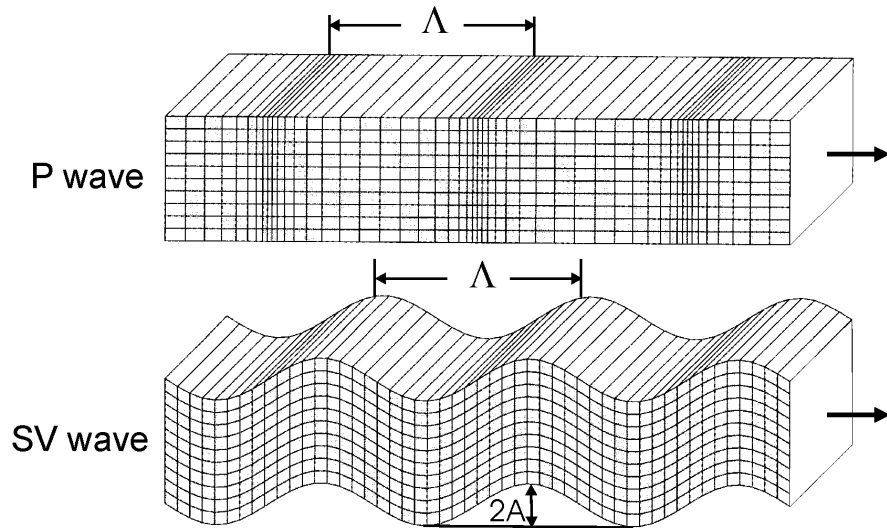


**Fig. 2.5** Relationship between near-field displacement, far-field displacement and velocity from isotropic or double-couple source earthquake shear sources (modified from Shearer, Introduction to Seismology, 1999; with permission from Cambridge University Press).

Fig. 2.6 depicts (exaggerated) the kind of displacements occurring from harmonic plane P and S waves. One clearly recognizes that P waves involve both a volume change and shearing (change in shape) while S-wave propagation is pure shear with no volume change. The P-wave particle motion is back and forth in the radial (R) direction of wave propagation (*longitudinal polarization*) but that of the S wave is perpendicular (*transverse*) to it, in the given case oscillating up and down in the vertical plane (SV-wave). However, S waves may also oscillate purely in the horizontal plane (SH waves) or at any angle between vertical and horizontal, depending on the source mechanism (Chapter 3), the wave propagation history, and the incidence angle  $i_0$  at the seismic station (see Fig. 2.27).

The wavelength  $\Lambda$  is defined by the distance (in km) between neighboring wave peaks or troughs or volumes of maximum compression or dilatation (see Fig. 2.6). The wave period  $T$  is the duration of one oscillation (in s) and the frequency  $f$  is the number of oscillations per second (unit [Hz] = [s<sup>-1</sup>]). The wavelength is the product of wave velocity  $v$  and period  $T$  while the wavenumber is the ratio  $2\pi/\Lambda$ . Tab. 2.2 summarizes all these various *harmonic wave parameters* and their mutual relationship.



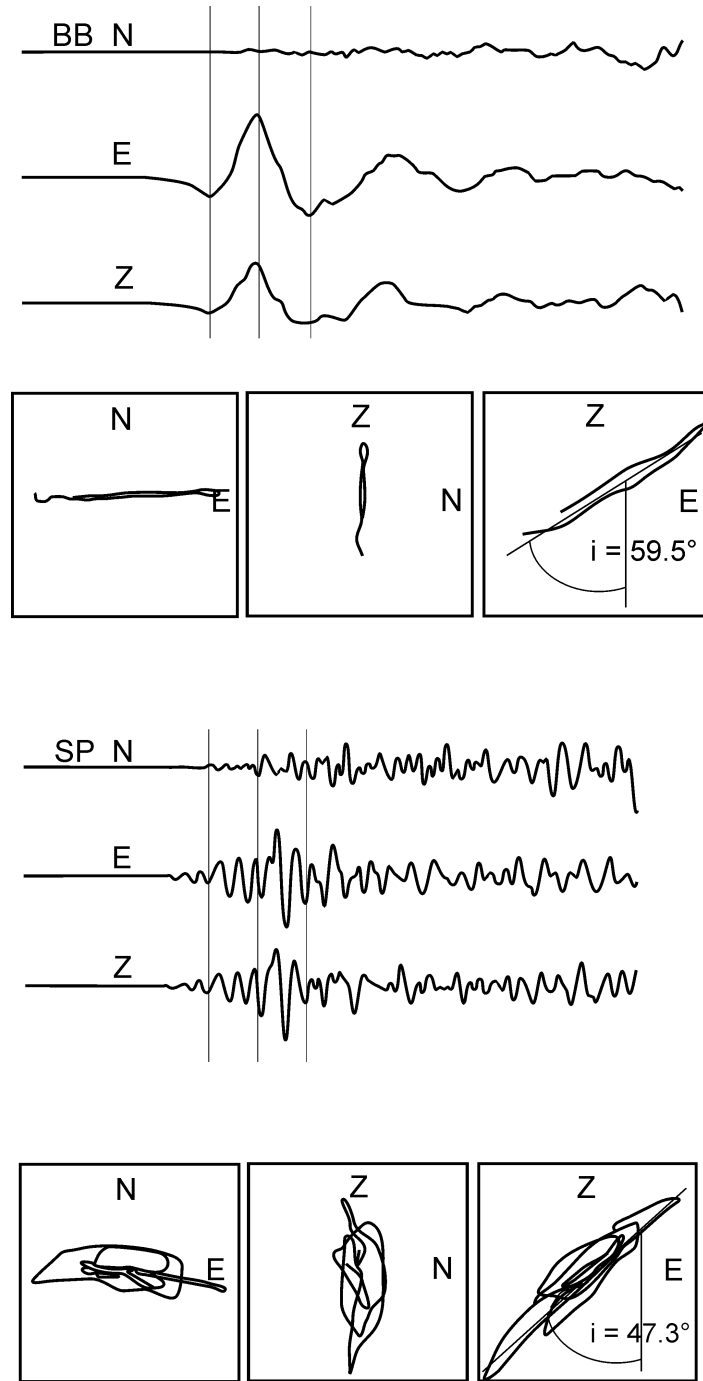


**Fig. 2.6** Displacement patterns of a harmonic plane P wave (top) and SV wave (bottom) propagating in a homogeneous isotropic medium.  $\Lambda$  is the wavelength and  $2A$  means double amplitude. The white surface on the right is a segment of the propagating plane wavefront where all particles undergo the same motion at a given instant in time, i.e., they oscillate *in phase*. The arrows indicate the seismic rays, defined as the *normal* to the wavefront, which points in the direction of propagation (modified according to Shearer, Introduction to Seismology, 1999; with permission from Cambridge University Press).

**Tab. 2.2** Harmonic wave parameters and their mutual relationship.

| Name              | Symbol    | Relationships                                |
|-------------------|-----------|----------------------------------------------|
| Period            | $T$       | $T = 1/f = 2\pi/\omega = \Lambda/v$          |
| Frequency         | $f$       | $f = 1/T = \omega/2\pi = v/\Lambda$          |
| Angular frequency | $\omega$  | $\omega = 2\pi f = 2\pi/T = v \cdot k$       |
| Velocity          | $v$       | $v = \Lambda/T = f \cdot \Lambda = \omega/k$ |
| Wavelength        | $\Lambda$ | $\Lambda = v/f = v \cdot T = 2\pi/k$         |
| Wavenumber        | $k$       | $k = \omega/v = 2\pi/\Lambda = 2\pi f/v$     |

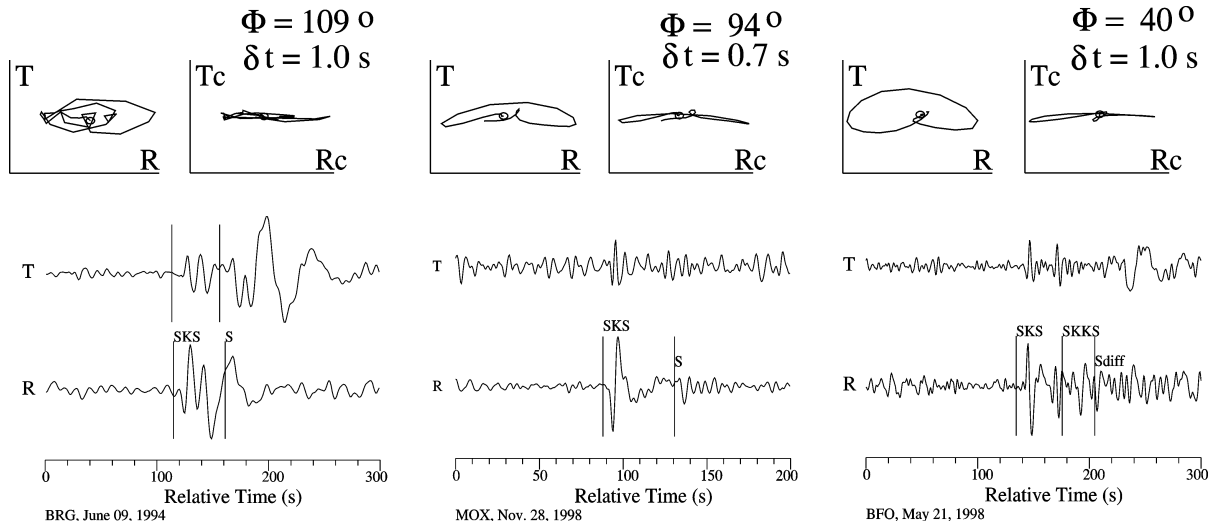
In any case, the polarization of both P and S waves, when propagating in a homogenous and isotropic medium, is linear. This is confirmed rather well by particle motion analysis of real seismic recordings, if they are broadband (or long period). But higher frequencies, which are more strongly affected by local inhomogeneities in the Earth, show a more elliptical or irregular particle motion. Fig. 2.7 shows an example. While the rectilinearity of P is almost 1 (0.95) in the BB record it is significantly less (0.82 as an average over 5 oscillations and down to 0.68 for some single oscillations) for the short-period filtered record.



**Fig. 2.7** 3-component records at station MOX (top traces) and related plots of particle motion in the horizontal (N-E) plane and two vertical planes (Z-N and Z-E, respectively) of the P-wave onset from a local seismic event (mining collapse) in Germany (13.03.1989;  $M_l = 5.5$ ; epicenter distance  $D = 112$  km, backazimuth  $BAZ = 273^\circ$ ). Upper part: broadband recording (0.1-5 Hz); lower part: filtered short-period recording (1- 5 Hz). **Note:** The incidence angle is  $59.5^\circ$  for the long-period P-wave oscillation and  $47.3^\circ$  for the high-frequency P-wave group.

S waves are also linearly polarized when propagating in homogeneous isotropic medium. However, in the presence of anisotropy, they split into a fast and slow component. These split waves propagate with different velocity that causes some time delay and related phase shift.

Accordingly, the two split S-wave components superimpose to an elliptical polarization (Fig. 2.8). The orientation of the main axis and the degree of ellipticity are controlled by the fast and slow velocity directions of the medium with respect to the direction of wave propagation and the degree of anisotropy. Therefore, shear-wave splitting is often used to study S-wave velocity anisotropy in the Earth (e.g., Vinnik et al., 1989a and b; Silver and Chan, 1991; Bormann et al., 1993).



**Fig. 2.8** Examples of SKS and SKKS recordings and plots of particle motion at three stations of the German Regional Seismograph Network. The horizontal radial (R) and transverse (T) components are shown. They were derived by rotation of the N-S and E-W horizontal components with the backazimuth angle. The T component at BFO has the same scale as the R component, while T is magnified two-fold relative to R at BRG and MOX. The top panels show the polarization in the R-T plane. Anisotropy is manifested in all three cases by the elliptical polarization. Linear polarization is obtained by correcting the R-T seismograms for the anisotropy effect using an anisotropy model where the direction of the fast shear wave is sub-horizontal and given by the angle  $\Phi$  measured clockwise from north, and the delay time (in seconds) between the slow and the fast shear wave is given by  $\delta t$  (courtesy of G. Bock †).

## 2.3 Surface waves (P. Bormann)

### 2.3.1 Origin

So far we have considered only body-wave solutions of the seismic wave equation. They exist in the elastic full space. However, in the presence of a free surface, as in the case of the Earth, other solutions are possible. They are called *surface waves*. There exist two types of surface waves, *Love waves* and *Rayleigh waves*. While Rayleigh (LR or R) waves exist at any free surface, Love (LQ or G) waves require some kind of a near surface *wave guide* formed by a velocity increase with depth (gradient- or layer-wise). Both conditions are fulfilled in the real Earth.

SH waves are totally reflected at the free surface. Love waves are formed through constructive interference of repeated reflections of teleseismic SH at the free-surface (i.e., S3, S4, S5, etc.; see Figs. 2.58 and 2.59). They can also result from constructive interference

between SH waves, which are *post-critically reflected* (see 2.5.3.5) within a homogeneous layer (or a set of  $i$  layers with increasing  $v_{si}$ ) overlaying a half-space of higher velocity. The latter is the case of crustal layers, overlaying the upper mantle with a significant velocity increase at the base of the crust, called the “Mohorovičić-discontinuity” or *Moho* for short. The Moho marks the transition between the usually more mafic (often termed “basaltic”) lower crust and the peridotitic uppermost mantle (for related velocities see Tab. 2.1) and may, together with other pronounced intra-crustal velocity discontinuities give rise to the formation of complex guided crustal waves (see 2.3.3).

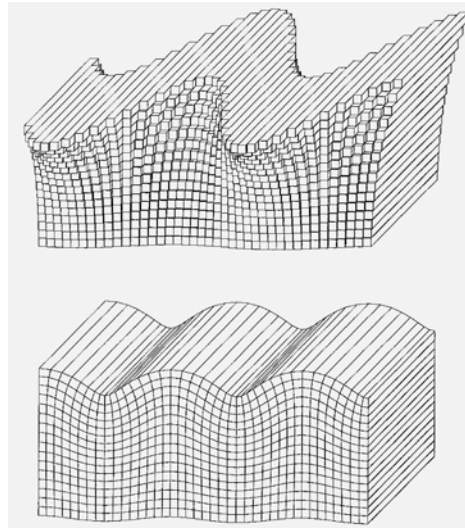
Generally, destructive interference of the upgoing and downgoing reflected SH waves will occur, except at certain discrete frequencies  $\omega$  and  $n$  multiples of it (with  $n$  as an integer). The values of  $\omega$  given for  $n = 0$  are termed the *fundamental modes* while larger values of  $n$  define the *higher modes* or *overtones*. Fig. 2.9 (top) shows the horizontal (SH type) of displacement and linear polarization of the fundamental Love-wave mode as well as the exponential decay of its amplitudes with depth.

When an SV (or P) wave arrives at the surface the reflected wave energy contains (because of *mode conversion*, see 2.5.3.4) both P and SV waves. Lord Rayleigh (1887) showed that a solution of the *wave equation* exists for two coupled *inhomogeneous* P and SV waves that propagate along the surface of a half-space. Physically, inhomogeneous waves are characterized by a *phase velocity* which is slower than the underlying body-wave speed and by amplitudes that grow or decay exponentially in directions of constant phase. P waves become inhomogeneous wave if the depth component of their *slowness*  $s$  is imaginary, i.e., when their *wave parameter*  $p = \sin i/v > 1/v_p$  (with  $i$  = *incidence angle*; for definition of terms and related formulas see section 2.5.2.1). Similarly, S waves form an inhomogeneous wave, when  $p > 1/v_s$ . Then the angle  $i$  is no longer real and it is no longer clear what the incidence angle of the wave is. Example: If for the upper crust  $v_p = 5$  km/s and  $v_s = 3$  km/s, then at an incidence angle of  $36.9^\circ$  for an SV wave at the Earth surface part of its energy is converted into a P wave which travels parallel to the surface. i.e., at an “incidence angle” of  $90^\circ$ . With further increase of  $i_{SV}$  and thus of the slowness  $p_{SV}$  the angle  $i_p$  can not grow further, it becomes imaginary and the physical displacement shows a phase shift of amount  $\phi$ , where  $\phi = \text{phase}(SP)$  for the horizontal component and  $\phi = \text{phase}(SP) + \pi/2$  for the vertical component, resulting in inhomogeneous wave with a retrograde elliptical particle motion. And further, if at the free surface of a half-space the condition  $1/v_p < 1/v_s < p$  holds then both P and SV waves become inhomogeneous. The coupled pair of these two inhomogeneous waves forms the Rayleigh waves.

While Rayleigh waves show no dispersion in a homogeneous half-space (with constant  $v_p$  and  $v_s$ ), they are always dispersed in media with layering and/or velocity gradients such as in the real Earth. Rayleigh waves travel - for a Poisson solid - with a phase velocity  $c = \sqrt{2 - 2/\sqrt{3}} v_s \approx 0.92 v_s$ , i.e., slightly slower than Love waves and thus have a larger slowness, which satisfies the condition  $1/v_p < 1/v_s < p$ . However, the exact value of  $c$  depends on  $v_p$  and  $v_s$  and is presented analytically by Malischewsky (2004).

Since Rayleigh waves originate from coupled inhomogeneous P and SV waves they are polarized in the vertical (SV) plane of propagation and due to the phase shift between P and SV the sense of their *particle motion* at the surface is *elliptical* and *retrograde* (counter clockwise). However, below a certain depth, which depends on frequency, the particle motion

is dominated by the SV component and becomes prograde elliptical (see Fig. 2.14 in the next section). Fig. 2.9 (bottom) shows schematically the displacement patterns of fundamental mode of Rayleigh waves with amplitudes decaying exponentially with depth. The short-period fundamental mode of Rayleigh type in continental areas is termed Rg.

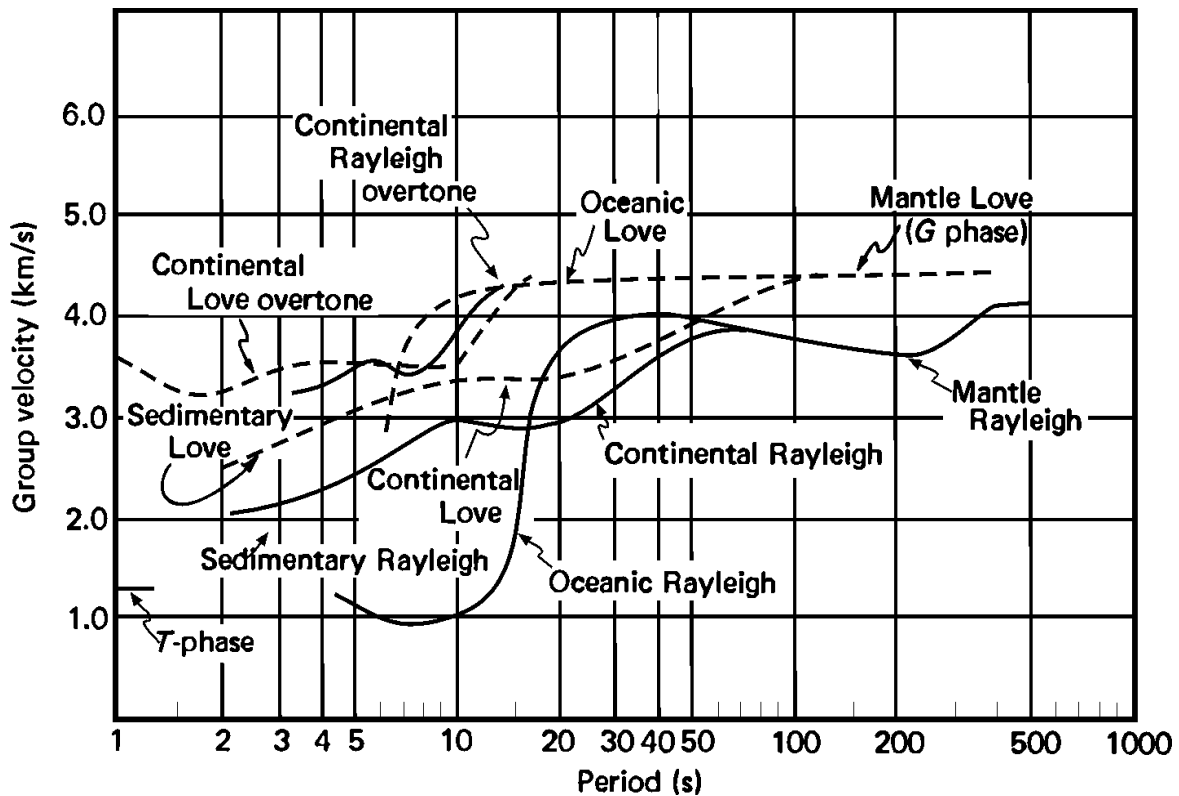


**Fig. 2.9** Displacement patterns caused by horizontally propagating fundamental Love (top) and Rayleigh waves (bottom). In both cases the wave amplitudes decay exponentially with depth (from Shearer, Introduction to Seismology, 1999; with permission from Cambridge University Press).

### 2.3.2 Dispersion and polarization

The *penetration depth* below the surface increases with  $\Lambda$ . This is comparable with the frequency-dependent *skin effect* of electromagnetic waves propagating in a conducting medium with a free surface. Since the types of rocks, their rigidity and bulk module change (usually increase) with depth, the velocities of surface waves change accordingly since the longer waves “sense” deeper parts of the Earth. This results in a frequency dependence of their horizontal propagation velocity, called *dispersion*. Accordingly, while body-wave arrivals with no or negligibly small dispersion only (due to intrinsic attenuation) appear in seismic records as rather impulsive onsets or short transient wavelets (with the shape and duration depending on the bandwidth of the seismograph; see Chapter 4.2 “?”), the dispersion of surface waves forms long oscillating wave trains. Their duration increases with distance.

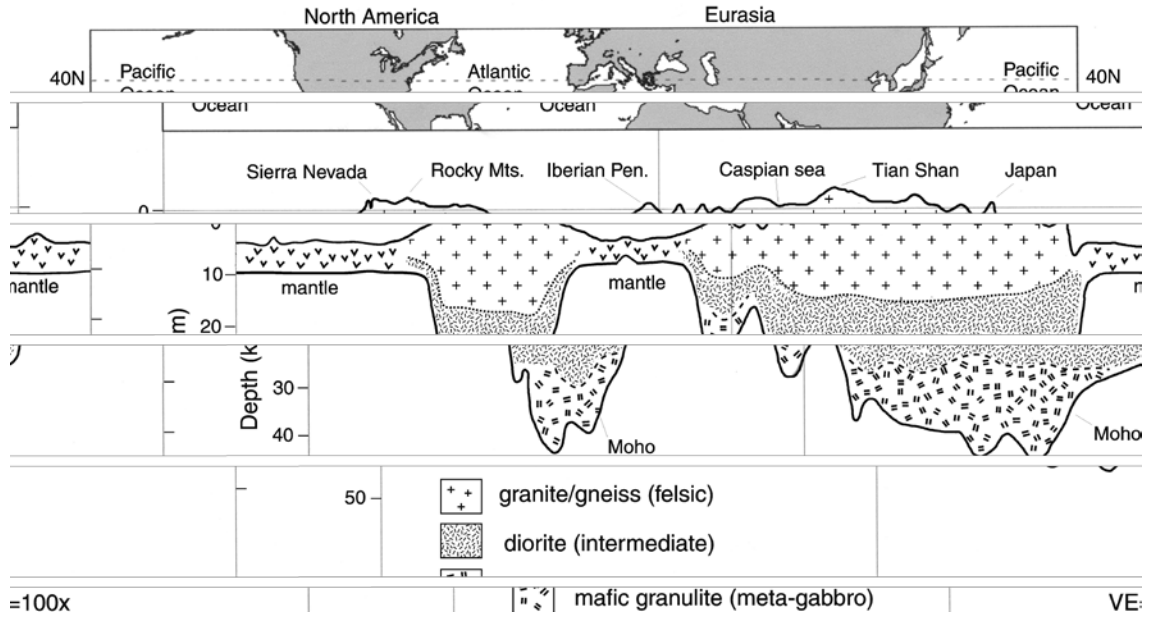
Usually, the more long-period surface waves arrive first (normal dispersion). But in some regions of the Earth low-velocity layers exist (e.g., the *asthenosphere* in the upper mantle; see the PREM model in 2.7, Fig. 2.77, in the depth range between about 80 and 220 km). This general trend may then be reversed for parts of the surface wave spectrum. Presentations of the propagation velocity of surface waves as a function of the period  $T$  or the frequency  $f$  are called *dispersion curves*. They differ for Love and Rayleigh waves and also depend on the velocity-depth structure of the Earth along the considered segment of the travel path (Fig. 2.10). Thus, from the inversion of surface wave dispersion data, information on the shear-wave velocity structure of the crust, and, when using periods up to about 500 s (mantle surface waves), even of the upper mantle and the transition zone to the lower mantle can be derived.



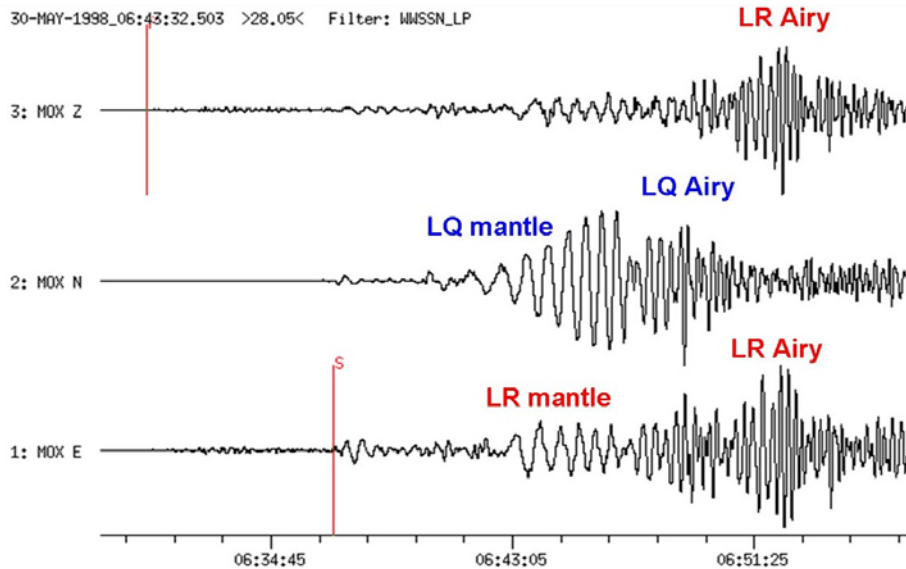
**Fig. 2.10** Group-velocity dispersion curves as a function of period for Love and Rayleigh waves (fundamental modes and overtones) (from Bullen and Bolt, *An Introduction to the Theory of Seismology*, 1985; with permission from Cambridge University Press).

The large differences in crustal thickness, composition and velocities between oceanic and continental areas (Fig. 2.11) result in significant differences between their related average group-velocity dispersion curves (Fig. 2.10). They are particularly pronounced for Rayleigh waves. While the velocities for continental Rayleigh waves vary in the period range from about 15 and 30 s only from 2.9 to 3.3 km/s, they vary much more in oceanic areas (from about 1.5 to 4.0 km/s within the same period range). Consequently, LR wave trains from travel paths over continental areas are shorter and look more clearly dispersed because the various periods follow each other at shorter time differences.

Fig. 2.12 presents an earthquake record at a *backazimuth* of  $BAZ = 85^\circ$ . This results in a rather good separation of the Rayleigh (LR) wave, recorded best in the E-W and Z component, and the Love wave (LQ), with largest amplitudes in the N-S component. Clearly recognizable are the large amplitude *Airy phases* for both waves with periods around 15 s. They are due to the group velocity minimum for continental travel paths (see Fig. 2.10). The preceding more long-period waves begin with periods around 50 s for LR and 90 s for LQ. These are already surface waves with a penetration depth into the upper mantle. The pronounced “normal” dispersion that follows is due to the slowly decreasing group velocities for both LQ and LR for periods  $T < 60$  s (LR), respectively  $T < 100$  s (LQ) until they reach their minima at periods around 15 s.



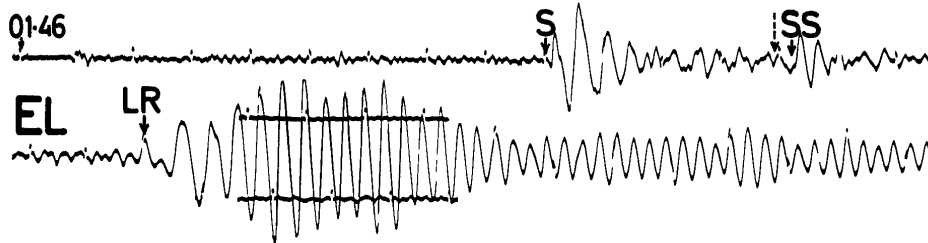
**Fig. 2.11** Cross-section through the crust along 40° northern latitude. Note the different signatures for the upper (granitic), intermediate (dioritic) and lower (mafic) continental crust and the basaltic oceanic crust. The crustal base is termed “Moho” (according to its discoverer, the Croatian seismologist Andrija Mohorovičić). The P-wave velocity increases at the Moho from about 6.5-6.8 km/s to 7.8-8.2 km/s. The continental crust is about 25 to 45 km thick (average about 35 km) and has “roots” under young high mountain ranges which may reach down to nearly 70 km. The oceanic crust is rather thin (about 8 to 12 km) with a negligible upper crust (courtesy of Mooney and Detweiler, 2002).



**Fig. 2.12** Three-component long-period seismogram (WWSSN-LP simulation filter) recorded from a shallow earthquake at the Afghanistan-Tadjikistan border region (30 May 1998,  $M_s = 6.9$ ) at station MOX ( $D = 43.2^\circ$ ,  $BAZ = 85^\circ$ ). Note that the displacement magnification of the WWSSN-LP at  $T = 50$  s and 90 s is already 3 times, respectively 9 times reduced as compared to the magnification at 15 s period (see Figure 1 in IS 3.3).

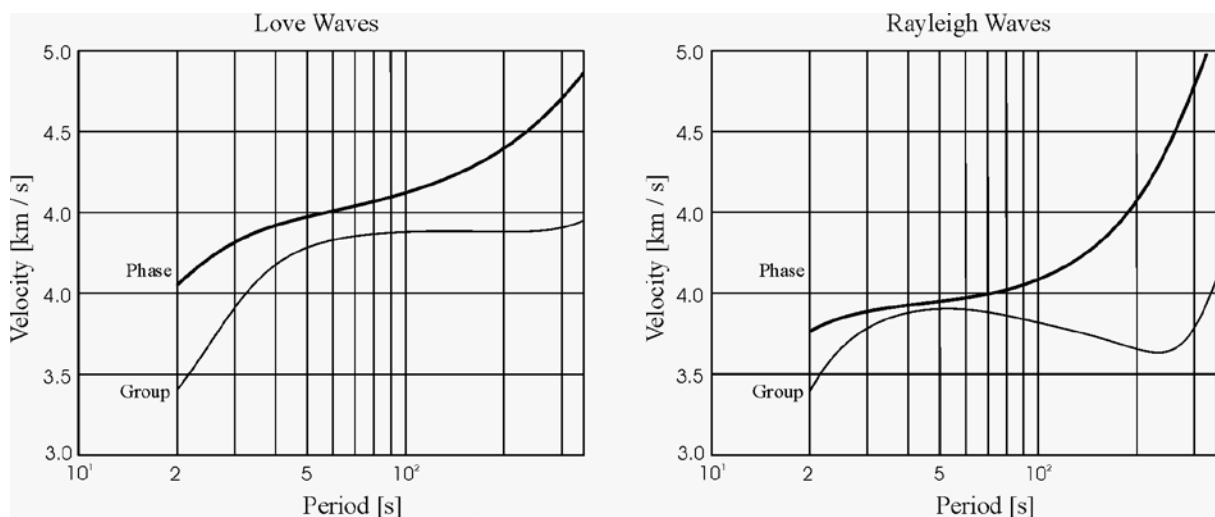


In contrast, LR wave trains with dominantly oceanic travel paths are much longer with almost monochromatic oscillations with periods between some 15 to 25 s over many minutes (Fig. 2.13), which is related to the steeply increasing group velocity branch for oceanic LR waves in this period range with velocities increasing from about 1.2 km/s to 3.8 km/s. Actually, the discovery of different surface-wave velocities along continental and oceanic paths were in the 1920s the first indication of the principle structural difference between oceanic and continental crust.



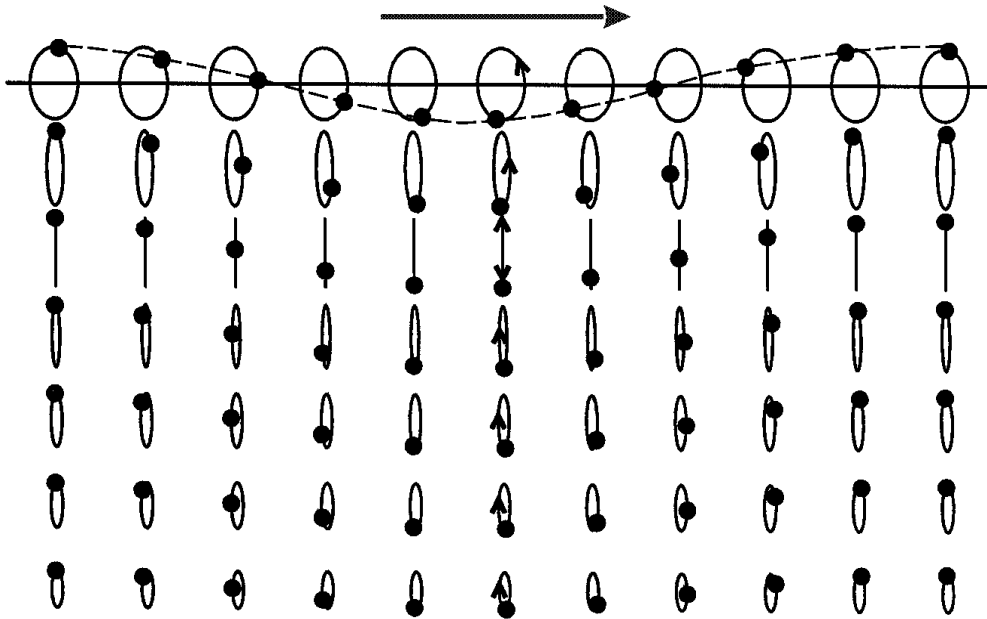
**Fig. 2.13** Record of Rayleigh waves in the long-period vertical component at the station Toledo, Spain, from an earthquake in the Dominican Republic ( $D = 6,622$  km; travel-path through the Atlantic Ocean) The largest amplitudes have a period of 25 s shortly after the LR onset and later, towards the end of the reproduced trace, a period of 16 s (copied cut-out from p. 35 of G. Payo, 1986).

Strictly speaking, when dealing with dispersive waves, one has to discriminate between the *group velocity*  $U(T)$ , with which the energy of the wave group moves and the *phase velocity*  $c(T)$ , with which the wave peaks and troughs at a given frequency travel along the surface. As seen from Fig. 2.14,  $c(T)$  is always increasing with period and larger than  $U(T)$ . When comparing Figs. 2.11 and 2.14 the significant differences between dispersion curves calculated for a global 1-D Earth model like PREM (see 2.7, Fig. 2.79 and DS 2.1) and averaged measured curves for different types of crust or mantle models become obvious.



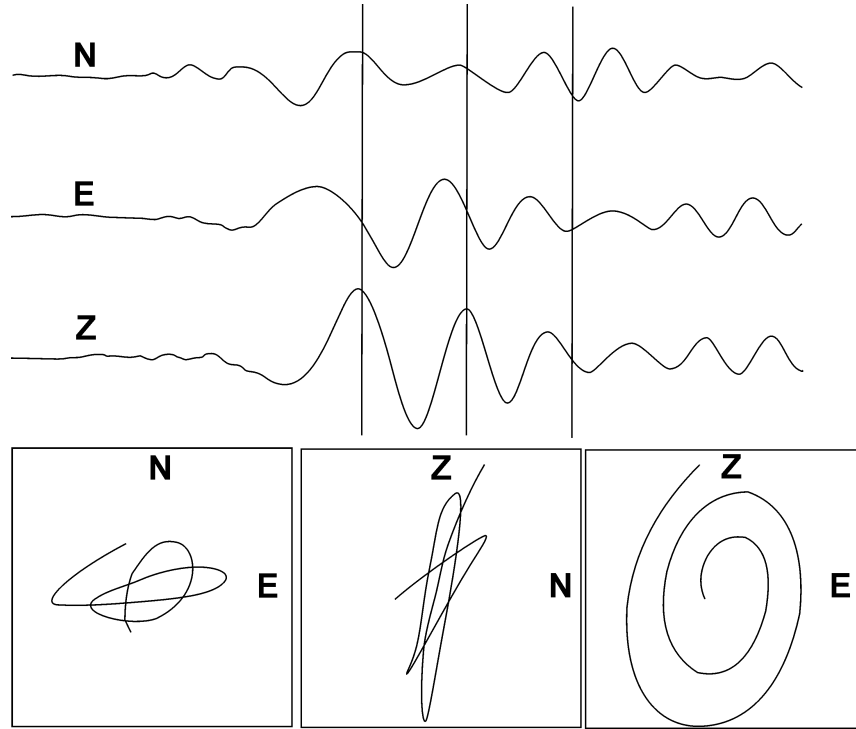
**Fig. 2.14** Fundamental mode Love- and Rayleigh-wave dispersion curves computed for the PREM model with anisotropy (courtesy of Gabi Laske).

As shown in Fig. 2.15, the particle motion of the fundamental Rayleigh-wave mode on the surface of a uniform half space is *retrograde* and the vertical component about 1.5 times larger than the horizontal one. The horizontal component will vanish at a depth of about  $\Lambda/5$  and thus the particle motion becomes vertical linear. At larger depth the particle trajectories will be elliptical again, but because of a now dominant SV component with a *prograde* (clockwise) sense of motion. The amplitudes decay rapidly with depth. At a depth of  $\Lambda/2$ , the horizontal particle motion is about 10% of that at the surface while the vertical particle motion at the surface is reduced to about 30%.



**Fig. 2.15** Particle motion for the fundamental Rayleigh mode in a uniform half-space. Shown is one horizontal wavelength. Note the change from retrograde to prograde sense of elliptical particle motion at a depth larger about  $\Lambda/5$ . The wave propagates from left to right. The dots show the position of the same particle at a fixed distance with time increasing from the right to the left (modified from Shearer, Introduction into Seismology, 1999; with permission from Cambridge University Press).

Fig. 2.16 shows, for the event in Fig. 2.7, the 3-component broadband record of the Rayleigh-wave group and the related particle motion trajectories in three perpendicular planes. There exists indeed a strikingly clear retrograde elliptical motion in the vertical-east (Z-E) plane, as predicted by theory, which is in this case almost identical with the vertical plane of wave propagation (BAZ =  $273^\circ$ ). Also the vertical component amplitude is somewhat larger than the horizontal one, as theoretically expected too. In the horizontal plane (N-E), however, there is also some transverse energy present in this wave group. It is due to some SH energy present in the N-S component. Generally, one should be aware that the theoretically expected complete separation of LQ and LR waves in a homogeneous isotropic (horizontally layered) half-space is not absolutely true in the real Earth because of heterogeneity and anisotropy. This may cause the coupling of some Rayleigh-wave energy into Love waves and vice versa (see e.g., Malischewsky, 1987, and Meier et al., 1997), similar to S-wave splitting in the presence of anisotropy (see Fig. 2.8).



**Fig. 2.16** 3-component broadband records (top traces) and related plots of particle motion in the horizontal (N-E) plane and two vertical planes (Z-N and Z-E, respectively) of the surface-wave group of the same event as in Fig. 2.6 ( $D = 112$  km; backazimuth BAZ =  $273^\circ$ ).

It should be noted, however, that even a simple model “layer over half-space” may produce prograde particle motion for Rayleigh waves on the model surface (Malischewsky et al., 2006 and 2008). This phenomenon strongly depends on the S-wave velocity contrast, the Poisson ratio  $\sigma_1$  in the layer and the frequency. This phenomenon disappears for Poisson ratios less than 0.2 and S-wave contrasts  $v_{s1}/v_{s2}$  higher than 0.45 with  $v_{s1}$  the S-velocity in the layer and  $v_{s2}$  the S-velocity of the half-space. The frequency range is inversely proportional to the layer thickness. This has been demonstrated by way of example for the Dead Sea transform fault region in Israel. For the simplified Kiryat Shmona test site model (see Malischewsky et al., 2008)  $v_{s1}/v_{s2}$  is 0.25 and  $\sigma_1 = 0.46$  and one observes prograde motion between 11.3 Hz and 22.5 Hz for a layer thickness of 10 m and between 2.8 Hz and 5.6 Hz for a layer thickness of 40 m.

Because of the strong decay of surface wave amplitudes with depth, earthquakes deeper than the recorded wavelengths will produce significantly reduced or no surface waves. The amplitude ratio between body and surface waves in broadband records is thus a reliable quick discriminator between shallow and deep earthquakes. For shallow teleseismic earthquakes the surface wave group has generally by far the largest amplitudes in broadband and long-period records (see Fig. 2.25). This is because of their 2-D propagation along the surface of the Earth and energy decay  $\sim 1/r$  as compared to the 3-D propagation of body-waves and energy decay  $\sim 1/r^2$ . Also, the local maxima and minima in the group-velocity curves (Figs. 2.10 and 2.14) imply that surface wave energy in a wider period range around these extremes will travel with nearly the same velocity and arrive at a seismic station at about the same time, thus superimposing to large amplitudes. These amplitude maxima in the dispersive surface wave train are called *Airy phases*. For continental travel paths a pronounced Rayleigh wave Airy

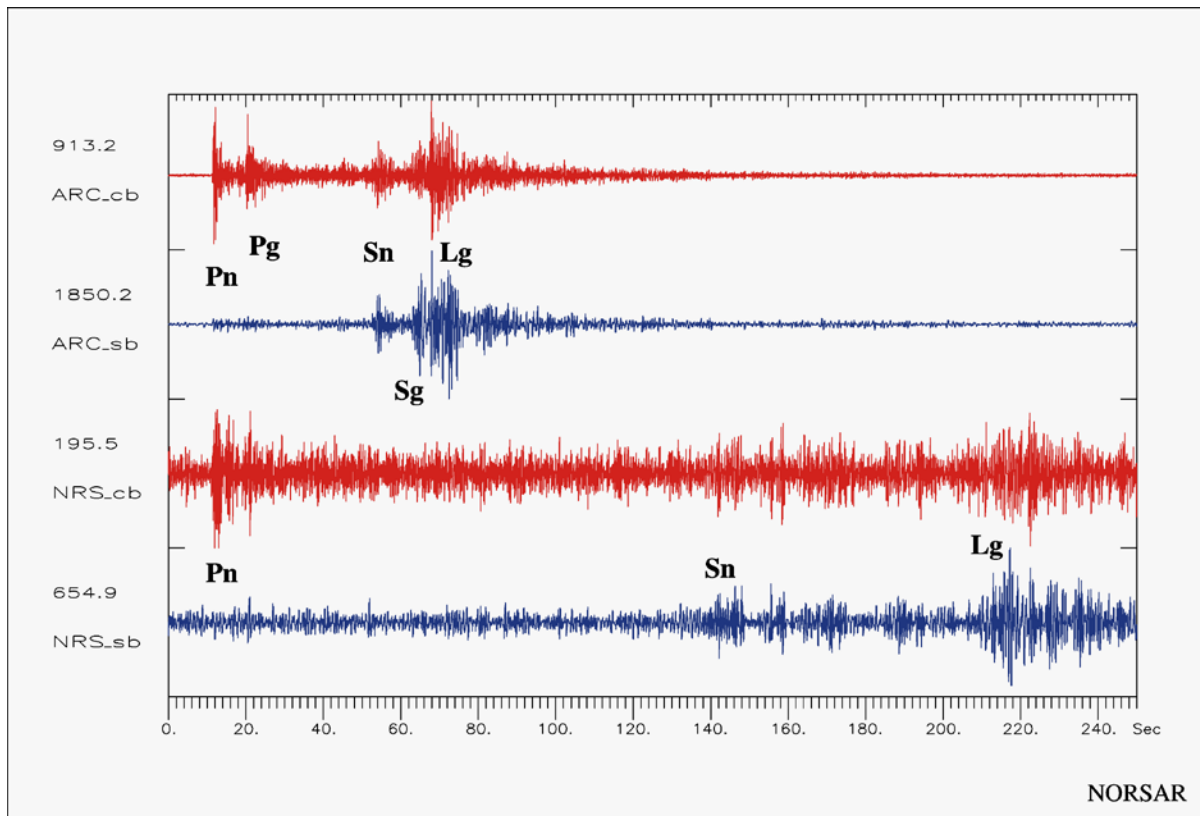
phase with periods around 15 s occurs which is rather stable. For oceanic travel paths the displacement amplitude maxima are due to the Airy phase periods associated with the group velocity maximum in Fig. 2.10 between about 25 s and 60 s. However, since in classical long-period records with WWSSN-LP response the maximum gain at periods of 15 s is already reduced, usually the period are measured around 20 s only. IASPEI standard surface-wave magnitudes are either measured on WWSSN-LP response records in the period range between 18 and 22 s (Ms\_20) or on velocity broadband records in a much wider period range between 3 s and 60 s (Ms\_BB; see IS 3.3 and IASPEI, 2011). Long-period mantle Rayleigh waves have another Airy phase around  $T \approx 200$  s, associated with a group velocity minimum (see Figs. 2.10 and 2.14). Such long-period mantle surface waves are used for measuring the (almost) not saturating mantle magnitude  $M_m$  (Okal, 1992a and b; Okal and Talandier, 1989), Chapter 3).

Higher mode surface waves have different depth dependence than fundamental modes and sample deeper structure than that probed by fundamental modes of the same period (see section 2.3.4 on mantle surface waves).

### 2.3.3 Crustal surface waves and guided waves

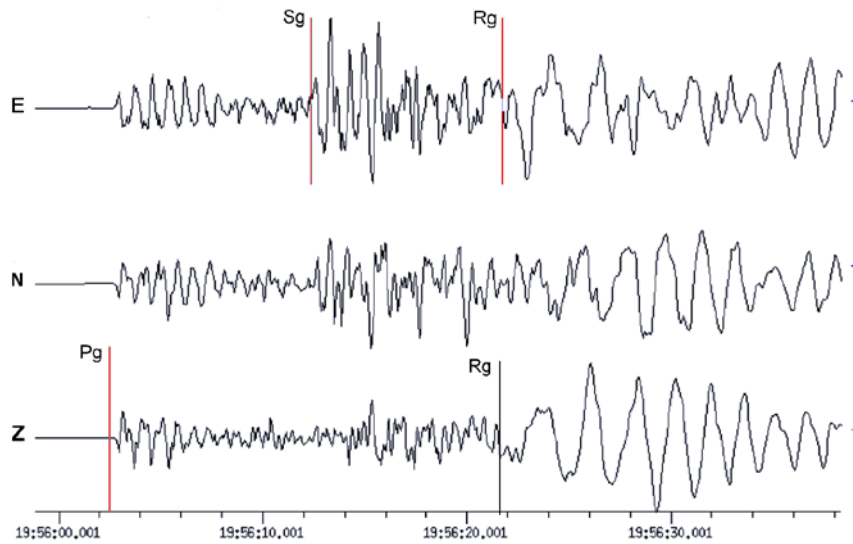
Because of the broad maximum of the group velocity of Love-wave continental overtones with periods between about 3 and 10 s and group velocities around 3.5 km/s they may appear in seismic records as an onset-like Lg-wave group with almost no dispersion, sometimes even in pairs (Lg1, Lg2) because of the nearby local minimum in the dispersion curve around 2 s (see Fig. 2.11). Since the group velocity of Lg-waves is higher than that of continental fundamental modes with  $T < 30$  s, they may appear in broadband records with an upper corner period around 20 s as clear forerunners to the surface wave group. The Lg-wave group, however, is not a pure continental Love wave but rather a complex *guided crustal* wave. It is caused by superposition of multiple S-wave reverberations between the surface and the Moho and SV to P and/or P to SV conversions as well as by scattering of these waves at lateral heterogeneities in the crust (for definition of phase nomenclature see IS 2.1 as well as Storchak et al. 2003 and 2011). Accordingly, Lg waves are also recorded on vertical components (see Fig. 2.17). Beyond epicenter distances of about  $3^\circ$  their amplitude maximum is usually well-separated from the earlier onset of Sg. Lg usually dominates seismic records of local and regional events and may propagate rather effectively along continental paths, in shield regions in particular, over a few thousand kilometers (see Fig. 2.17).

Because of the stability of Lg amplitude-distance relationships in continental areas this phase is well suited for reliable magnitude estimates of regional events (see 3.2.6.6) and mb\_Lg has become a new IASPEI standard (see IS 3.3 and IASPEI, 2011). The propagation of Lg may be barred by lateral changes in the velocity structure such as sedimentary basins, Moho topography, the transition between oceanic and continental crust and the boundaries between different tectonic units.



**Fig. 2.17** Records of Lg, together with other crustal phases, in records of a Kola peninsula mining blast ( $M_l = 2.4$ ) at the Norwegian array stations ARCES (distance  $D = 391$  km; upper two traces) and NORES ( $D = 1309$  km, bottom traces). cb and sb – P- and S-wave beams (see Chapter 9) of the vertical elements of the array, filtered with 2-8 Hz and 1-4 Hz, respectively (courtesy of J. Schweitzer).

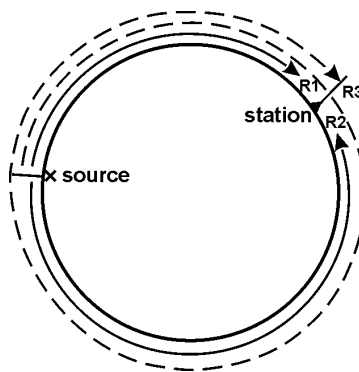
Near-surface seismic events such as industrial or underground nuclear explosions, rock-bursts etc. also generate short-period fundamental-mode Rayleigh waves, termed Rg. Rg waves show normal dispersion and have relatively large amplitudes on vertical components (see Fig. 2.18). They are not excited by seismic events deeper than about one wavelength and thus a good discriminator between often man-made seismic sources near the surface and most natural earthquakes with depths most likely between 5 and 25 km (crustal earthquakes) or even deeper (intermediate or deep earthquakes in subduction zones). Rg is more strongly attenuated than the guided wave Lg. Its range of propagation is limited to less than about 600 km. However, up to about 200 km distance Rg may dominate the recorded wave train from local near-surface seismic events.



**Fig. 2.18** Mining induced rock burst south of Saarbrücken, Germany, recorded at station WLF in Luxemburg ( $D = 80$  km,  $h = 1$  km,  $M_l = 3.7$ ). Note the strong dispersive Rg phase with periods between 1.2s and 2.8 s.

### 2.3.4 Mantle surface waves

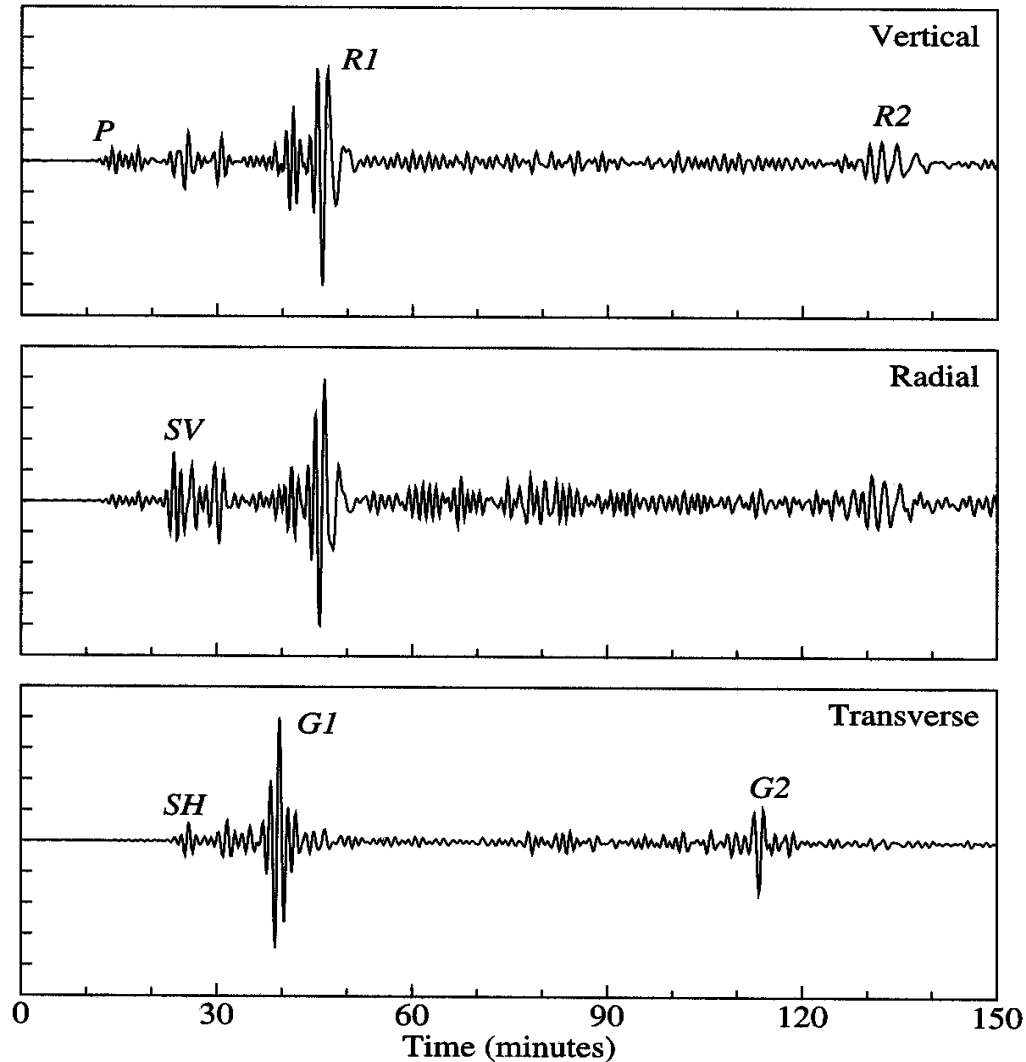
Love and Rayleigh waves travel along great circle paths around the globe. Surface waves from strong earthquakes may travel several times around the Earth. They are termed *global surface waves*. The first surface wave group arriving at a seismic station at the epicenter distance  $\Delta^\circ$  will have taken the shorter great circle while the later arrival has traveled the major arc path over  $360^\circ - \Delta^\circ$  (Fig. 2.19).



**Fig. 2.19** Great circle paths for the first three arrivals of global Rayleigh waves.

These arrival groups are called R1, R2, R3, R4 etc. for Rayleigh waves and G1, G2, G3, G4 etc. for Love waves, respectively. R3 (or G3) have traveled over  $360^\circ + \Delta^\circ$  and R4 over  $720^\circ - \Delta^\circ$  etc. Fig. 2.20 gives an example for long-period records of P, SV, SH, R1, R2, G1 and G2 in the vertical (Z) and the two horizontal components rotated into the radial (wave propagation) direction R and transverse direction T. As expected, P appears only on Z and R while S has both SV and SH energy. The Love wave groups G1 and G2 are strongest in T and arrive ahead of R1 in R2, which are visible only on the R and Z components. But Fig. 2.20 is also a

good example for inverse (negative) dispersion in the mantle Rayleigh-wave groups. Their periods range in the record from about 60 s to almost 200 s, with the longest periods arriving at the end of the R1 and R2 groups. This is just the period range of inverse dispersion according to Fig. 2.11 for both continental and oceanic mantle Rayleigh waves. This inverse dispersion is not seen in records of classical long-period WWSSN -LP seismographs with a peak magnification around 15 s because the corresponding periods are filtered out by the system response with a hundred times lower gain at 200s than at 15 s.



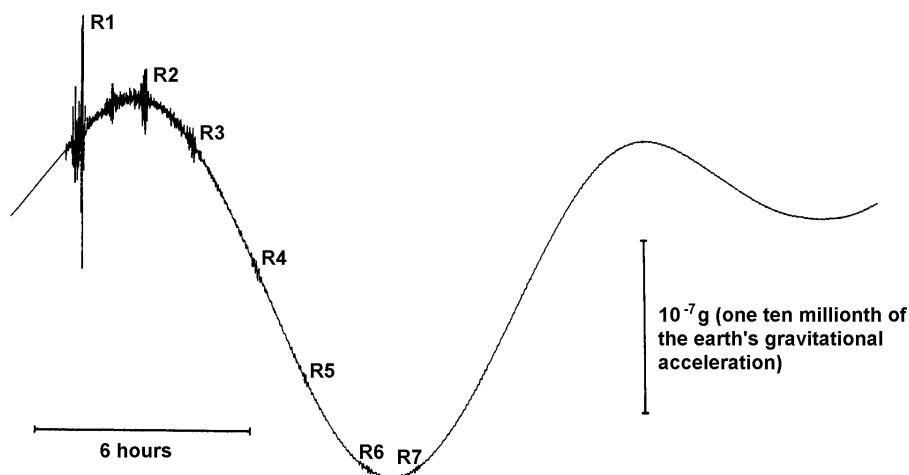
**Fig. 2.20** Records of the March 11, 1989 Tonga trench earthquake ( $h = 235$  km) in the Z, R and T components of the IRIS/IDA station NNA in Peru ( $D = 93.7^\circ$ ) (from Shearer, Introduction to Seismology, 1999; with permission from Cambridge University Press).

Further, one should note in Fig. 2.20 that these surface waves originate from an earthquake in the Tonga trench subduction zone at a depth of  $h = 230$  km. This seems to contradict the above statement, that no or only weak surface waves can be observed from deep earthquakes. However, there is no contradiction. As discussed above, the depth of penetration (and thus constructive interference) of surface waves increases with their wavelength. For the periods considered in Fig. 2.20  $\Lambda$  ranges between about 230 and 880 km, i.e., it is comparable or larger than the source depth. Therefore, we still can expect significant surface wave energy in



that period range for the largest amplitudes in Fig. 2.20. However, no periods below 50 s, as recorded in classical narrow-band long-period records, are recognizable in these long-period bandpass-filtered accelerometer records. Note, however, that in such records the displacement amplitudes are suppressed proportional to  $1/T^2$ , thus not representing correctly the spectral amplitude ratios. The longest period recognizable in the radial component record of Fig. 2.20 has a period of about 200 s, corresponding to the group velocity minimum of the fundamental mantle Rayleigh wave (Airy phase). In contrast, the pronounced wave group with much shorter periods ( $T$  around 70 s), with largest amplitudes arriving some 10 min. earlier, are mantle penetrating higher modes. According to Romanovicz (2011; courtesy of Yuancheng Gung) computations based on the PREM model (Dziewonski and Anderson, 1981) show that the fundamental Rayleigh-wave mode has at  $T = 150$  s its sensitivity peak around 200 km depth whereas the first two overtones have at a period of  $T = 100$  s still significant sensitivity down to the top of the lower mantle at depths around 800 km and 1200 km, respectively.

With modern very broadband (VBB) recording systems of high dynamic range (see Chapter 5) it is possible to record such long-period global mantle surface waves up to about R7, riding on oscillations of solid Earth's tides of even longer period (more than 12 hours). Fig. 2.21 shows a striking example. The successive groups of R reveal an exponential decay of amplitudes. This allows the determination of the intrinsic frequency-dependent attenuation in the crust and mantle (see 2.5.4.2).

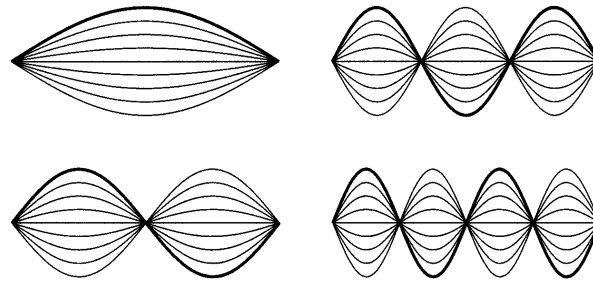


**Fig. 2.21** Example of a very broadband (VBB) record with high dynamic range by the STS1 seismograph operated by the Nagoya University, Japan. The seismic wave groups from a magnitude 8.2 earthquake in the Kermadec Islands (October 20, 1986) are superimposed to solid Earth's tides (modified from a pamphlet of the Japanese Global Seismology Subcommittee for the POSEIDON project).

## 2.4 Normal modes (P. Bormann)

Since the Earth is not an infinite half-space but a finite body, all wave motions must be confined too. Body waves are reflected back from the surface into the Earth, surface waves orbit along great circle paths. Thus, there will be a multitude of different seismic phases arriving at a given point on the surface. Depending on their timing and periods they will interfere with each other, either in a more destructive or more constructive manner. The latter will be the case at certain resonant frequencies only. They are termed the Earth's *normal*

*modes* and provide another alternative way of representing wave propagation. An analogy is the standing wave modes of a vibrating string fixed at both ends (Fig. 2.22). The lowest frequency is called the fundamental mode; the higher modes are the overtones. This can be treated as an eigenvalue problem: the resonant frequencies are called *eigenfrequencies*; the related displacements are termed the *eigenfunctions*.



**Fig. 2.22** The first four modes of vibration of a string between fixed endpoints (from Shearer, Introduction to Seismology, 1999; with permission from Cambridge University Press).

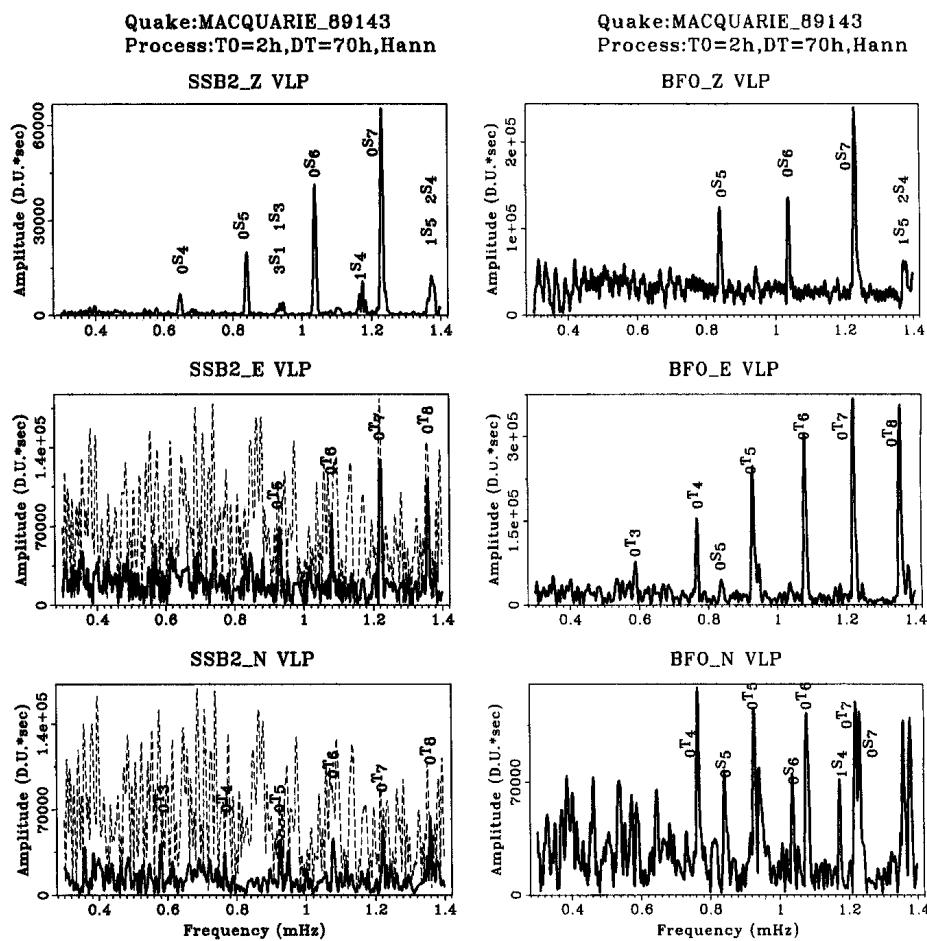
One should be aware of the following points about normal modes in observatory practice:

- Any wave motion within the Earth may be expressed as a sum of normal modes with different excitation factors;
- There exist, in analogy to P/SV (Rayleigh) and SH (Love) -wave motion, *spheroidal modes* and *toroidal modes*, respectively;
- Toroidal modes involve no radial motion and are only sensitive to the shear velocity;
- Spheroidal modes have both radial and horizontal motion. They are sensitive to both compressional and shear velocities;
- Long-period spheroidal modes are sensitive to gravity and thus provide information about the density structure of the Earth that may not be obtained in any other way;
- The ellipticity of the Earth, its rotation as well as its 3-D velocity variations will cause a splitting of the eigenfrequency spectral lines. Thus the investigation of normal mode splitting allows to constrain 3-D structures of the Earth;
- Normal modes do (besides PKPdf amplitudes) provide information about the shear-wave velocity of the *inner core*;
- The decay of mode amplitudes with time has provided important information about the attenuation properties of the Earth at very long periods;
- Normal modes provide a complete set of basis functions for the computation of synthetic seismograms for surface-wave and long-period body-wave seismology.

Therefore, the collection of high-quality broadband data that also allow retrieval of normal modes is an important function of high-tech seismological broadband observatories. This requires very stable installation conditions, for horizontal seismometers in particular, e.g., in boreholes (see 7.4.5) or deep non-producing underground mines in order to reduce near surface tilt noise caused by barometric pressure variations. The latter may also be filtered out by correlating parallel recordings of seismometers and micro-barometers (e.g., Warburton and Goodkind, 1977; Beauduin et al., 1996; see Fig. 2.23). For another very good low-noise free normal mode spectrum recorded from the disastrous N-Sumatra-Andaman earthquake on December 25, 2004 see Figure 4 in IS 5.1.

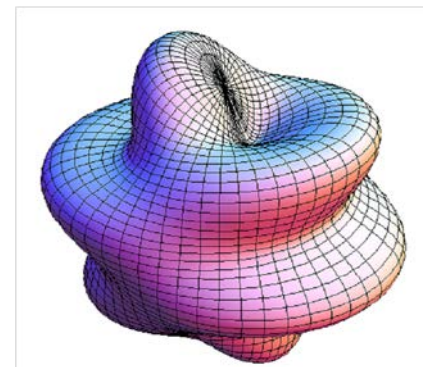
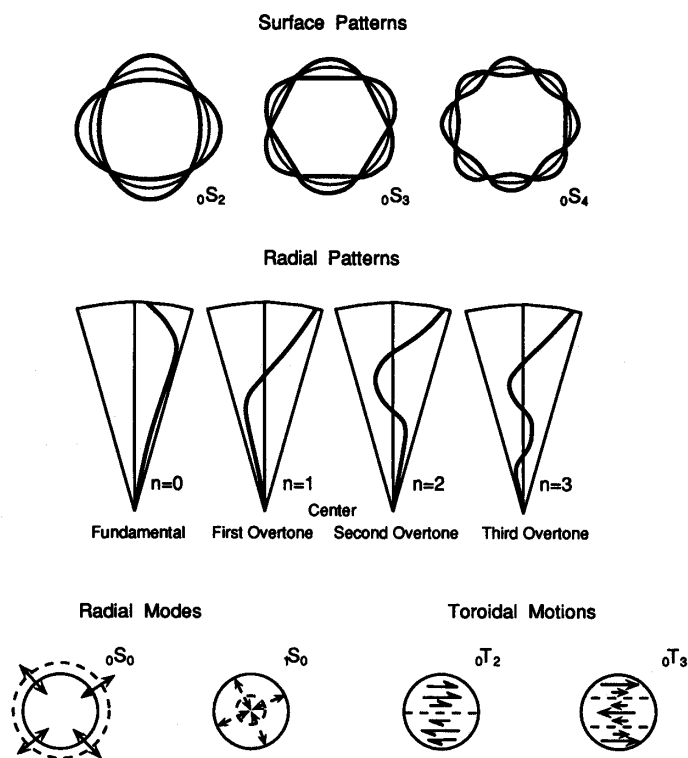
First observations of some normal modes were made in conjunction with the strongest earthquake of the 20<sup>th</sup> century (Chile, 1960). Since then, further progress in seismometry and data analysis have permitted the identification of over a thousand modes and on that basis, to significantly refine velocity, density and attenuation models of the Earth (see 2.7; PREM model).

In fact, normal mode analysis in the *spectral domain* (see Chapter 4.1) is the only practical way to examine seismic records at very long periods (> 500 s) and thus with wavelengths of 2000 and more kilometers. But normal mode studies themselves are beyond the scope of routine data analysis at seismological observatories and will not be considered in this Manual. (For further readings see Gilbert and Dziewonski, 1975; Aki and Richards, 1980 and 2002; Lapwood and Usami, 1981; Lay and Wallace, 1995; Dahlen and Tromp, 1998; Kennett, 2001; Longnonné and Clévéde, 2002).



**Fig. 2.23** Normal mode spectra excited by an  $M_s = 8.2$  earthquake in the Macquarie Island region and recorded with STS1 at the stations SSB2 in France and BFO in Germany. BFO is located in an old silver mine and has very low tilt noise. The latter is high at SSB2 (broken lines) but could be filtered out (solid lines) by correlation with micro-barometric recordings (reproduced from Beauduin et al., The Effects of the Atmospheric Pressure Changes on Seismic Signals ..., Bull. Seism. Soc. Am., Vol. 86, No. 6, Fig. 8, page 1768, 1996; © Seismological Society of America).

Fig. 2.24 shows the patterns of surface and radial motions related to some of the spheroidal and toroidal modes. Their general nomenclature is  ${}_nS_l$  and  ${}_nT_l$ .  $n$  is the number of zero crossings of amplitudes with depth while  $l$  is the number of zero (nodal) lines on the surface of the sphere. Accordingly, the fundamental spheroidal “breathing” mode of the Earth is  ${}_0S_0$  because it represents a simple expansion and contraction of the Earth. It has a period of about 20 min.  ${}_0S_2$  has the longest period ( $\approx 54$  min) and describes an oscillation between an ellipsoid of horizontal and vertical orientation, sometimes termed “rugby” mode. The toroidal mode  ${}_0T_2$  corresponds to a purely horizontal twisting motion between the northern and southern hemisphere and has a period of about 44 min. Overtones  ${}_iS$  and  ${}_iT$  with  $i = 1, 2, 3, \dots$  have one, two, three or more nodal surfaces at constant radii from the center of the Earth across which the sense of radial or twisting motions reverses. A 3-D presentation of the spherical harmonic of 6<sup>th</sup> degree and 1<sup>st</sup> order shows Fig. 2.24.



Spherical harmonic of 6<sup>th</sup> degree and 1<sup>st</sup> order, corresponding to  ${}_0S_6^1$ . The distinct corresponding multiplet spectral line with average frequency of 1.03755 mHz is clearly seen in the upper two spectral plots of Fig. 2.23 for stations SSB2 and BFO.

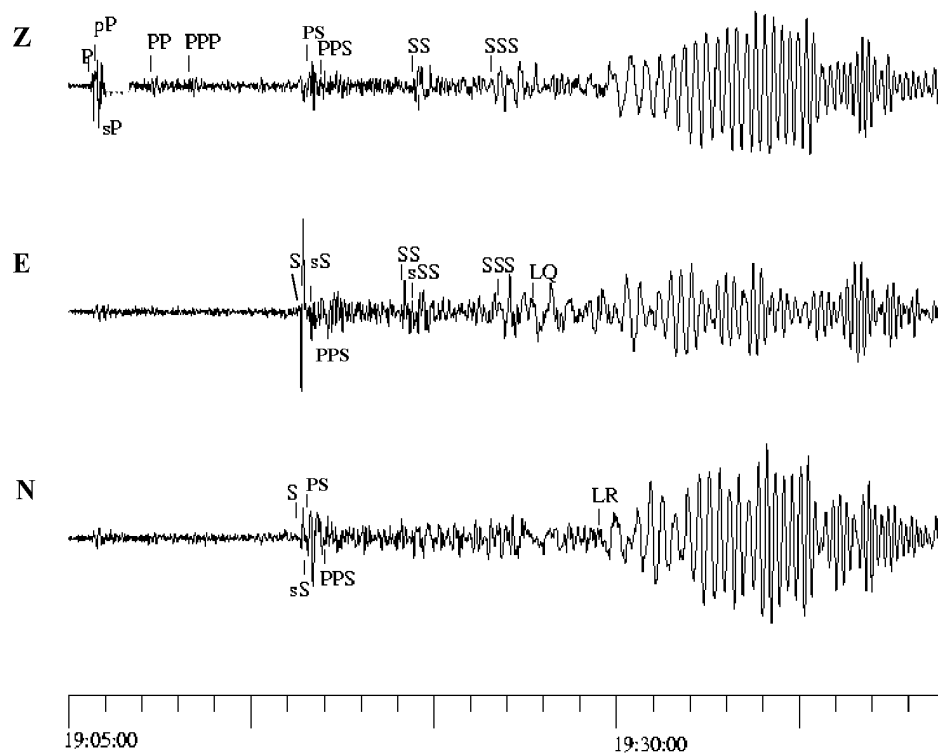
**Fig. 2.24 Left:** Top - Surface and radial patterns of motions of spheroidal modes; Bottom: Purely radial modes involve no nodal pattern on the surface but have nodal surfaces at depth. Toroidal modes involve purely horizontal twisting of the Earth. For explanation of mode nomenclature see text (after Bolt, 1982; from Lay and Wallace, 1995, Fig. 4.24, p. 160; with permission of Elsevier Science, USA). **Right:** Singlet figure  ${}_0S_6^1$  (courtesy of P. Malischewsky, 2012; modified version of Figure 3 in Malischewsky, 2011; calculated with Mathematica).

In summary, strong earthquakes can make the planet Earth ring like a bell. Seismologists may be compared with experienced bell-makers who are able to infer from the complex sound spectra of a bell not only its size and general shape but also the composition of the alloy of which it is made.

## 2.5 Seismic rays, travel times, amplitudes and phase shifts (P. Bormann)

### 2.5.1 Introduction

So far we have introduced seismic body and surface waves more from a physical and phenomenological point of view. We have learned why these different wave types travel with different velocities and consequently appear in the seismogram at different times. We have seen that body waves form short transient wavelets (see Figs. 2.4), in contrast to the prolonged and dispersed wave trains of surface waves (e.g., Figs. 2.12 and 2.13). Fig. 2.25 shows another seismic record of an earthquake  $73^\circ$  away. Besides the discussed primary body and surface waves (P, S, LQ, and LR), several additional arrivals are marked in the seismogram and their symbols are given. These energy pulses are mainly caused by reflection or mode conversion of primary P or S waves either at the free surface of the Earth or at velocity-density discontinuities inside the Earth.



**Fig. 2.25** Digital broadband record of the Seattle  $M_w = 6.8$  earthquake on 28 February 2001 at the station Rüdersdorf (RUE), Germany (epicenter distance  $D = 73^\circ$ ). Note the detailed interpretation of secondary phase onset (courtesy of S. Wendt).

A proper understanding of these arrivals is essential for a correct phase identification that in turn is of great importance for event location (see IS 11.1) and magnitude determination (see IS 3.3 and EX 3.1) but also for later determination of seismic velocities inside the Earth, especially also for S waves. We will introduce and use the concept of seismic rays to understand and illustrate the formation and propagation of these different wave arrivals.

Seismic ray theory is a very convenient and intuitive way to model the propagation of seismic energy and in particular of body waves. It is generally used to locate earthquakes and to determine focal mechanisms and velocity structure from body wave arrivals. Seismic ray theory is essentially analogous to optical ray theory, including phenomena like ray-bending, focusing and defocusing. We just summarize the very basics. For more detailed treatments see Červený (2001) and Chapman (2002).

Using ray theory, it is important to keep in mind that it is an approximation that does not include all aspects of wave propagation. Ray theory is based on the so-called *high-frequency approximation* which states that fractional changes in velocity gradient over one seismic wavelength are small compared to the velocity. In other words, we may use ray theory only if the dimensions of structures to be considered are larger than the seismic wavelengths used.

These conditions are valid for most parts of the Earth (see global model in Fig. 2.79) and for the wavelengths that are usually recorded and analyzed in seismological observatory practice. The problem of relatively sharp boundaries, as for example the crust-mantle interface (Moho - discontinuity), discontinuities in the upper mantle, and the core-mantle boundary (CMB) or the inner-core boundary (ICB) can be tackled by matching the boundary conditions between neighboring regions in which the ray solutions are valid. Yet, the derivation and practical application of more detailed 3D-models also for earthquake location gain nowadays strong momentum (see section 2.7.2).

## 2.5.2 Huygens's and Fermat's Principle and Snell's Law

In classical optics, *Huygens's Principle* governs the geometry of a wave surface. It states that every point on a propagating wavefront can be considered the source of a small secondary wavefront that travels outward at the wave velocity in the medium at that point. The position of the wavefront at a later time is given by the tangent surface of the expanding secondary wavefronts. Since portions of the primary wave front, which are located in relatively high-velocity material, produce secondary wavefronts that travel faster than those produced by points in relatively low-velocity material, this results in temporal changes of the shape of the wavefront when propagating in an inhomogeneous medium. Since rays are defined as the normals to the wavefront, they will change accordingly. Rays are a convenient means for tracking an expanding wavefront. Fig. 2.26 depicts the change of direction of a plane wavefront and associated rays when traveling through a discontinuity which separates two homogeneous media with different but constant wave propagation velocity.

*Fermat's Principle* governs the geometry of the ray path. It states that the energy (or ray) will follow a *minimum time path*, i.e., it takes that path  $d$  between two points, which takes an extreme travel-time  $t$  (i.e., the shortest or the longest possible travel time, with  $\partial t / \partial d = 0$ ). Such a path is called *stationary*. In case of a stationary time path there exist three possibilities, depending on the value (sign) of the higher derivatives of  $\partial t / \partial d$ :

|     |                                   |                                                                                                             |
|-----|-----------------------------------|-------------------------------------------------------------------------------------------------------------|
| for | $\partial^2 t / \partial d^2 > 0$ | the ray follows a true <i>minimum time path</i> ,                                                           |
| for | $\partial^2 t / \partial d^2 < 0$ | the ray follows a <i>maximum time path</i> and                                                              |
| for | $\partial^2 t / \partial d^2 = 0$ | i.e., in case of an inflection point of the travel-time curve, the ray follows a <i>minimax time path</i> . |

Different kinds of seismic waves follow different time paths, e.g., the reflected waves pP (see Fig. 2.62) a true minimum path, the PP or the SKKS reflection (Fig. 2.52) a minimax path and the reflected wave P'P' (PKPPKP) (Fig. 2.53) a true maximum path. Note that the character of the stationary path influences the character (phase shift) of the reflected waveform. Whenever a seismic ray travels in some parts of its ray path as a maximum time ray, it touches a *caustic*. This caustic can be a focusing point (see 2.5.3.3 or 2.5.3.4) or a surface along which seismic rays superimpose each other (see 2.5.4.3). In any case, prominent phase distortion can be observed and may have to be taken into account when analyzing seismograms, depending on the task.

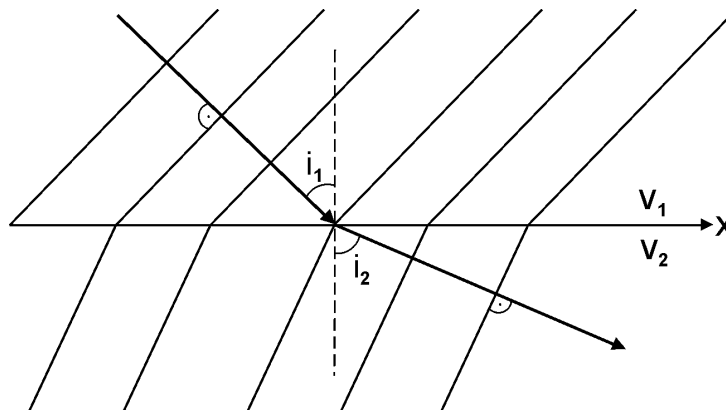
### 2.5.2.1 Snell's Law for a flat Earth

From *Fermat's Principle* follows, with some simple geometry and mathematics, *Snell's Law* of wave refraction (e.g., Aki and Richards, 1980 and 2002; Lay and Wallace, 1995; Shearer, 1999; Červený, 2001; Kennett, 2001):

$$\sin i/v = s \sin i = s_x = 1/v_{\text{app}} \equiv p = \text{constant} \quad (2.12)$$

where  $i$  is the angle of incidence, measured between the ray and the vertical (see Fig. 2.26),  $v$  is the velocity (speed) of wave propagation in the medium,  $s = 1/v$  is called slowness, and  $p$  is the so-called ray parameter,  $v/\sin i = v_{\text{app}}$  is the apparent horizontal wave propagation velocity in  $x$ -direction with  $v_{\text{app}} = \infty$  for  $i = 0$  (vertical incidence of the ray) and  $s_x = 1/v_{\text{app}}$  is the horizontal component of the slowness vector  $s$ . Note, however, that  $p$  is constant for laterally homogeneous media only. In Fig. 2.26 the refraction of a seismic wavefront and of a related seismic ray across the interface of two half spaces with different but constant seismic velocities  $v_1$  and  $v_2$  is sketched. Such an instantaneous velocity jump is called first-order discontinuity. Because the ray parameter must remain constant across the interface, the ray angle has to change:

$$\sin i_1/v_1 = \sin i_2/v_2 = s_1 \sin i_1 = s_2 \sin i_2. \quad (2.13)$$



**Fig. 2.26** A plane wavefront with the associated ray crossing a medium boundary with  $v_2 > v_1$ . Because of the velocity increase in medium 2 the ray is refracted away from the vertical, i.e.,  $i_2 > i_1$ .



### 2.5.2.2 Snell's Law for the spherical Earth

Above, a flat-layered case was considered. Yet the Earth is a sphere and curvature has to be taken into account at distances greater than about  $12^\circ$ . In this case the ray parameter has to be modified. In Fig. 2.27 a ray is sketched in a sphere composed of two concentric shells 1 and 2 of different but constant velocity  $v_1$  and  $v_2$  or slowness  $s_1 = 1/v_1$  and  $s_2 = 1/v_2$ , respectively. At the first interface between medium 1 and 2, Snell's Law must be satisfied locally, i.e.,:

$$s_1 \sin i_1(r_1) = s_2 \sin i_2(r_2) \quad (2.14)$$

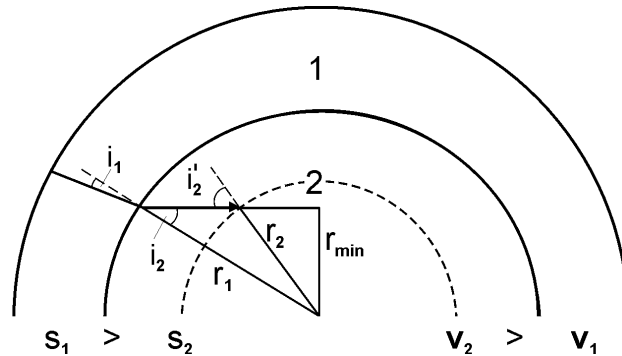
for  $r_1 = r_2$ . Inside shell 2, however, despite  $v_2 = \text{const.}$ , the incidence angle changes as the ray progresses, namely,  $i_1(r_1) \neq i'_2(r'_2)$ . If we project the ray in medium 2 further to its turning point where  $r = r_{\min}$  we see from the set of right triangles that the following relationship holds:

$$s_1 r_1 \sin i_1 = s_2 r'_2 \sin i'_2.$$

This is true along the entire ray path and we can generalize

$$s r \sin i = r \sin i/v \equiv p, \quad (2.15)$$

which is the modified Snell's Law for a spherical Earth.



**Fig. 2.27** Ray geometry for an Earth model consisting of two spherical shells of constant but different velocity  $v_1$  and  $v_2$ .

In fact, it has been Benndorf (1905 and 1906) who first published that in the case of a spherically symmetric Earth model apparent velocity and thus the seismic ray parameter  $p$  are constant along their whole ray path (*Benndorf's Law*).

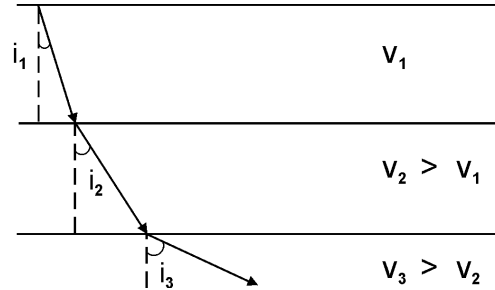
## 2.5.3 Rays and travel times in laterally homogeneous (1-D) media

### 2.5.3.1 Velocity gradient

It is true for most parts of the Earth that the seismic velocity increases with depth due to compaction of the material. Consider a ray travelling downwards through a stack of layers with constant velocities  $v_i = 1/s_i$  each, however, increasing layer velocities with depth (Fig. 2.28). Applying Snell's law

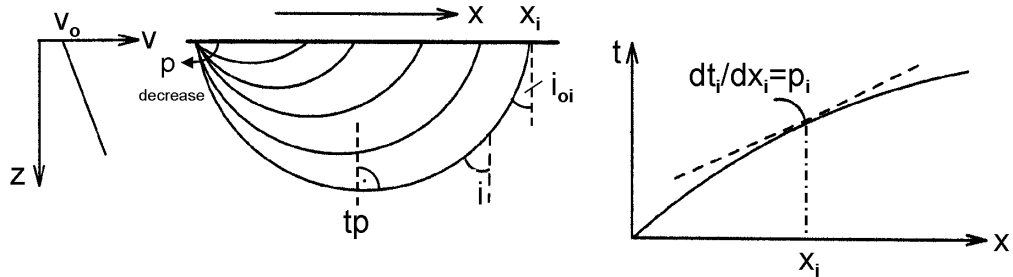
$$p = s_1 \sin i_1 = s_2 \sin i_2 = s_3 \sin i_3 \dots \quad (2.16)$$

we can derive the incidence angle  $i$ , that is continuously increasing with depth, and finally approaching  $90^\circ$ . At  $i = 90^\circ$  the ray is at its *turning point*  $tp$ .



**Fig. 2.28** Ray through a multi-layered model with constant velocity within the layers but increasing velocity with depth of the layers. The ray angle  $i$  increases accordingly with depth.

This can be generalized by modeling a velocity gradient with depth as a stack of many thin layers with constant velocity. Rays and travel times for this case are sketched in Fig. 2.29. The plot of arrival times  $t$  versus distance  $x$  is generally called the *travel-time curve*. The tangent  $dt_i/dx_i$  on the travel-time curve at any distance  $x_i$  corresponds to the inverse of the horizontal wave propagation velocity  $1/v_{appi}$  and thus to the ray parameter  $p_i$  of that ray which comes back to the surface at  $x_i$ . Because of  $\sin i = \sin 90^\circ = 1$  at the turning point of the ray, we can determine the velocity  $v_{tp}$  at the turning point of the ray either from the gradient of the travel-time curve at  $x_i$  via  $p_i = dt_i/dx_i = 1/v_{tp}$  or by knowing the sub-surface velocity  $v_{oi}$  at station  $x_i$  and measuring the incidence angle  $i_{oi}$  at that station ( $v_{tp} = v_{oi} / \sin i_{oi}$ ).



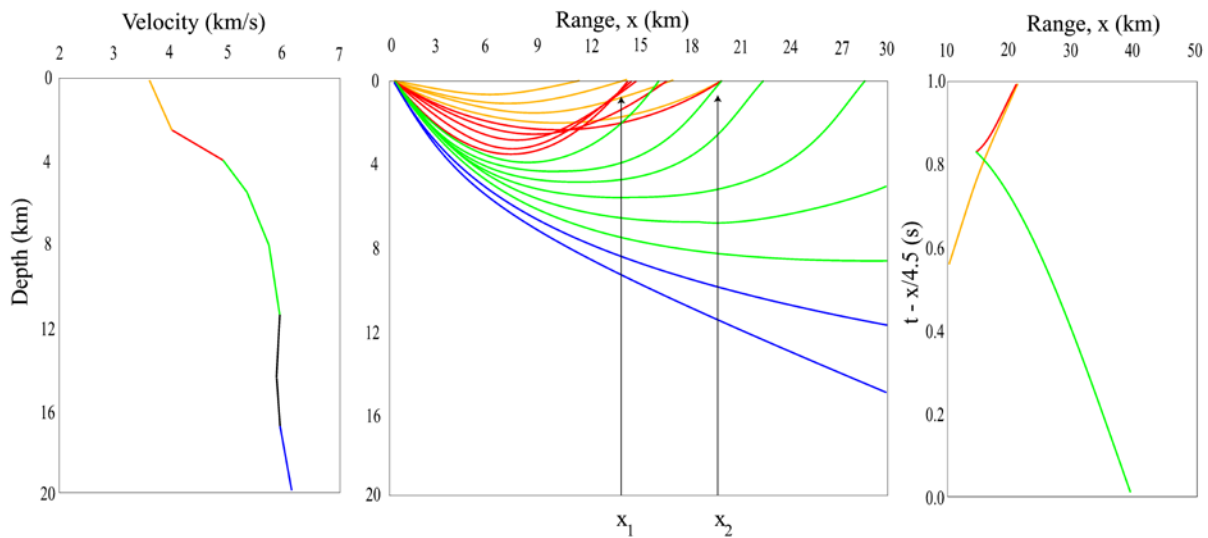
**Fig. 2.29** Ray paths (middle) and travel-time curve (right) for a model with velocity  $v$  gradually increasing with depth  $z$  (left). The incidence angle  $i$  increases continuously until it reaches  $90^\circ$  at the turning point  $tp$ , then the rays turn up again to reach the surface at  $x_i$ . On the travel-time curve each point comes from a different ray with a different slowness and ray parameter  $p$ . The gradient of the tangent on the travel time curve at  $x_i$  is the ray parameter  $p_i = dt_i/dx_i$ . In the considered case of modest velocity increase with depth the distance  $x$  increases with decreasing  $p$ . The related travel-time curve is termed *prograde*.

### 2.5.3.2 Effect of a sharp velocity increase

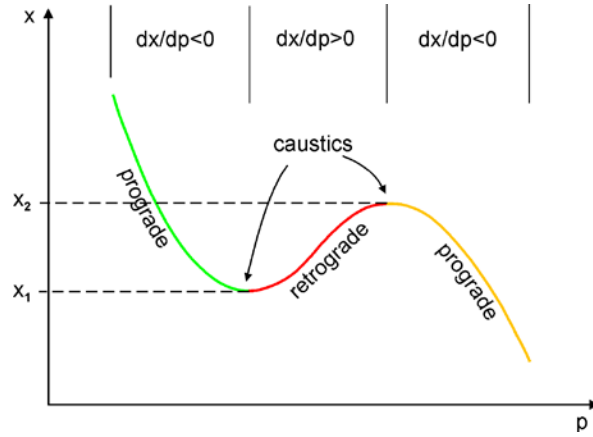
Next we consider the effect of a sharp velocity increase, which may be an increase in gradient (second-order discontinuity) or an instantaneous velocity jump (first-order discontinuity). Fig.

2.30 shows on the left side a hypothetical velocity-depth model in the upper crust of the Earth together with the related seismic rays and on the right the corresponding travel-time curves in the reduced-time presentation  $t_{\text{red}} = t - (x/v_{\text{red}})$ . Travel-time usually increases with distance. Consequently, presenting absolute travel-time curves for large epicenter distances would require very long time-scale axes. Moreover, small changes in  $dt/dx$  are then not so well recognizable. Therefore, in order to reduce the time scale and to increase the resolution of changes in slowness, travel-time curves are often represented as *reduced travel-time curves*, in which  $t_{\text{red}} = t - x/v_{\text{red}}$  is plotted (for some constant  $v_{\text{red}}$ ) as a function of  $x$ . The reduction velocity  $v_{\text{red}}$  is usually chosen so as to be close to the mean velocity in the considered depth range or of the considered seismic phase. Its reduced travel-time is then constant and positive or negative slowness deviations are clearly recognizable.

In the ray diagram of Fig. 230 one recognizes that at certain distances  $x$ , rays with different incidence angles may emerge. Modest velocity gradients in the upper and lower part of the velocity profile result in rays which return to the surface with increasing distance  $x$  for decreasing ray parameter  $p$ . This produces prograde travel-time branches (yellow and green branches in the  $t_{\text{red}}-x$  plot). In contrast, a strong velocity gradient leads to decreasing  $x$  with decreasing  $p$  and thus to a receding (retrograde) travel-time branch (red). Thus, a strong gradient zone between two weak gradient zones results in a *triplication* of the travel-time curve. The endpoints of the triplication are called *caustics*. At the caustics (positions  $x_1$  and  $x_2$ ) rays, which have left the source under different take-off angles, arrive at the surface at the same time. This causes a focusing of energy, large amplitudes and a waveform distortion (see 2.5.4.3). Fig. 2.31 shows qualitatively, with the same color coding as in Fig. 2.30, the changes in the ray parameter  $p$  with distance  $x$  for the prograde and retrograde travel-time branch(es) of a triplication.

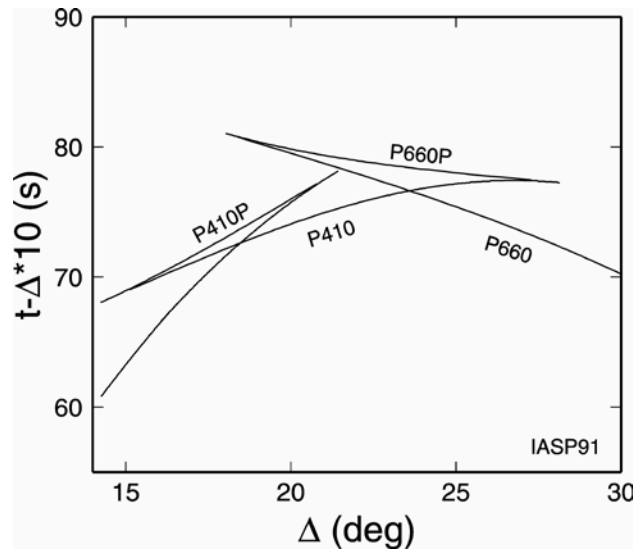


**Fig. 2.30 Left:** Velocity-depth profile in a model of the upper crust with a strong velocity gradient between about 2.5 and 6 km depth and related seismic rays from a surface source. **Right:** ray diagram and  $t_{\text{red}}-x$  relation for the given model;  $v_{\text{red}} = 4.5$  km/s. Note the differently colored segments of the velocity-depth distribution and of the travel-time branches that relate to the seismic rays given in the same color. Yellow and green: prograde travel-time curves, red: retrograde travel-time curve. Note the two lowermost blue rays that have already been affected by a low-velocity zone below 10 km depth (courtesy of P. Richards, 2002.)



**Fig. 2.31** Distance  $x$  as a function of ray parameter  $p$  for *triplications*. Note that the colors in this diagram correspond to the colors of the related rays and velocity segments in Fig. 2.30.

The gradient of the retrograde travel-time branch and the position  $x_1$  and  $x_2$  of the caustics are controlled by the thickness and the velocity-gradient in this strong-gradient zone. Similar triplications develop in the presence of first-order discontinuities with positive velocity jump. In this case the retrograde branch relates to the post critical reflections from such a discontinuity (see Fig. 2.35 in 2.5.3.6). The identification and quantification of such first- and second-order discontinuities is of greatest importance for the understanding of related changes in physical and/or compositional properties in the Earth. This necessitates, however, that not only first arrivals of seismic waves but also later, secondary arrivals are identified and their amplitudes measured. Since the latter may follow rather closely to the former, their proper identification and onset-time measurement may be difficult in very narrow-band filtered recordings because of their strong signal distortion (see figures in Chapter 4, section 4.2).

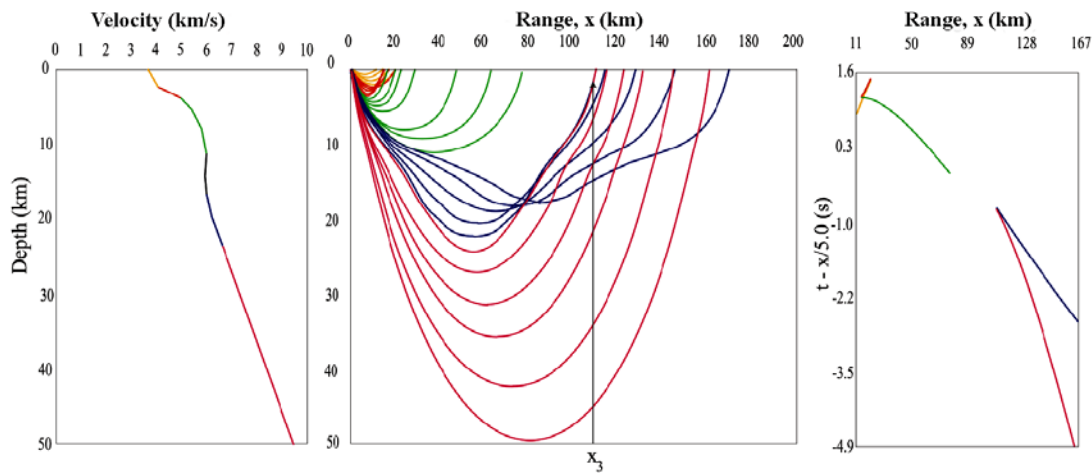


**Fig. 2.32** Triplications of the P-wave travel-time curve (here in reduced presentation) due to the 410 km and 660 km upper mantle/transition zone discontinuities, calculated according to the IASP91 velocity model (Kennett and Engdahl, 1991) (section 2.7, Fig. 2.76). The P waves diving directly below the 410 km (660 km) are called P410 (P660); the phases P410P and P660P are the over- reflections from the outer side of these discontinuities, respectively.

Two of the most pronounced velocity and density increases occur at about 410 and 660 km below the surface (section 2.7, Figs. 2.51 and 2.53). They mark the lower boundary of the upper mantle and of the transition zone from the upper mantle to the lower mantle, respectively. Both are caused by phase transitions of the mantle material into states of higher density at critical pressure-temperature (P-T) conditions. These two pronounced discontinuities result in triplications of the P-wave travel-time curves in the distance range between about  $14^\circ$  and  $28^\circ$  (Fig. 2.31) associated with a strong increase of P-wave amplitudes around  $20^\circ$  (so-called  $20^\circ$  discontinuity; see Fig. 3.34 in Chapter 3).

### 2.5.3.3 Effect of a low-velocity zone

Velocity generally increases with depth due to compaction, however, lithologic changes or the presence of water or melts may result in low-velocity zones (LVZ). Fig. 2.33 shows the effects of an LVZ on seismic rays and the travel-time curve. The latter becomes discontinuous, forming a shadow zone within which no rays emerge back to the surface. Beyond the shadow zone the travel-time curve continues with a time off-set (delay) from a caustic with two branches: one retrograde branch (blue) beginning with the same apparent horizontal velocity as just before the beginning of the shadow zone and another prograde branch with higher apparent velocity (smaller  $dt/dx$ ). Fig. 2.33 is in fact a continuation of the model shown in Fig. 2.30 towards greater depths. One recognizes a low-velocity zone between 12 and 18 km depth. The related ray diagram clearly shows how the rays that are affected by the LVZ jump from an arrival at distance 79 km to 170 km, and then go back to a caustic at 110 km before moving forward again. The related prograde travel-time branches and rays have been color-coded with **green**, **blue** and **violet**. The corresponding  $t_{\text{red}}-x$  plot on the right side of Fig. 2.33 shows nicely the travel-time offset and caustic beyond the shadow zone with two branches: a) retrograde (blue) and b) prograde (violet).



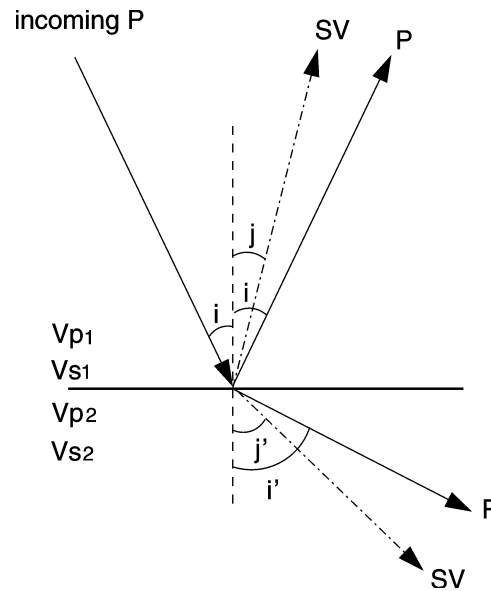
**Fig. 2.33** **Left:** Velocity-depth profile and seismic rays in the crust with a low-velocity zone between  $12 \text{ km} < h < 18 \text{ km}$  depth. The black segment in the velocity-depth curve produces the shadow zone. **Right:** ray diagram and  $t_{\text{red}}-x$  relation for the considered model. The reduction velocity is  $v_{\text{red}} = 5.0 \text{ km/s}$ . Note the additional colored travel-time branches which relate to the seismic rays given in the same color. Green and violet: prograde travel-time curves, blue and red: retrograde travel-time curves. There is a caustic at distance  $x_3$ . Therefore, the end of the shadow has strong amplitudes (courtesy of P. Richards).

An outstanding example for an LVZ, which shows these features very clearly, is the outer core. At the core-mantle boundary the P-wave velocity drops from about 13.7 km/s in the lowermost mantle to about 8 km/s in the liquid outer core. This causes a shadow zone for short-period direct P waves between around 100° and 144°, however slightly “illuminated” by reflected arrivals from the inner-core boundary (PKiKP) and by rays that have been refracted backward to shorter distances (retrograde travel-time branch) due to the strong velocity increase in the inner core (phase PKPdf = PKIKP) (Chapter 11, Fig. 11.73). The travel-time branch PKPab corresponds qualitatively to the blue branch and the branch PKPdf beyond the caustic to the violet branch in Fig. 2.33 (compare with Fig. 2.59). There may exist, however, also low-velocity zones (LVZs) in the crust and in the upper mantle (asthenosphere; see PREM model in Fig. 2.77). LVZs often more pronounced in S-wave velocity than in P-wave velocity because material weakening due to (partial) melting reduces more strongly the shear module  $\mu$  than the bulk module  $\kappa$  [see Eqs. (2.9) and (2.10)].

#### 2.5.3.4 Refraction, reflection, and conversion of waves at a boundary and formation of inhomogeneous waves

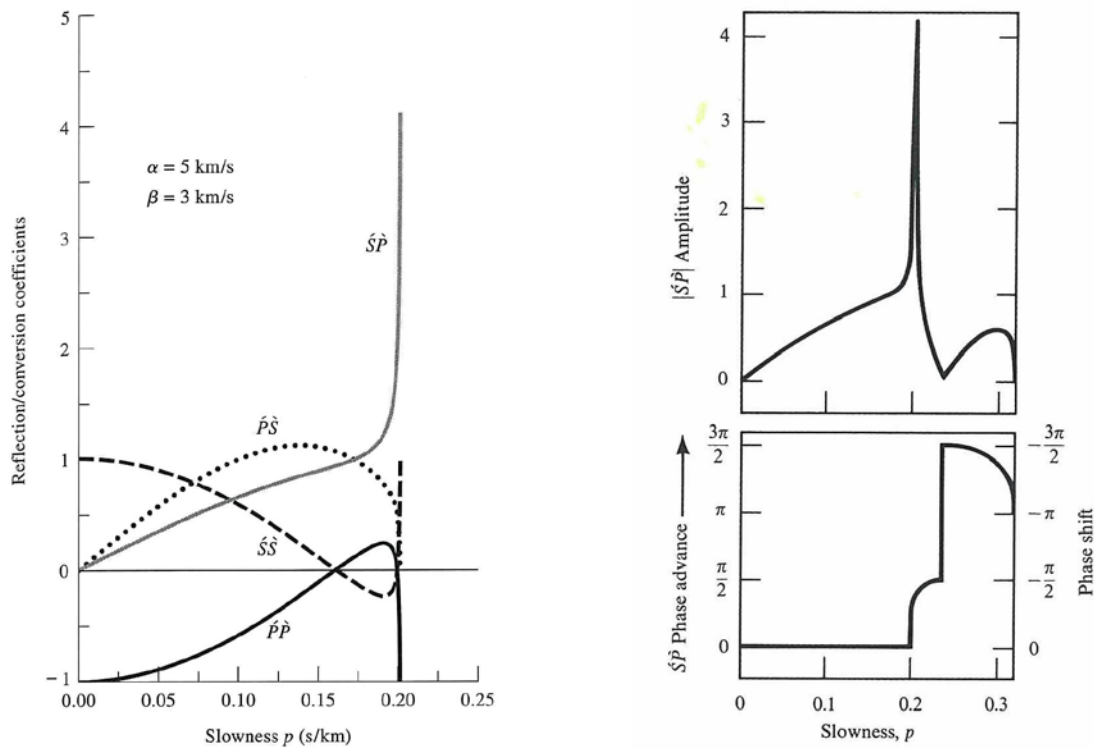
So far we have only considered transmission of seismic waves at a boundary. However, generally not all energy is transmitted; parts are reflected or converted. If a P wave hits a boundary between different seismic velocities, four different waves may be generated: a transmitted P wave; a converted transmitted S wave purely polarized in the vertical plane of propagation (SV-wave); a reflected P wave; and a reflected converted SV wave (Fig. 2.34). The geometry of these waves is also governed by Snell's Law:

$$\sin i/v_{p1} = \sin j/v_{s1} = \sin i'/v_{p2} = \sin j'/v_{s2}. \quad (2.17)$$



**Fig. 2.34** An incident P wave at a solid-solid boundary (shown is the case  $v_1 < v_2$ ) generates a reflected and a transmitted P wave and a reflected and transmitted SV wave. Snell's Law governs the angular relationship between the rays of the resultant waves.

In the case of an SH wave hitting the boundary, which is purely polarized in the horizontal plane, there is only a transmitted and a reflected SH wave, but no conversion into P or SV possible. If a single incident wave is split into multiple scattered waves, energy must be partitioned between these waves. Coefficients governing the partitioning between transmitted, reflected, and converted energy will generally depend on the incidence angle of the incoming wave and the *impedance contrast* at the boundary. Impedance is the product of wave velocity and density of the medium. Derivation of the expressions for reflection, transmission, and conversion coefficients is beyond the scope of this book. However, as a telling example, Fig. 2.35 shows for a free surface the four possible P-SV reflection/conversion coefficients (displacement amplitude ratios) against horizontal slowness  $p = \sin i/v_x$  as well as the coefficients for SP and SS, together with the related phase shift, for an incident SV-wave over an extended range of  $p$ . Note the rather large coefficient of 4.1 for SP when  $p = 0.2$  and thus  $i$  for the converted P wave  $90^\circ$ .



**Fig. 2.35 Left:** The four possible reflection/conversion coefficients for a P wave incident on a free surface and either reflected back into the half-space as P or being converted into SV as a function of slowness  $p$ , and these same for an incident SV being either reflected as SV or converted into P. **Right:** The same again for incident SV but for an extended range of slowness  $p$ . In both cases the velocities of  $v_p$  and  $v_s$  in the half-space are 5 km/s and 3 km/s, respectively (courtesy of P. Richards, 2012).

In the left-hand panel of Fig. 2.35 the range for  $p$  is restricted to  $0 \leq p \leq 1/v_p$ . In this case the incidence angle  $i$  is always real and all related waves are homogeneous ones. If, however,  $p > 1/v_p$  then P is an inhomogeneous wave with an imaginary depth component of slowness vector  $\mathbf{s}$ , that is, the horizontal phase velocity,  $p^{-1}$ , is less than the propagation speed of P-waves. The amplitudes of such an inhomogeneous wave decay exponentially with depth in the half-space. If  $p$  is even greater, so that  $p > 1/v_p > 1/v_s$  holds then the associated S wave is an

inhomogeneous wave as well. In section 2.3.1 we had already mentioned, that Rayleigh waves originate from a coupled pair of inhomogeneous P and SV waves that propagate along the surface of a half-space. According to Aki and Richards (2002), all the reflection/transmission coefficients associated with a particular interface will become complex if at least one of the waves set up at the interface by an incident wave is an inhomogeneous wave. An alternative solution to the wave equation, in the case of inhomogeneous waves, is the solution where the amplitudes grow with depth. If there is another interface, then both solutions are present in the layer between two interfaces, if the horizontal slowness is sufficiently great. Note, that in the right-hand panel of Fig. 2.35 SP is a real wave in the slowness range  $0 \leq p \leq 1/v_p = 0.2$  but for larger  $p$  values an inhomogeneous P wave with a strong phase advance is formed.

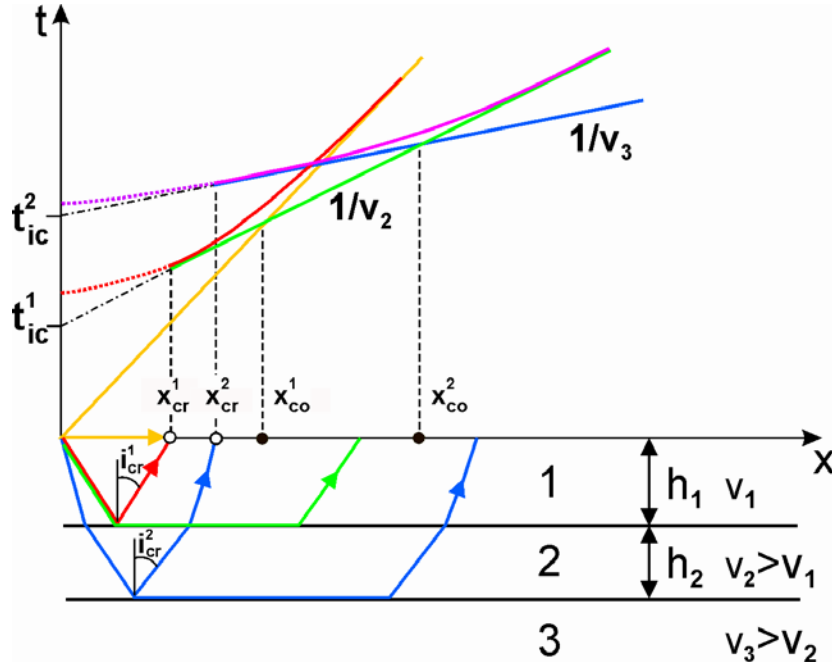
For more details we relate for a complete treatment to the classic textbook of Aki and Richards (1980 and 2002) and for a condensed overview to Müller (1985) or Shearer (1999). The same applies to the following considerations on seismic energy, amplitudes and phases.

### 2.5.3.5 Seismic rays and travel times in homogeneous models with horizontal and tilted layers

Below we consider a horizontal two-layer model above a half-space. Within the layers and in the half space the wave velocities are constant with  $v_1 < v_2 < v_3$ . The discontinuities between them are of first order, i.e., with instantaneous velocity jumps (see Fig. 2.36). For an incidence angle  $i_1 = i_{cr}^1$ , with  $\sin i_{cr}^1 = v_1/v_2$  and  $v_2 > v_1$ , no wave energy can penetrate into the layer 2, because  $\sin i_2 = 1$  and thus  $i_2 = 90^\circ$ . The angle  $i_{cr}$  is called the *critical angle* because for  $i > i_{cr}$  all energy incident at a first-order discontinuity is totally reflected back into the overlaying layer. However, part of it may be converted. The point in the travel-time curve at which a critically reflected ray (reflection coefficient 1) comes back to the surface is termed the *critical point*  $x_{cr}$ . The travel-time curve has a caustic there. Reflected rays arriving with  $i < i_{cr}$  are termed *pre-critical* (or steep angle) reflections (with reflection coefficients  $< 1$ ), those with  $i > i_{cr}$  as *post-critical*, *super-critical* or *wide-angle reflections*. However, in this case the reflection coefficient becomes a complex number which results in the above discussed phase distortion of over-critical reflections. Note that the travel-time hyperbola of the reflected waves from the bottom of the first layer (red curve) merges asymptotically at larger distances with the travel-time curve of the direct wave in this layer (yellow curve).

Seismic rays incident with  $i_n = i_{cr}^n$  on the lower boundary of layer  $n$  are refracted with  $i_{n+1} = 90^\circ$  into the boundary between the two layers  $n$  and  $n+1$ . They form so-called seismic *head waves* (green and blue rays and travel-time curves, respectively, in Fig. 2.36). Head waves are inhomogeneous boundary waves that travel along the discontinuity with the velocity of layer  $n+1$  and radiate upward wave energy under the angle  $i_{cr}^n$ . The amplitudes of pure head waves decay theoretically with  $1/r^2$ . The full description of this kind of wave, however, is not possible in terms of ray theory but requires a wave-theoretical treatment. In the real Earth, with non-ideal first-order layer boundaries, true *head waves* will hardly exist but rather so-called *diving waves* which slightly penetrate - through the high-gradient zone between the two media - into the underlying high-velocity medium. There they travel sub-parallel to the discontinuity and are refracted back towards the surface under an angle  $\approx i_{cr}$ . In terms of travel time there is practically no difference between a diving wave and a pure head-wave along a first-order velocity discontinuity; diving waves, however, have usually larger amplitudes and can still be observed at much larger distance from the source.





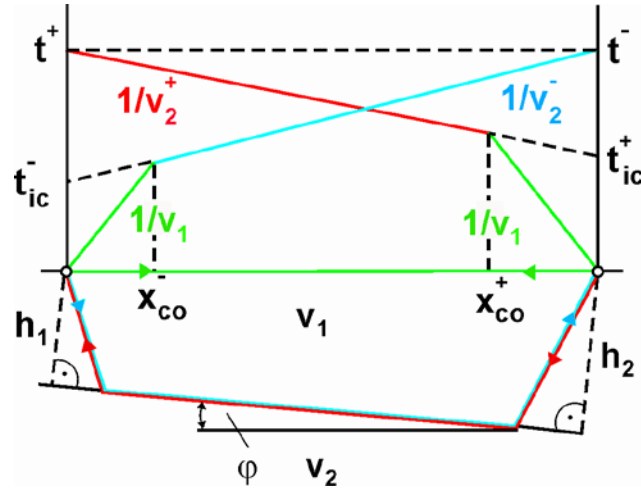
**Fig. 2.36** Schematic local travel-time curves (time  $t$  over distance  $x$  from the source) for a horizontal two-layer model with constant layer velocities  $v_1$  and  $v_2$ , layer thickness  $h_1$  and  $h_2$  over a half-space with velocity  $v_3$ . Other abbreviations stand for:  $t_{ic}^1$  and  $t_{ic}^2$  – intercept times at  $x = 0$  of the extrapolated travel-time curves for the “head-waves”, which travel with  $v_2$  along the intermediate discontinuity between the layers 1 and 2 and with  $v_3$  along the discontinuity between layer 2 and the half-space, respectively.  $x_{cr}^1$  and  $x_{cr}^2$  mark the distances from the source at which the critically reflected rays from the bottom of the first and the second layer, respectively, return to the surface. Beyond  $x_{co}^1$  and  $x_{co}^2$  the head-waves from the bottom of the first and the second layer, respectively, become the first arriving waves ( $x_{co}$  - crossover distance). Rays and their corresponding travel-time curves are shown in the same color. The full **red** (respectively **violet**) travel-time curve relates to the super-critical reflections ( $i > i_{cr}$ ) from the intermediate (respectively lower) discontinuity while the **dotted red** (**violet**) travel-time curve refers to the respective pre-critical ( $i < i_{cr}$ ) steep angle reflections.

In the case of horizontal layering as in Fig. 2.36 the layer and half-space velocities can be determined from the gradients  $dt/dx$  of the yellow, green and blue travel-time curves which correspond to the inverse of the respective layer velocities. When determining additionally the related intercept times  $t_{ic}^1$  and  $t_{ic}^2$  by extrapolating the green and blue curves, or with help of the crossover distances  $x_{co}^1$  and  $x_{co}^2$ , then one can also determine the layer thickness  $h_1$  and  $h_2$  from the following relationships:

$$h_1 = 0.5 x_{co}^1 \sqrt{\frac{v_2 - v_1}{v_1 + v_2}} = 0.5 t_{ic}^1 \frac{v_1 \cdot v_2}{\sqrt{v_2^2 - v_1^2}} \quad \text{and} \quad h_2 = \frac{t_{ic}^2 - 2h_1 \sqrt{v_2^2 - v_1^2} (v_1 \cdot v_2)^{-1}}{2\sqrt{v_3^2 - v_2^2} \cdot (v_2 \cdot v_3)^{-1}}. \quad (2.18)$$

For calculating crossover distances for a simple one-layer model as a function of layer thickness and velocities see Equation (6) in IS 11.1.

In the case where the layer discontinuities are tilted, the observation of travel-times in only one direction away from the seismic source will allow neither the determination of the proper sub-layer velocity nor of the differences in layer thickness. As can be seen from Fig. 2.37, the intercept times, the cross-over distances and the apparent horizontal velocities for the critically refracted head-waves differ when observed down-dip or up-dip from the source although their total travel times to a given distance from the source remain constant. Therefore, especially in controlled-source seismology, *countershot* profiles are deliberately designed so as to identify changes in layer dip and thickness.



**Fig. 2.37** Schematic travel-time curves for direct waves and head waves in a single-layer model with inclined lower boundary towards the half-space. Note the difference between up-dip and down-dip observations (“countershot profile”).  $t_{ic}^-$  and  $v_2^-$  are the intercept time and related apparent velocity of the down-dip head wave,  $t_{ic}^+$  and  $v_2^+$  the respective values for the up-dip travel-time curve.

For the considered simple one-layer case in Fig. 2.37 the dip angle  $\phi$  and the orthogonal distance  $h_1$  to the layer boundary underneath the seismic source on the left can be determined from the following relations:

$$\phi = \frac{1}{2} [\sin^{-1} (v_1/v_2^-) - \sin^{-1} (v_1/v_2^+)] \quad (2.19)$$

and

$$h_1 = \frac{1}{2} t_{ic}^- [v_1 v_2 / \sqrt{(v_2^2 - v_1^2)}]. \quad (2.20)$$

### 2.5.3.6 Wiechert-Herglotz inversion

In the case of velocity  $v = f(z)$  increasing monotonously with depth  $z$ , as in Fig. 2.29, a continuous travel-time curve is observed because all rays return back to the surface. The epicenter distance  $x = D$  of their return increases with decreasing  $p$ , i.e.  $dx/dp < 0$ . The related travel-time curve, with  $dt/dx > 0$  is termed *prograde*. In this case an exact analytical solution of the inverse problem exists, i.e., when knowing the apparent horizontal velocity  $c_x(D) = v_o/\sin i_o = dD/dt$  at any point  $D$ , we know the velocity  $v_{tp}$  at the turning point of the ray that returns to the surface at  $D$ . Thus one can calculate the depth  $z(p) = z_{tp}$  of its turning point. The following relations were developed by Wiechert and Herglotz for the return distance  $D(p)$  and the depth of the turning point  $z(p)$  of a given ray (see Wiechert and Geiger, 1910):

$$D(p) = 2 \int_0^{z(p)} \frac{p v(z)}{\sqrt{1 - p^2 v(z)^2}} dz \quad (2.21)$$

and

$$z(p) = \frac{1}{\pi} \int_0^D \cosh^{-1} \frac{c_x(D)}{c_x(x)} dx \quad (2.22)$$

Note, however, that the velocity  $v_{tp}(p)$  determined from  $dx/dt$  at distance  $x = D$  does always relate to the respective depth half way between source and station! Nevertheless, practically all one-dimensional Earth models have been derived this way assuming that lateral variations of velocity are negligible as compared to the vertical velocity variations.

## 2.5.4 Amplitudes and phase distortions

### 2.5.4.1 Energy and amplitudes of seismic waves

The energy density  $E$  contained in a seismic wave may be expressed as the sum of kinetic ( $E_{kin}$ ) and potential ( $E_{pot}$ ) energy densities :

$$E = E_{kin} + E_{pot} \quad (2.23)$$

The potential energy results from the distortion of the material (strain; see. Figs. 2.2) working against the elastic restoring force (stress) while the kinetic energy density is

$$E_{kin} = \frac{1}{2} \rho a_v^2, \quad (2.24)$$

where  $\rho$  is the density of the material,  $a_v = A \omega \cos(\omega t - kx)$  is the ground-motion particle velocity, with  $A$  - wave amplitude,  $\omega$  - angular frequency  $2\pi f$  and  $k$  - wavenumber. Since the mean value of  $\cos^2$  is  $\frac{1}{2}$  it follows for the average kinetic energy density  $\bar{E}_{kin} = \frac{1}{4} \rho A^2 \omega^2$ , and with  $E_{kin} = E_{pot}$  in case of an isotropic stress-strain relationship in a non-dispersive (closed) system for the average energy density

$$\bar{E} = \frac{1}{2} \rho A^2 \omega^2. \quad (2.25)$$

The energy-flux density per unit of time in the direction of wave propagation with velocity  $v$  is then

$$E_{flux} = \frac{1}{2} v \rho A^2 \omega^2 \quad (2.26)$$

and the total energy-flux density  $E_{flux}$  through a small surface area  $dS$  of the wavefront bounded by neighboring rays which form a *ray tube*

$$E_{flux} = \frac{1}{2} v \rho A^2 \omega^2 dS. \quad (2.27)$$

When considering only waves with wavelengths being small as compared to the inhomogeneities of the medium of wave propagation (high-frequency approximation), then we can assume that the seismic energy only travels along the rays. According to the energy conservation law, the energy flux within a considered ray tube must remain constant although

the surface area  $dS$  of the wavefront related to this ray tube may vary along the propagation path due to focusing or defocusing of the seismic rays (e.g., Figs. 2.30 and 2.33). Considering at different times two surface patches of the propagating wavefront  $dS_1 \neq dS_2$ , which are bounded by the same ray tube, and assuming that  $v$  and  $\rho$  are the same at these two locations then

$$A_1/A_2 = (dS_2/dS_1)^{1/2}, \quad (2.28)$$

i.e., the amplitudes vary inversely as the square root of the surface area of the wavefront patch bounded by the ray tube. Thus amplitudes increase due to ray focusing, which is particularly strong at caustics (see 2.5.3.2) and decrease when the wavefront spreads out (e.g., Figure 3a in IS 2.1).

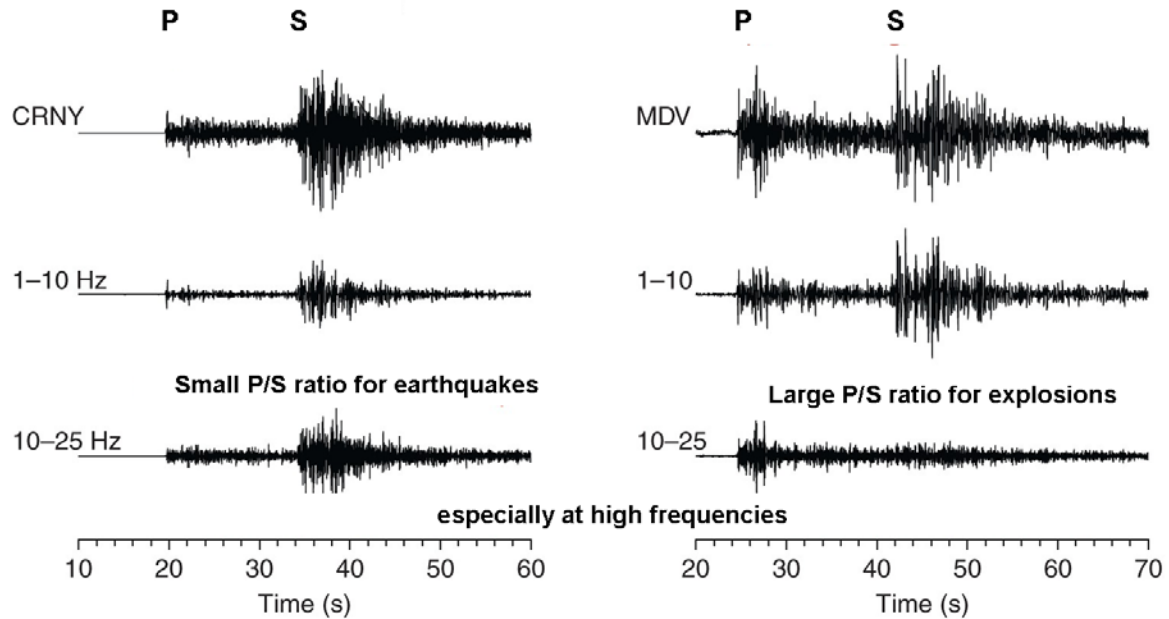
Also, for a spherical wavefront (e.g., body-wave propagation in a homogeneous isotropic medium) the surface area grows with  $r^2$  and for a cylindrical wavefront (e.g., for surface waves) with distance  $r$ . Accordingly, the wave-amplitude decay is in the former case  $\sim r$  and in the latter case  $\sim \sqrt{r}$ . This difference in *geometrical spreading* is the main reason for the domination of surface wave amplitudes in seismic records of shallow events (e.g., Fig. 2.12 and Fig. 3.34).

However, wave amplitudes will also change, even in the absence of geometrical spreading, when density  $\rho$  and velocity  $v$  vary at different locations along the ray path. We then get

$$A_1/A_2 = [(\rho_2 v_2)/(\rho_1 v_1)]^{1/2}. \quad (2.29)$$

The product  $\rho v$  is termed the *impedance* of the material and  $(\rho_2 v_2)/(\rho_1 v_1)$  is the *impedance contrast* between the two adjacent media  $m_1$  and  $m_2$ . The latter largely controls the reflection and transmission coefficients at the media discontinuity. From Eq. (2.29) it follows that seismic amplitudes will increase as waves propagate into media of lower density and wave propagation velocity. This has two important implications. On the one hand, seismic stations on hard bedrock tend to record smaller amplitudes and thus to slightly underestimate event magnitudes as compared to stations on average or soft-soil conditions. On the other hand, ground shaking from strong earthquakes is usually more intense on top of unconsolidated sediments as compared with nearby rock sites. Additionally, reverberations and resonance within the unconsolidated near-surface layers above the basement rocks may significantly amplify the amplitudes at soft-soil sites. This may increase significantly local seismic hazard.

In this context one should mention that in earthquake records amplitudes of S waves are at comparable period on average about five times larger than those of P waves (see Fig. 2.3 and Figure 1 in EX 3.4). This follows from Eq. (3.2) in Chapter 3 or from the far-field term of the Green's function when modeling earthquake shear sources [see Eq. (24) in the IS 3.1], taking into account that  $v_P \approx v_S \sqrt{3}$ . This would correspond to an average ratio between S-wave and P-wave energy radiation of about 25. The ratio, however, depends on the specifics of the considered source mechanisms. Published numbers vary between about 13.7 and 40.5 (e.g., Aki and Richards, 1980; Boatwright and Fletcher, 1984; Boatwright and Quin, 1986, Duda and Yanovskaya, 1993), yet empirical data hint to values closer to the lower bound. The S-/P-wave amplitude ratio in individual seismic records additionally depends strongly on the position (azimuth and distance) of the station with respect to the radiation pattern and directivity of the source (see Chapter 3, section 3.4). For explosions, however, the average ratio is usually much smaller and may become even less than 1, depending also on frequency (see Fig. 2.38).



**Fig. 2.38** Typical records of a local earthquake ( $D = 125$  km) and a local explosion ( $D = 149$  km) in broadband (top traces) and high-frequency filtered records (middle and bottom traces). According to data from Kim et al. (1993), *Geophysical Research Letters*, **20**, 1507–1510.

Also, the periods of S waves tend to be longer than those of P waves, at least by a factor of  $\sqrt{3}$ , due to the differences in wave propagation velocity and the related differences in the corner frequencies of the P- and S-wave source spectrum.

For shallow earthquakes the amplitudes of surface waves are generally much larger than those of body-waves (see Fig. 2.12 and many more examples in DS 2.1-2.2), whereas the amplitude ratios of primary P and S waves with respect to their secondary (often multiple) reflected and converted waves varies strongly according to distance and incidence-angle dependent reflection/conversion coefficients (e.g., Fig. 2.35). No simple rules can be given and proper phase identification on the basis of relative travel-times, slowness, amplitude and period relationships requires a lot of practical experience.

For overview papers with many references on the calculation of released seismic wave energy, based on the integration of squared broadband velocity records or moment rate spectra, see IS 3.6 by G. Choy in this Manual, section 3.2.7.2 in Chapter 3 and Di Giacomo and Bormann (2011).

#### 2.5.4.2 Wave attenuation

Amplitudes of seismic waves are not only controlled by geometrical spreading, focusing/defocussing and by the reflection and transmission coefficients at discontinuities. Wave amplitudes may also be reduced because of energy loss due to viscoelastic material behavior and thus internal friction and heat generation during wave propagation. This effect is called *intrinsic attenuation*. Also, scattering of energy at heterogeneities along the travel paths with scale lengths in the order of wavelength may reduce amplitudes of seismic waves, change the shape of the initial cycle of a body wave pulse as well as of the motion immediately following the initial cycle, termed *coda*, with high-frequency energy being

transferred into the coda (see Fig. 2.44). The separation of intrinsic and scattering attenuation is not trivial. For a review see Cormier (2009). In the case of *scattering attenuation*, however, the integrated energy in the total wavefield remains constant, while *intrinsic attenuation* results in loss of mechanical wave energy, e.g., by transformation into heat. The wave attenuation is usually expressed in terms of the dimensionless *quality factor*  $Q$

$$Q = 2\pi E/\Delta E \quad (2.30)$$

where  $\Delta E$  is the dissipated energy per cycle. Thus,  $Q$  is a measure of the area contained in the hysteresis loop of a stress-strain cycle. **Large energy loss means low  $Q$  and vice versa, i.e.,  $Q$  is inversely proportional to the attenuation.** In a simplified way we can write for the decay of source amplitude  $A_0$  with distance  $x$

$$A = \frac{A_0}{x^n} e^{-\frac{\omega t}{2Q}} = \frac{A_0}{x^n} e^{-\frac{\pi x}{QTv}} = \frac{A_0}{x^n} e^{-\frac{\omega x}{2Qv}}, \quad (2.31)$$

with  $A_0/x^n$  – the geometrical spreading term,  $\exp(-xt/2Q) = \exp(-\pi/QTv)$  – the attenuation term,  $\omega$  – angular frequency  $2\pi/T$ ,  $T$  – period of wave,  $t$  – travel time,  $v$  – propagation velocity of wave, and  $n$  – exponential factor controlled by the kind of geometric spreading. According to experimental data,  $n$  varies between about 0.3 and 3, depending also on the type of seismic wave and distance range considered.

In ray theoretical methods, attenuation may be modeled through the use of the parameter  $t^*$  that is defined as the integrated value of the travel time divided by  $1/Q$

$$t^* = \int_{\text{path}} \frac{dt}{Q(\vec{r})}, \quad (2.32)$$

where  $\vec{r}$  is the position vector. Accordingly, we can then write Eq. (2.31) as

$$A(\omega) = A_0(\omega) e^{-\omega t^*/2}. \quad (2.33)$$

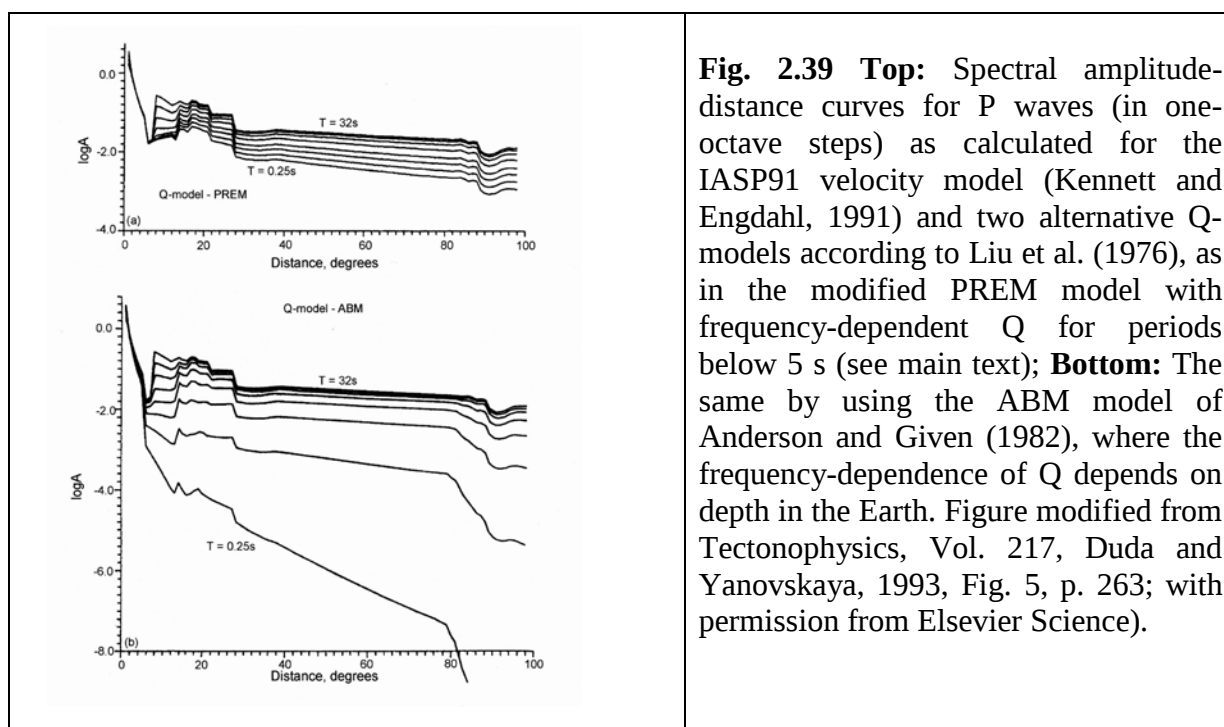
Note, that P-wave attenuation  $Q_\alpha$  and S-wave attenuation  $Q_\beta$  differ. They are related to the shear attenuation  $Q_\mu$  and the bulk attenuation  $Q_\kappa$  by the relationships

$$Q_\beta = Q_\mu \quad \text{and} \quad 1/Q_\alpha = (4/3)(\beta/\alpha)^2/Q_\mu + [1 - (4/3)(\beta/\alpha)^2]/Q_\kappa. \quad (2.34)$$

with P-wave velocity  $\alpha = v_p$  and S-wave velocity  $\beta = v_s$ . In the Earth shear attenuation is much stronger than bulk attenuation. While  $Q_\mu$  is smallest (and thus shear attenuation strongest) in the upper mantle and zero in the outer core (see DS 2.1),  $Q_\kappa$  is generally assumed to be close to infinite, except in the inner core. While the P- and S-wave velocities are rather well known and do not differ much between different Earth models, the various model assumptions with respect to  $Q_\alpha$  and  $Q_\beta$  as a function of depth still differ significantly (see Fig. 2.79). According to the PREM model,  $Q_\mu$  is 600 for less than 80 km depth. It then drops between 80 and 220 km to 80, increases to 143 from 220 to 670 km, and is 312 for the lower mantle below 670 km depth. For pure viscoelastic attenuation in Poisson solid  $t_s^*$  is about 4 times larger than  $t_p^*$ . However, there are indications that this difference decreases for higher

frequencies because apparent  $1/Q_P$  approaches  $1/Q_S$  at frequencies  $> 1\text{Hz}$ . This hints to an increasing effect of scattering attenuation on P waves at higher frequencies (Cormier, 2009).

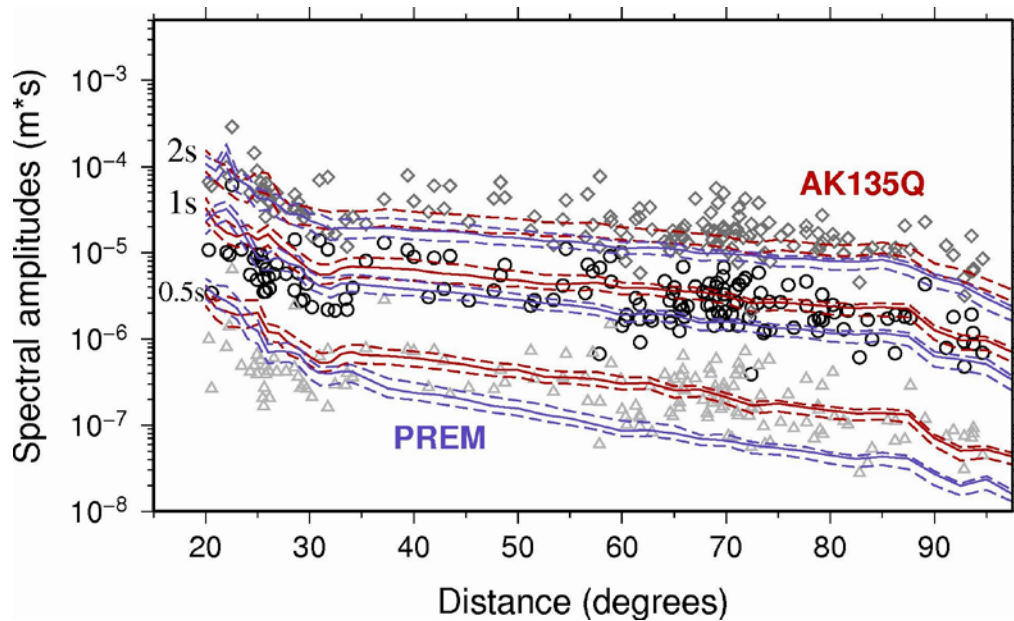
One of the important questions in this context is, whether the quality parameter  $Q$  itself depends on frequency or not. According to Dziewonski and Anderson (1981), seismological observations are consistent with the hypothesis of  $Q$  being independent on period only down to periods 5-10 s but that they provide no information about attenuation of the waves at periods less than 5 s. However, experimental studies (Berckhemer et al., 1982; Ulug and Berckhemer, 1984) showed that within the absorption band  $Q$  varies with frequency. Choy and Cormier (1986) proved this by direct measurement of the mantle attenuation operator from broadband P and S waves, thus resolving the difference between results from the normal-mode long-period community and the high-frequency (e.g., nuclear monitoring) communities. In order to explain the amplitudes of seismic spectra in the 1-10 Hz frequency band, it was necessary to assume an increase of  $Q$  with frequency, even when assuming white source spectra (Cormier, 2009). Therefore, Duda and Yanovskaya (1993) calculated a hybrid set of theoretical spectral  $\log A$ - $\Delta$  curves for P-waves (upper diagram in Fig. 2.39). It rests on the frequency independence of the  $Q$  in the PREM model (see Fig. 2.79 and DS 2.1) down to 5 s but introduces for shorter periods a frequency-dependence of  $Q$  on the basis of the absorption-band model proposed by Liu et al. (1976) and Anderson et al. (1976), assuming a relaxation time of 0.16 s. This has the effect that at 1 Hz  $Q^{-1}$  has already only half the value of the frequency independent  $Q^{-1}$  in the PREM model.



According to the “hybrid” PREM model no strong increase of frequency-dependent absorption has to be expected for periods below 4 s and the difference in  $\delta \log A$  of spectral amplitudes at 2 s and 0.5 s, which covers the main range of periods at which the short-period body-wave magnitude  $m_b$  is measured, is only about 0.3. When investigating regional variations and frequency-dependence of anelastic attenuation in the mantle under the United States in the 0.5-4 Hz band Der et al. (1982) found about the same difference for the respective spectral P-wave amplitudes, whereas the difference for S-wave  $\log A$  was about 1. In contrast, recent investigations by D. Di Giacomo (personal communication 2011) yielded

on the basis of the frequency-independent AK135 and PREM Q-models as well as empirically much larger attenuation-caused differences of spectral amplitudes between 2 s and 0.5 s, ranging between about 1.8 to 2.2 log-units (see Fig. 2.40). This compares even better with the calculations by Duda and Yanovskaya (1993) for the alternative ABM model proposed by Anderson and Given (1982) in which the parameters of the absorption-band vary with depth in the Earth (Fig. 2.39, lower diagram), resulting in  $\delta \log A$  for periods of 2 s and 0.5 s in the distance range  $20^\circ < \Delta < 100^\circ$  between about 1.2 and 2.6 log-units. Such large differences between currently existing attenuation models are not yet fully understood or supported by high-quality empirical data and the way in which the latter are processed and interpreted, usually again on the basis of simplified model assumptions. This requires more thorough specialized in-depth studies in future.

For periods longer than about 4 s, however, all the here presented different models yield about the same results in agreement with the assumption of frequency-independent Q. Most importantly, however, is to realize that theoretical calculations based on global 1-D models can at best describe and allow to account for the average trend of amplitude decay with distance due to geometric spreading and inelastic attenuation but not for the very significant scatter of the real data in the order of about  $\pm 0.5$  to 0.7 log-units (see Fig. 2.40). The latter is mainly controlled by local effects at the station sites such as spectral amplification due to soft-layer cover over bedrock (see Chapter 14), focussing and de-focussing effects due to 3-D velocity and topographic anomalies and/or source radiation differences into the direction of the considered different station azimuth at comparable distances, rather than by local or regional anomalies in inelastic attenuation.



**Fig. 2.40** Displacement spectral amplitudes at 0.5 s (2 Hz, light gray triangles), 1 s (1 Hz, black circles) and 2 s (0.5 Hz, gray diamonds) for the  $M_w = 7.9$  Wenchuan, China, earthquake of 2008 May 12, superimposed to the theoretical spectral amplitude decay functions for the same frequencies for the AK135 model (median, 25<sup>th</sup> and 75<sup>th</sup> percentile in solid and dashed red lines, respectively) and the PREM model (median, 25<sup>th</sup> and 75<sup>th</sup> percentile in solid and dashed violet lines, respectively). An arbitrary offset has been added to the theoretical curves to make easier the comparison with the real data. (Courtesy of D. Di Giacomo, 2011).



According to Fig. 2.40 the AK135Q model curves seem better than the ones calculated with the PREM model to agree towards higher frequencies with the observed average increments in frequency-dependent attenuation, although both models assume **no** frequency dependence of  $Q$ . Thus, the significance of the absorption-band effect and the necessity of accounting for a frequency dependence of  $Q$  for periods  $0.5 \text{ s} \leq T < 4 \text{ s}$  is still difficult to prove with available noisy empirical data.

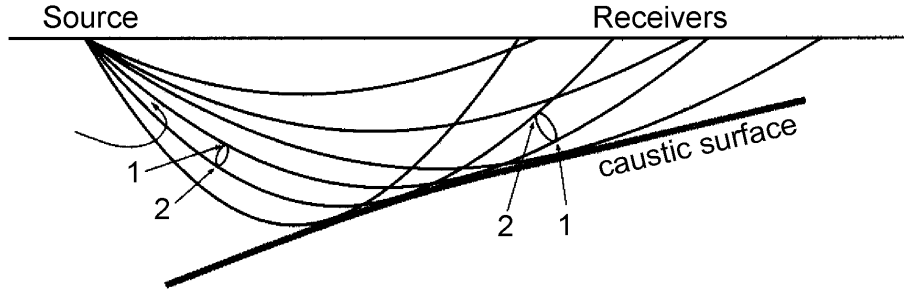
In practice there is another difficulty, namely to separate intrinsic (inelastic)  $Q$  and scattering  $Q$ . Particularly in local earthquake records, which are strongly affected by scattering on crustal inhomogeneities, scattering  $Q$  dominates. Scattering  $Q$  is usually determined from the decay of coda waves following  $S_g$  and  $S_mS$  onsets (e.g., Fig. 2.47) and is called accordingly  $Q_c$ . For an elaborate discussion of these topics see Aki and Richards (1980; pp. 170-182) and for a review about  $Q$  of the Earth from crust to core Romanovicz and Mitchell (2007).

Also note that  $S$  waves are much stronger attenuated than  $P$  waves (see also  $Q$  models and formulas in DS 2.1), thus filtering out more strongly the higher frequencies. Moreover,  $S$  waves do not propagate in the fluid outer core because of vanishing shear module (see Fig. 2.79). Therefore, no direct  $S$  waves are observed beyond  $100^\circ$  epicenter distance (with the exception of long-period  $S$  diffracted around the core-mantle boundary; see Fig. 2.60).

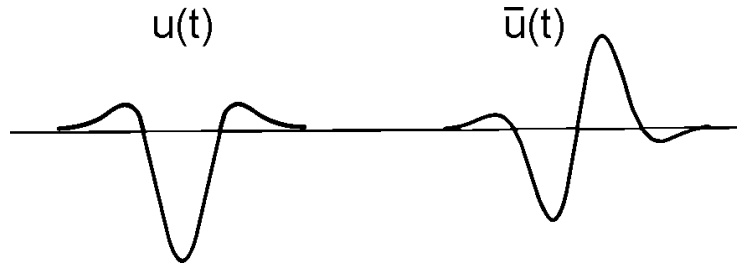
The discussed differences in amplitude-distance relationships have to be compensated by wave-type dependent calibration functions in order to be able to derive comparable magnitude values for seismic events based on amplitude readings from different types of seismic waves (see 3.2). However, for teleseismic body- and surface-wave magnitudes, based on measuring the ratio  $(A/T)$ , or directly ground motion velocity (see IS 3.3), the effect of exponential frequency-dependent attenuation on  $A$  is reduced by multiplying  $A$  with  $f = 1/T$ , making the related calibration functions less dependent on variations in period.

### 2.5.4.3 Phase distortions and Hilbert transform

As shown in Fig. 2.27 seismic rays will curve in the case of a vertical velocity gradient and thus seismic wavefronts will no longer be planar. Nevertheless, locally, between adjacent rays defining a *ray tube*, the wavefront still can be considered as a plane wavefront. In the case of strong gradients, retrograde travel-time branches will develop because rays bend stronger, cross each other and the wavefront folds over itself at the turning point (Fig. 2.41). Accordingly, a local plane wavefront traveling through a strong vertical velocity gradient will experience a constant, frequency-independent  $\pi/2$  phase advance at the turning point. The envelope of turning points of these crossing bended rays is termed an *internal caustic surface*. Because of the  $-\pi/2$  phase shift the up-going plane wave is the *Hilbert transform* of the down-going wave. More generally, whenever a ray has a non-pure minimum ray path (see 2.5.2) it touches such a caustic. Consequently, its pulse shape is altered (see Fig. 2.42). Example: The Hilbert transform of a pure sine wave is a cosine wave. In the case of seismic waves this phase shift by  $-\pi/2$  has to be applied to each single frequency represented in the seismic pulse. This results in the known pulse shape alterations.



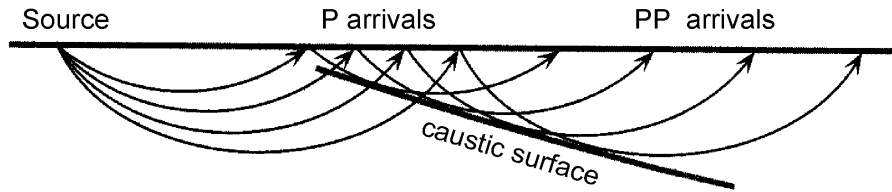
**Fig. 2.41** In a medium with steep vertical velocity gradient, seismic rays with larger take-off angles from the source turn back towards the source thus forming a retrograde branch of a travel-time curve. The crossing of ray paths forms an internal caustic surface that produces a  $-\pi/2$  phase shift in the waveforms (according to Choy and Richards, 1975; modified from Shearer, Introduction to Seismology, 1999; with permission of Cambridge University Press).



**Fig. 2.42 Left:** A typical seismic pulse; **Right:** Its Hilbert transform.

Generally, in the case of a steep velocity gradient producing a retrograde travel-time branch, ray theory predicts that seismic wave arrivals along this branch are Hilbert transformed compared to the prograde branches. One should note, however, that also in case of relatively weak vertical velocity gradients, which do not produce a retrograde travel-time branch for the direct wave, the related reflected phase might nevertheless be Hilbert transformed (see Fig. 2.43 for PP waves). On the other hand, when the gradient becomes too steep, or in the case of a first-order velocity step discontinuity, the post-critical reflection coefficients for such an interface involve a continuous change in phase with ray angle. The phase shift may then acquire any value other than a constant  $-\pi/2$  phase shift.

Without exception, all the distorted waveforms bear little or no resemblance to the original waveforms. Accordingly, neither their onset times (first arrival of energy) nor the relative position of peaks and troughs of the distorted waveforms appear at the times that are theoretically predicted by ray theory. This biases onset-time picking, related travel-time determinations as well as waveform correlations between primary and Hilbert transformed phases. Therefore, modern digital data-analysis software can routinely apply the inverse Hilbert transform to phases distorted by internal caustics. The following major teleseismic body-wave phases are Hilbert transformed: PP, PS, SP, SS, PKPab, pPKPab, sPKPab, SKKSac, SKKSdf, P'P', S'S'ac. For the nomenclature of these phases and their travel paths see Figures 2.57-2.58 and IS 2.1). However, many phases pass caustics several times in the Earth and then the final pulse shapes are the sum of all internal caustic effects.



**Fig. 2.43** Ray paths for the surface reflected phase PP. Note that the rays after the reflection points cross again and form an internal caustic. Accordingly, PP is Hilbert transformed relative to P and additionally has an opposite polarity (phase shift of  $\pi$ ) due to the surface reflection (from Shearer, Introduction to Seismology, 1999; with permission from Cambridge University Press).

#### 2.5.4.4 Effects not explained by ray theory

As mentioned above, ray theory is a high-frequency approximation that does not cover all aspects of wave propagation. Although detailed wave-theoretical considerations are beyond the scope of this Manual we will shortly mention three major phenomena that are of practical importance and not covered by ray theory.

##### *Head waves*

As mentioned in 2.5.4.1, seismic waves impinging at a discontinuity between the layers  $n$  and  $n+1$  with  $v_n < v_{n+1}$  under the critical incidence angle  $i_c$  with  $\sin i_c = v_n/v_{n+1}$  are refracted into this discontinuity with the angle  $i_{n+1} = 90^\circ$ . There they travel along (or just below) this discontinuity with the velocity  $v_{n+1}$  of the lower faster medium. Such kind of *inhomogeneous* waves (see section 2.5.3.4) are usually referred to as *head waves*. They have the unique property to transmit energy back into the upper medium at exactly the critical angle  $i_c$ , however with amplitudes which decay with distance  $\sim 1/r^2$ . Therefore, the amplitudes of true head-waves are rather small as compared to those of direct, reflected or converted waves. The travel-time curve of a head wave is a straight line with the slope of  $1/v_{n+1}$  (see Fig. 2.36). This provides a convenient and direct measure of the sub-discontinuity velocity. Head waves are of particular importance for crustal studies and in the analysis of seismic records from local and regional seismic events (see 2.6.1).

##### *Seismic Diffraction*

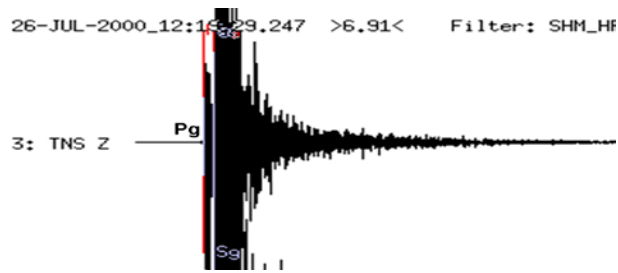
Diffraction, analog to optics, is the phenomenon of transmission of energy by non-geometric ray paths. In optics, the classic example is the diffraction of light “leaking” around the edge of an opaque screen. In seismology, diffraction occurs whenever the radius of curvature of a reflecting interface approaches the order of wavelength of the propagating wave. Seismic diffraction is important for example in steep-angle reflection data in the presence of sharp boundaries. But there are also long-period diffracted waves such as Pdif and Sdif which are “bended” around the core-mantle boundary into the core shadow zone beyond about  $100^\circ$  epicenter distance. Only little short-period P- and S-wave energy is observed in this shadow zone. In fact, the edge of a discontinuity/impedance contrast acts like a secondary source according to Huygens's principle and radiates energy forward in all directions.

Diffractions can also be understood from the standpoint of *Fresnel zones*. This concept states that waves are not only reflected at a considered point of the discontinuity (like a seismic ray) but also from a larger surrounding area. The radius of the so-called first Fresnel zone is about  $\frac{1}{2}$  wavelengths around a considered ray arriving at a station, i.e., the range within which reflected energy might interfere constructively. The wavelength-dependent width of this Fresnel zone also determines the geometrical resolution of objects/impedance contrasts that can be at best achieved by seismic (or optical) methods.

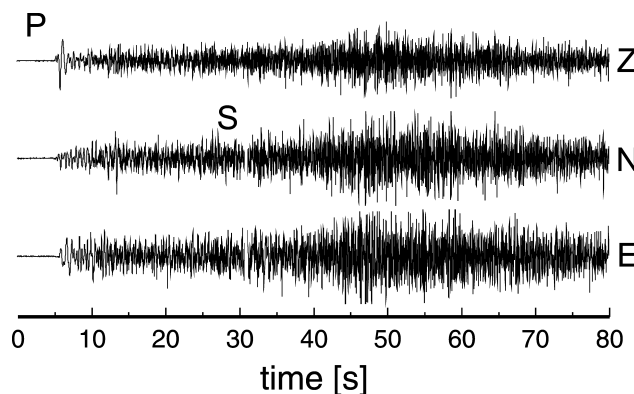
Since the real Earth may significantly deviate from simplified global one-dimensional models, scattering and diffraction effects render not only amplitudes but also travel times of more low-frequency waves sensitive to the 3-D structure off the seismic rays. This has to be taken into account when making use of recent developments of automated travel-time measurement techniques which use cross-correlation of observed body wave phases in digital broadband records with the corresponding synthetic phases possible in spherical Earth. Marquering et al. (1999) showed that for an SS wave observed at an epicenter distance of  $80^\circ$ , near-surface heterogeneities situated more than  $15^\circ$  from the bounce point at  $40^\circ$  can exert a significant influence upon the travel time of an SS wave. They conclude that geometrical ray theory, which has been a cornerstone of seismology for about a century and proven useful in most practical applications, including earthquake location and tomography, is, however, valid only if the scale length of the 3-D heterogeneities is much greater than the seismic wavelengths. In other words, the validity of ray theory is based on a high-frequency (short-period) approximation. However, intermediate-period and long-period seismic waves, with wavelength of the order of 100 – 1000 km, already have comparable scale lengths with 3-D anomalies in current global tomographic models. When investigating smaller 3-D structures and applying new methods of waveform correlation, these wave-theoretical considerations gain growing importance, probably even in future observatory routines. In this context we refer to section 2.7.2 on 3-D Earth models and their use in event location and to the establishment, in 2012, of an international “Working Group on the Utilization of Global 3-D Models for Seismic Observatory Applications” by the IASPEI Commission on Seismic Observation and Interpretation (CoSOI).

### ***Scattering of seismic waves***

Often the primary arrivals are followed by a multitude of later arrivals that can not be explained by simple 1-D models. The complex wave train following the primary arrival(s), often with exponentially decaying amplitudes, is called coda (Figs. 2.44 and 2.45). Coda arrivals are produced by scattering, that is, the wavefield's interaction with small-scale heterogeneities. Heterogeneity at different length scales is present almost universally inside the Earth. Seismic coda waves can be used to infer stochastic properties of the medium, i.e., scale the amplitude of the average heterogeneities, to estimate coda  $Q_c$  which is particularly needed for correcting source spectra prior to deriving spectral source parameters from records of local events (see exercise EX 3.4), and to estimate earthquake magnitude (see Chapter 3). Moreover, according to Rautian et al. (2007) there is great potential in using the later part of the coda for spectral source parameter estimation such as seismic energy, seismic moment, apparent stress, etc. by means of multi-bandpass filtering, first proposed by Rautian and Khalturin (1978).



**Fig. 2.44** Clearly developed coda waves of the high-pass filtered (0.7 to 4 Hz) filtered record of a local earthquake on 26.07.2000 near Limburg/Lahn in Germany ( $D = 29$  km,  $M_l = 3.5$ ) at station TNS.



**Fig. 2.45** Three-component seismogram of a local, 100 km deep earthquake recorded at a portable station on the active volcano Lascar in northern Chile. The P-wave arrival is followed by coda waves produced by heterogeneous structure in the vicinity of the volcano (courtesy of B. Schurr, 2001).

## 2.6 Seismic phases and travel times in real Earth (P. Bormann)

Seismic body waves propagate three-dimensionally and are more strongly affected than surface waves by refraction, reflection and mode conversion at the main impedance contrasts in the radial direction of the Earth. This gives rise, with growing distance, to the appearance of more and more secondary seismic body-wave phases following the direct P- and S-wave arrivals in seismic records (e.g. Fig. 2.25). And since body waves show no dispersion in the considered frequency range these phases can usually be well observed and discriminated from each other as long as their travel-time curves do not overlap.

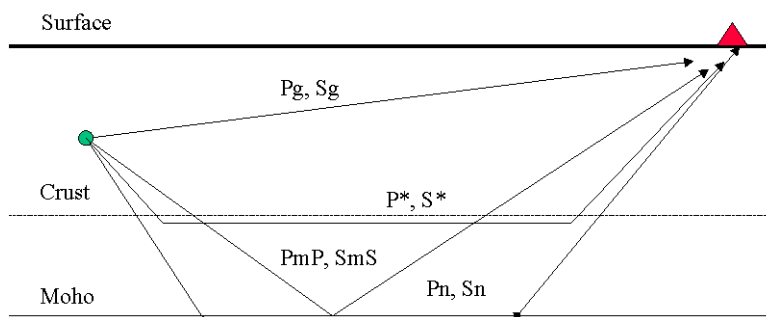
All of these secondary phases have a special story to tell about the geometrical and physical properties of the discontinuities which they encountered during their travel through the Earth's interior and which have shaped their waveforms and influenced their amplitude and frequency content. Therefore, the proper identification and parameter or waveform reporting about later phases in seismic records to relevant data centers is an important duty of seismological observatories. In addition, the complementary use of secondary phases significantly improves the precision and accuracy of seismic event locations, their source depth in particular (see Figure 7 in IS 11.1 and section 2.6.3). In the following, we will introduce the main types of seismic body-wave phases that can generally be observed at local, regional and teleseismic distance ranges. They should be identified, analyzed and reported by

seismic observatories or data analysis centers. Basic features of their travel-time curves, polarization, waveforms and frequency range of observation, which can guide their identification, will be presented.

### 2.6.1 Seismic phases and travel times from local and regional events

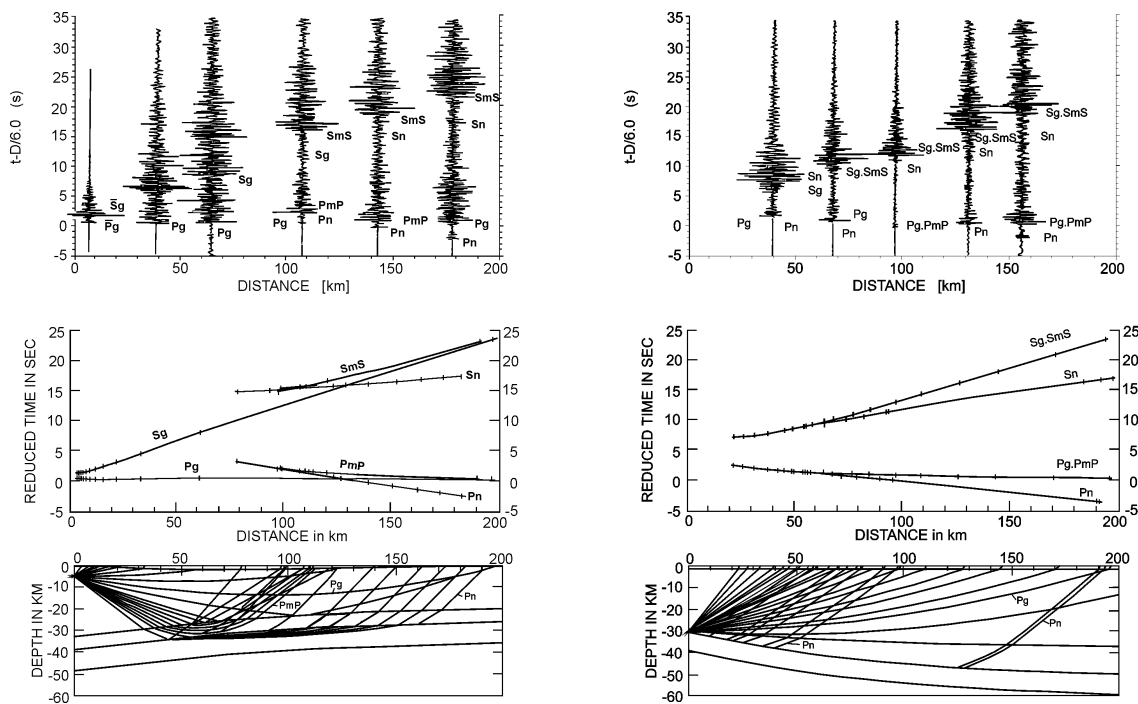
Seismic waves arriving at stations at local distances of up to about 150 km or regional distances of up to about 15° (1° = 111.2 km) from the seismic source have traveled exclusively or dominantly through the crust or the sub-crustal uppermost mantle. The crust varies strongly in its thickness (see Fig. 2.11), petrologic composition and internal structure due to folding and faulting processes in the past. The resulting strong heterogeneities in physical properties at scale length of several decameters to several km cause intensive scattering of P and S waves in the typical frequency range for the recording of near seismic events (about 0.5 to 50 Hz). Therefore, primary wave onsets are usually followed by signal-generated noise or coda waves that make it difficult to identify later seismic phases reflected or refracted from weaker intra-crustal discontinuities. It is usually only the significant velocity increase of about 20% at the base of the crust towards the upper mantle (*Mohorovičić discontinuity*, or *Moho* for short), which produces first or later wave onsets besides the direct P and S waves that are strong enough to be recognizable above the ambient or signal-generated noise level. Only in some continental regions may an intermediate discontinuity, named the *Conrad discontinuity* after its discoverer, produce recognizable critically refracted (Pb = P\*; Sg = S\*) or reflected waves (see Fig. 2.46). Accordingly, for purposes of routine seismological observatory practice, it is usually sufficient to represent the crust as a horizontal one-layer model above the half-space (upper mantle).

Both of the widely used global 1-D Earth velocity models, IASP91 by Kennett and Engdahl (1991) and AK135 by Kennett et al. (1995) assume a homogeneous 35 km thick two-layer crust with the intermediate crustal discontinuity at 20 km depth (see DS 2.1). The respective average velocities for the upper and lower crust and the upper mantle are for P waves 5.8 km/s, 6.5 km/s and 8.04 km/s, and for S waves 3.36 km/s, 3.75 km/s and 4.47 km/s, respectively. The *impedance contrast* at the Conrad discontinuity and the Moho is about 1.3. Fig. 2.46 is a simplified depiction of such a two-layer crust and of the seismic rays of the main crustal/upper mantle phases to be expected. These are: Pg, Sg, Pb, Sb, Pn, Sn, PmP and SmS. For detailed definition of the phase names see IS 2.1 or Storchak et al. (2003 and 2011).

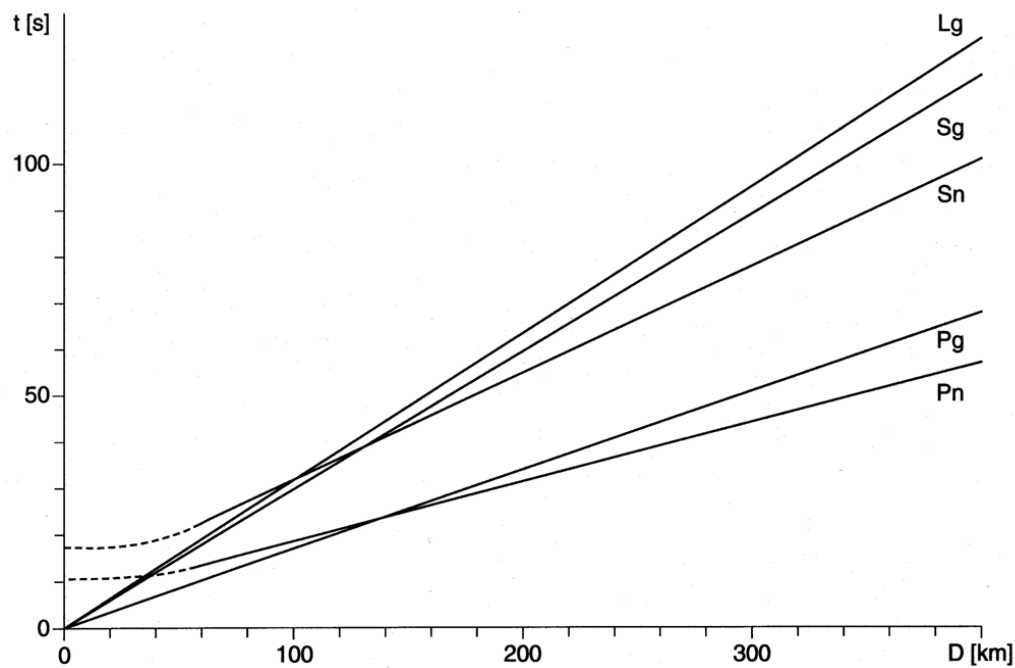


**Fig. 2.46** A simplified model of the crust showing the ray traces of the main “crustal phases” observed for near (local and regional) earthquakes. Note: P\* = Pb and S\* = Sg.

The apparent horizontal velocity of the reflected PmP and SmS waves varies with distance according to their changing incidence angle on the surface. Their travel-time branches form hyperbolae that approach asymptotically the travel-time curves for Pg and Sg (or Pb and Sb) with increasing distance (see Fig. 2.47). Note that Pn, if a head wave, has usually smaller amplitudes than Pg and Sg, at least for distances up to about 300 km. Pn can be usually identified above the noise level only when it becomes the P-wave first arrival. At larger distances, because of the stronger attenuation of upper crustal Pg and Sg and with Pn and Sn being less attenuated as typically upper mantle diving phases, Pn and Sn may become clear P and S first arrivals (see Fig. 2.17). Beyond the critical point (at about 70-80 km distance for an average crust) the super-critically reflected waves PmP and SmS have generally the largest amplitudes, however, arriving always closely after Pg and Sg, their onset times can usually not be picked reliably enough as to be of value for earthquake location. Therefore, these phases are usually not explicitly reported in routine observatory practice. However, reporting of Pg, Sg, Pn and Sn, if recognizable, is a must. This also applies to the reporting of the maximum amplitudes in records of near seismic events for the determination of local magnitudes  $M_L$  (see Chapter 3, section 3.2.4). Depending on source depth too, this amplitude maximum may be related to Sg/SmS, Lg, or Rg (see Figs. 2.17, 2.18 and 2.47). A simple average local travel-time curve for a single-layer crust in Central Europe had been derived from related reliable phase observations at European observatories from an earthquake in Hungary decades ago. It is depicted in Fig. 2.48 and proved to be a suitable single-layer crust approximation for routine analysis, phase identification and location of seismic events in the considered area (see EX 11.1). The crustal and sub-Moho velocity data given in DS 2.1 for IASP91 and AK 135, respectively, would permit to calculate a similar global average travel-time curve.



**Fig. 2.47** Records (above) of two regional earthquakes of Oct. 9, 1986 at Sierre (left) and of July 7, 1985 at Langenthal, Switzerland together with the calculated reduced travel-time curves (middle) and ray-tracing crustal models which best fit the observations (below), redrawn and complemented from Kulháněk (1990), *Anatomy of Seismograms*, plate 4, pp. 83-84, © (with permission from Elsevier Science).



**Fig. 2.48** Empirically determined average travel-time curves for the phases Pn, Pg, Sn, Sg and Lg observed at distances up to 400 km, assuming a single-layer crust in Central Europe.

However, crustal travel-time curves based on a global velocity model may not be representative at all for certain regions and may serve as a starting model only to work with. It is one of the main tasks of operators of local and regional seismic networks to derive from their own carefully analyzed data of near events not only local/regional magnitude calibration functions (see 3.2.4.3 and Table 2 in DS 3.1) but also average local/regional 1-D travel-time curves. The latter will not only allow significantly improved seismic event locations but may serve also as starting models for tomographic studies of crustal heterogeneities and the derivation of 3-D velocity models.

Fig. 2.47 above shows real short-period seismic network records of two local earthquakes in Switzerland in the distance range between about 10 km and 180 km along different profiles together with the modeling of their reduced travel-time curves and inferred structural profiles. While one event was at a depth of only 5 km, the other event was about 30 km deep. The first one was observed by stations situated up-dip while the latter event was observed down-dip. The striking differences in the shape and gradient of the travel-time curves and in the crossover distance between Pg and Pn, in particular, are obvious. In the case of the deeper event near to the Moho, Pn becomes the first arrival beyond 70 km distance, whereas for the shallower event Pn outruns Pg only at more than 130 km epicenter distance. In both cases Pg (Sg) and/or PmP (SmS) are the prominent P and S arrivals. The Pn first arrivals are relatively small. No Pb, Sb or reflected waves from a mid-crustal discontinuity are recognizable in Fig. 2.47. Note, however, depending on the orientation of the earthquake rupture and thus of the related radiation characteristic of the source, it may happen that a maximum of energy is radiated in the direction of the Pn ray and a minimum in the direction of the Pg ray. Then the usual relationship  $A_{Pn} < A_{Pg}$  may be reversed (examples are given in Chapter 11, section 11.5.1).

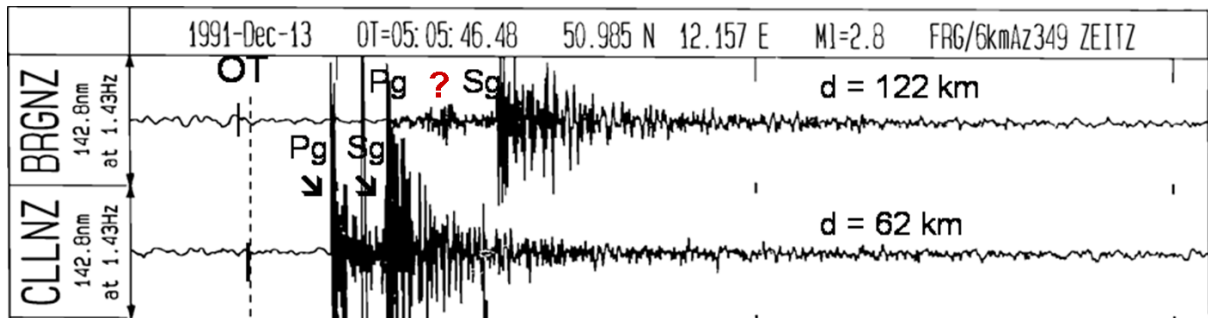


Misinterpretation of Pn as Pg or vice versa may result in large errors of event location. Therefore, one should have at least a rough idea at which distance in the region under study, depending on the average crustal thickness and velocity, one may expect Pn to become the first arrival. A “rule-of-thumb” for calculating the crossover distance  $x_{co}$  is given in Eq. (6) of IS 11.1. For an average single-layer crust and a surface source,  $x_{co} \approx 5 z_m$  with  $z_m$  – Moho depth. However, as demonstrated with Fig. 2.47,  $x_{co}$  is only about half as large for near Moho earthquakes and also the dip of the Moho and the direction of observation (up- or down-dip) do play a role. Yet, lower crustal earthquakes are rare in a continental (intraplate) environment. Mostly they occur in the upper crust. Rules-of-thumb for calculating the source distance from the travel-time differences Sg-Pg and Sn-Pn are:

$$\text{hypocenter distance } d \text{ [in km]} \approx \Delta t(\text{Sg-Pg}) \text{ [in s]} \times 8 \text{ (near range only)} \quad (2.35)$$

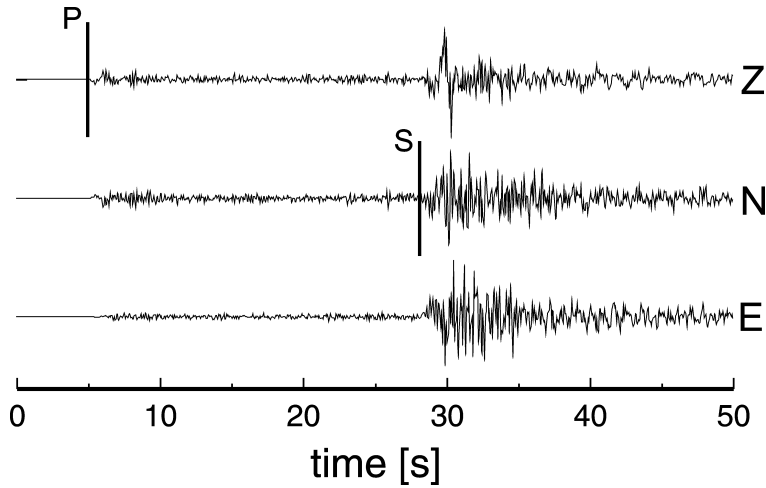
$$\text{epicenter distance } D \text{ [in km]} \approx \Delta t(\text{Sn-Pn}) \text{ [in s]} \times 10 \text{ (in Pn-Sn range } < 15^\circ) \quad (2.36)$$

Sometimes, very strong onsets after Pg, well before Sn or Sg are observed and may complicate proper interpretation of local phases on the basis of an oversimplified travel-time model as the one in Fig. 2.48. Then special modelling may be required, which is beyond the scope of seismic routine analysis. According to Bock et al. (1996) these strong onsets (?) may be related to local depth phases (e.g., sPmP in Fig. 2.49).



**Fig. 2.49** Two records of a local earthquake at a source depth of  $h = 6$  km in eastern Germany. The strong seismic phase ? between the onsets of Pg and Sg at hypocenter distance of 122 km in the record of station BRG can not be identified by using the simplified set of travel-time curves for surface foci in Fig. 2.48. According to model calculations it is most likely the depth phase sPmP of the large amplitude overcritical reflection PmP from the Moho discontinuity.

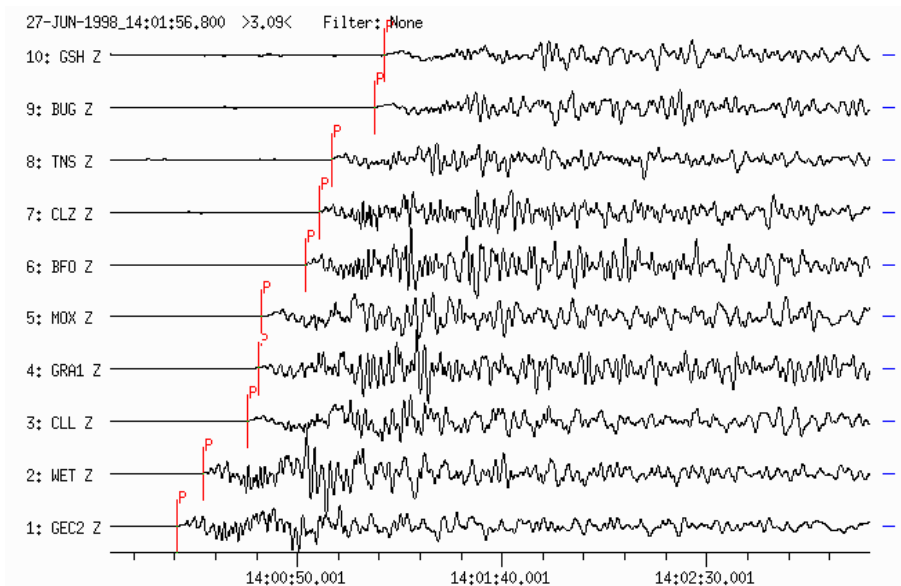
Also be aware that in the case of sub-crustal earthquakes, which are common in subduction zones, none of the crustal phases discussed above exist. In this case, the first arriving longitudinal and shear wave onsets are usually termed P and S, respectively, as for teleseismic events (see Fig. 2.50) although the IASPEI standard phase nomenclature terms them also Pn and Sn for upgoing rays from the source that are recorded in the local distance (see Figure 1c in IS 2.1; Storchak et al. 2003 and 2011).



**Fig. 2.50** P- and S-wave onsets from a local earthquake in northern Chile at a depth of 110 km and a hypocenter distance of about 240 km (courtesy of B. Schnurr, 2001).

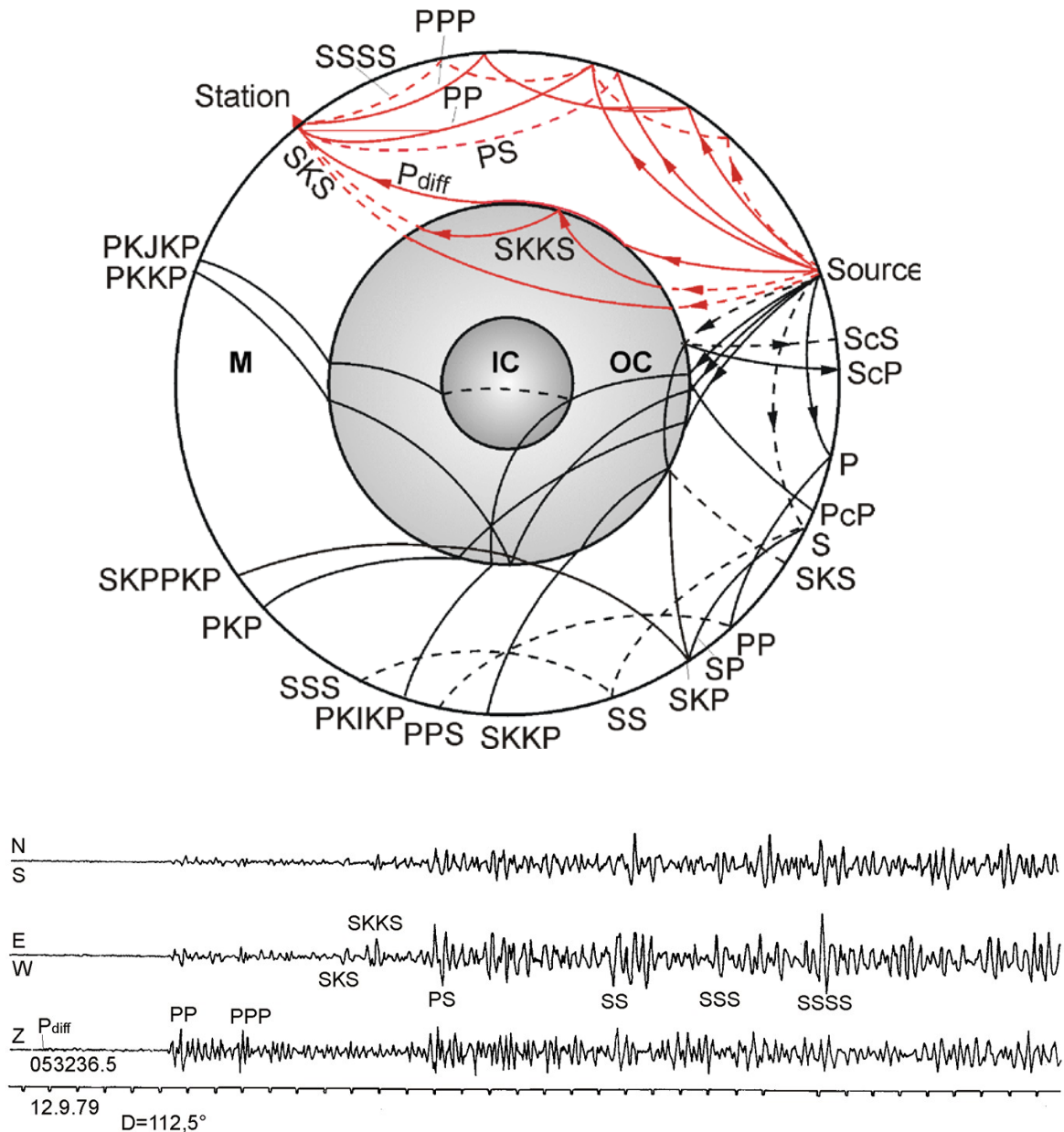
## 2.6.2 Shallow source body-waves and travel times at teleseismic distances

Seismic waves arriving at distances beyond  $10^\circ$  up to about  $30^\circ$  have mainly traveled through the upper mantle (from Moho to about 410 km depth) and the transition zone to the lower mantle (between about 410 km and 660 km depth). The strong discontinuities which mark the upper and lower boundary of the transition zone are associated with strong increases in seismic impedance (i.e., of both velocity and density; see Fig. 2.7). This results in two remarkable triplications of the travel-time curve for P waves (see Fig. 2.32) and S waves, which give rise to complicated waveforms of rather long duration (about 10 s and more) and consisting of a sequence of successive onsets with different amplitudes (Fig. 2.51).



**Fig. 2.51** Broadband seismograms with high time resolution showing the complex P-wave groups in records of an earthquake in southern Turkey at stations of the German Regional Seismograph Network (GRSN) at epicenter distances between  $D = 19.7^\circ$  (GEC2) and  $24.8^\circ$  (GSH).

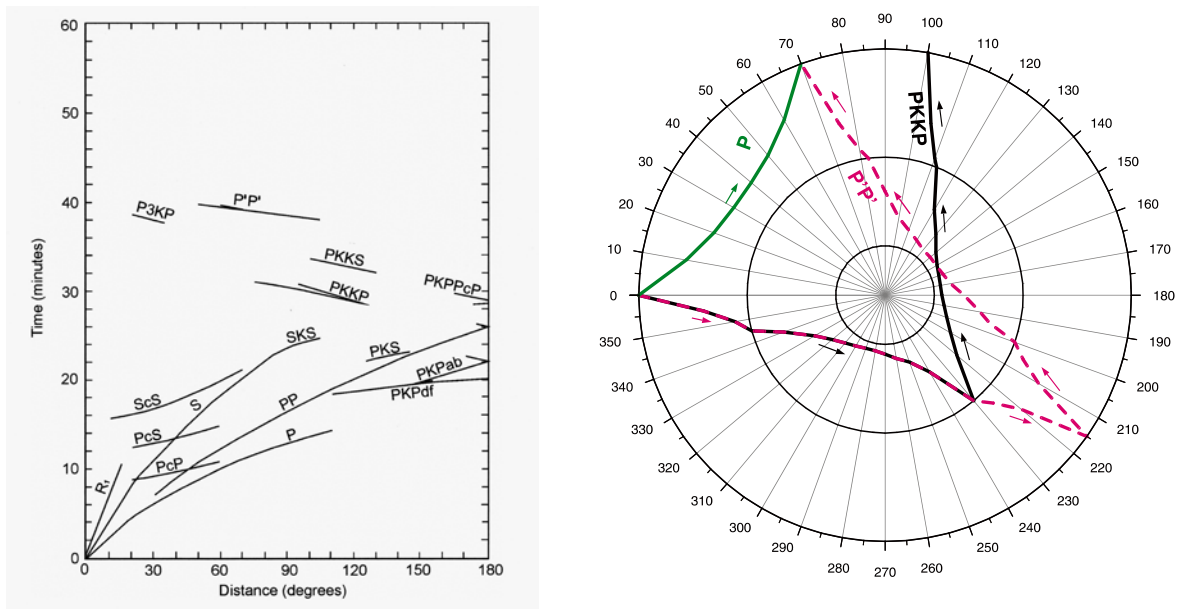
For epicenter distances  $D > 30^\circ$  P and S waves are followed by an increasing number of secondary waves, mainly phases, which have been reflected or converted at the surface of the Earth or at the core-mantle boundary(CMB). Fig. 2.52 depicts a typical collection of possible primary and secondary ray paths together with a three-component seismic record at a distance of  $D = 112.5^\circ$  that relates to the suit of seismic rays shown in red in the upper part of the cross section through the Earth. The phase names are standardized and in detail explained in IS 2.1



**Fig. 2.52 Top:** Seismic ray paths through the mantle (**M**), outer core (**OC**) and inner core (**IC**) of the Earth (above) with the respective phase symbols according to the international nomenclature (see IS 2.1, also for detailed ray tracing). Full lines: P rays; broken lines: S rays. Related travel-time curves are given in Fig. 2.60. Red rays relate to the seismic phases identified in the 3-component Kirnos SKD broadband seismogram recorded at station MOX, Germany (**bottom**) of body-waves from an earthquake at an epicenter distance of  $112.5^\circ$ .

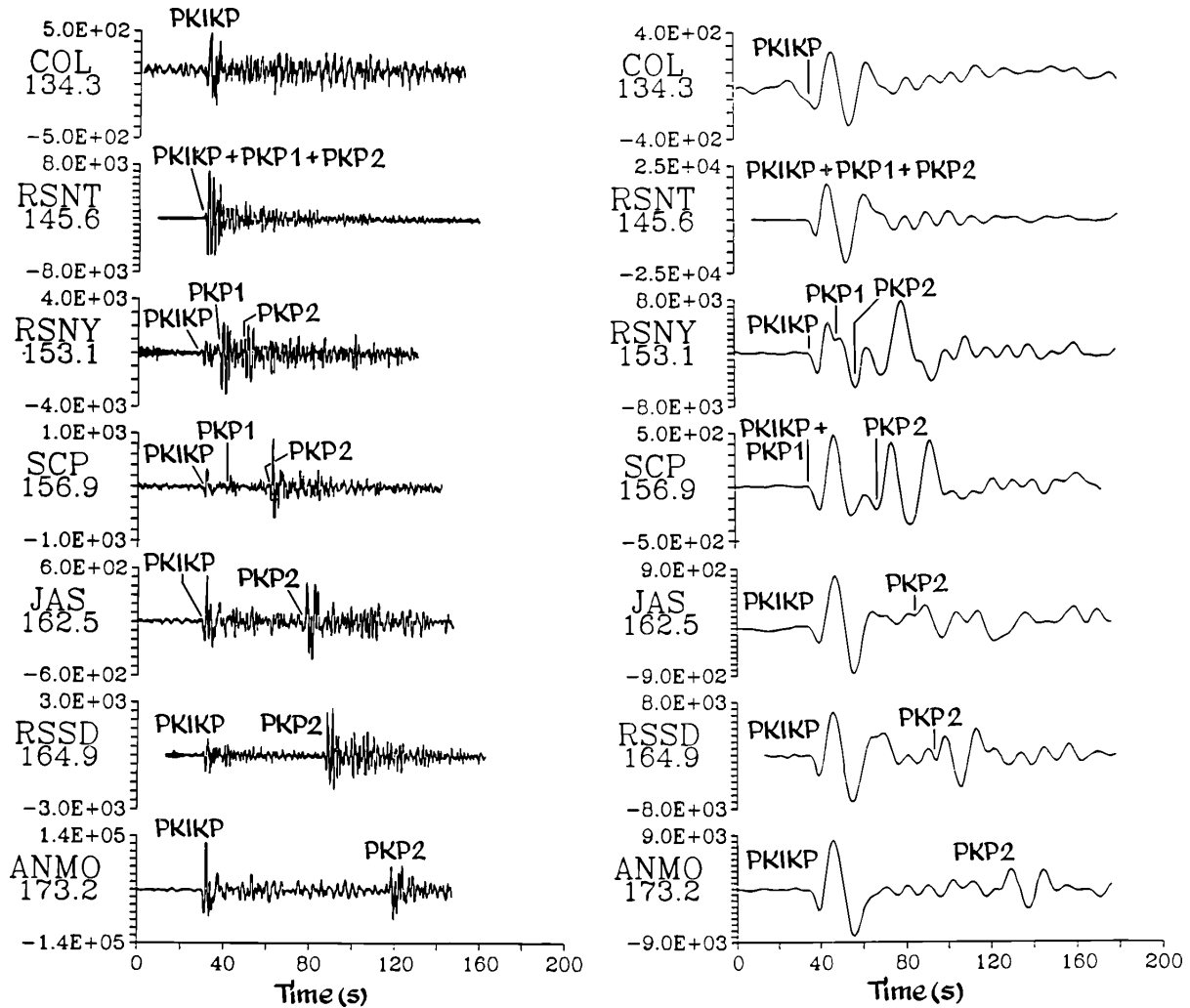
Between about  $30^\circ$  and  $100^\circ$  epicenter distance P and S have traveled through the lower mantle, which is characterized by a rather smooth positive velocity and density gradient (see Fig. 2.79). In this distance range, seismograms are relatively clearly structured with P and S (or beyond  $80^\circ$  with SKS) being the first, prominent longitudinal and transverse wave arrivals, respectively, followed by multiple surface and core-mantle boundary (CMB) reflections or conversions of P and S such as PP, PS, SS and PcP, ScP etc. (see travel-time curves in Fig. 2.60).

Within about 15 to 35 min after the first P arrival multiple reflections of PKP from the inner side of the CMB (PKKP; P3KP) or from the surface (PKPPKP = P'P') may be recognizable in short-period seismic records. Their travel-time curves have been plotted in the left-hand panel and their ray traces in the right-hand panel of Fig. 2.53. Many more of such multiple core phase records are presented in Chapter 11, section 11.5.3.



**Fig. 2.53 Left:** Short-period teleseismic travel-time curves and **right:** ray paths of PKKP and P'P' (= PKPPKP) with respect to the direct P phase (courtesy of S. Wendt, 2001).

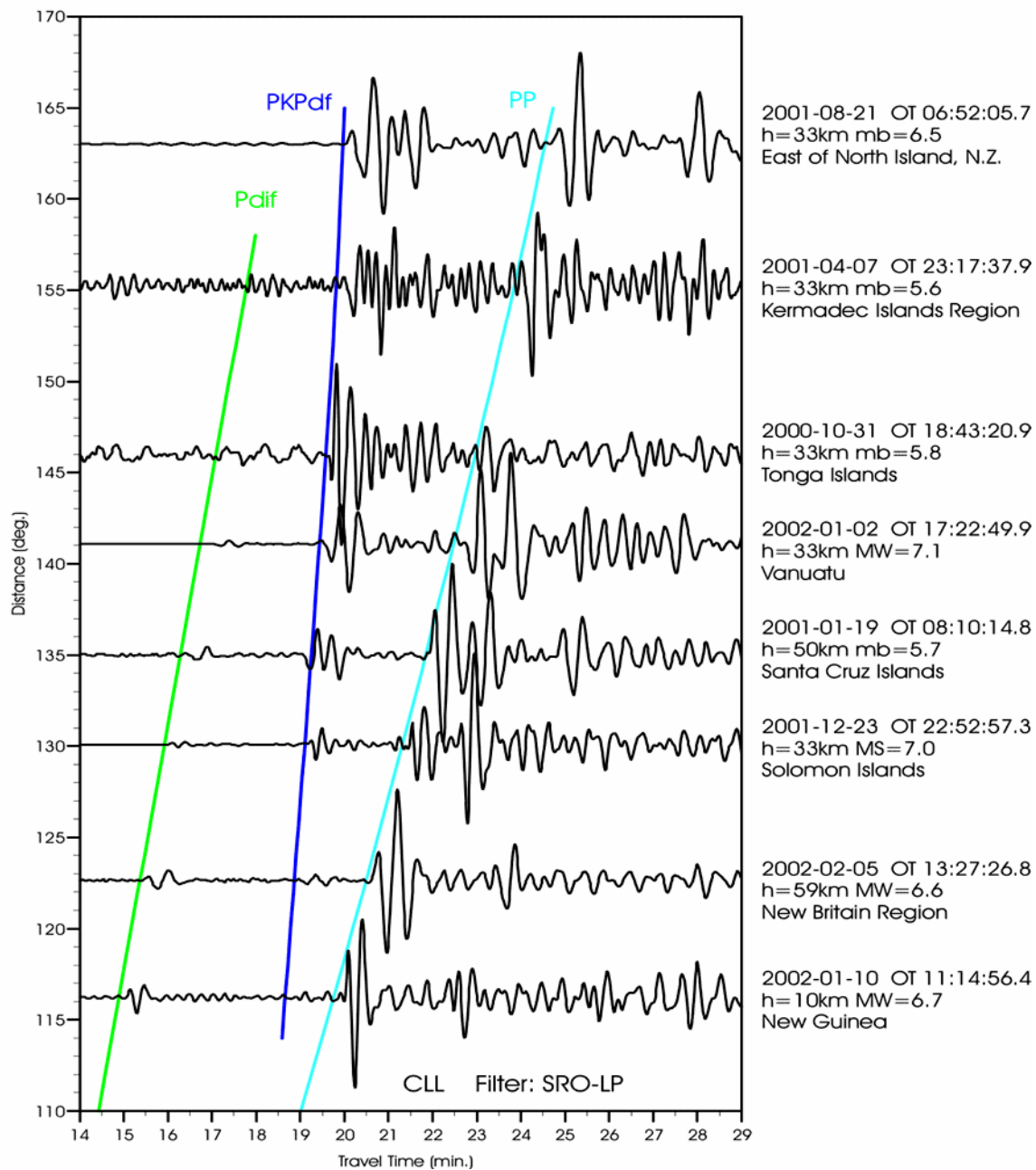
Beyond  $100^\circ$ , P-wave rays, which entered the outer core after strong downward refraction, will return with time delay to the surface as PKP core phases (Fig. 2.53 left). This is due to the strong reduction of the P-wave velocity at the CMB from about 13.7 km/s in the lowermost mantle to 8.0 km/s in the upper outer core. Thus, short-period P waves form a *core shadow* whereas PKP forms a caustic around  $145^\circ$  due to the slow velocity increase in the liquid outer core and the strong positive velocity gradient in the transition to the solid inner core (see 2.76). This results in strong amplitudes that are comparable with those of P at much shorter distances around  $50^\circ$  (see Chapter 3, Fig. 3.34). At the caustic of PKP three different branches of PKP meet (see Figs. 2.53 and 2.54 and Chapter 11, Figs. 11.72 and 11.73).



**Fig. 2.54** Short-period (left) and long-period (right) seismograms for the Mid-Indian Rise earthquake on May 16, 1985 ( $M = 6.0$ ,  $h = 10$  km) in the range  $D = 145.6^\circ$  to  $173.2^\circ$ . (From Kulháněk (1990), *Anatomy of Seismograms*, plate 55, pp. 165-166, © with permission from Elsevier Science). Note: The figure above gives still the old names of the core phases. According to the new IASPEI phase names PKP2 should be replaced by PKPab, PKP1 by PKPbc and PKIKP by PKPdf (see IS 2.1, also for the detailed ray tracing of these phases).

Note, however, that long-period P-wave energy is diffracted around the CMB into this shadow zone for short-period P-wave energy up to about  $150^\circ$  (Fig. 2.55). According to the new IASPEI nomenclature of phase names (see IS 2.1) the diffracted P wave is termed Pdif, however the old phase symbol Pdiff is still widely used. The amplitudes of Pdif are much smaller than those of PP, which is the strongest longitudinal arrival in the far teleseismic range (see Fig. 2.55 and Figs. 11.74 and 11.75 in Chapter 11).

In more detail, the types of seismic phases appearing at teleseismic distances and their peculiarities are discussed in Chapter 11, where many additional record examples are given, complemented by Datasheets DS 11.2 and 11.3.

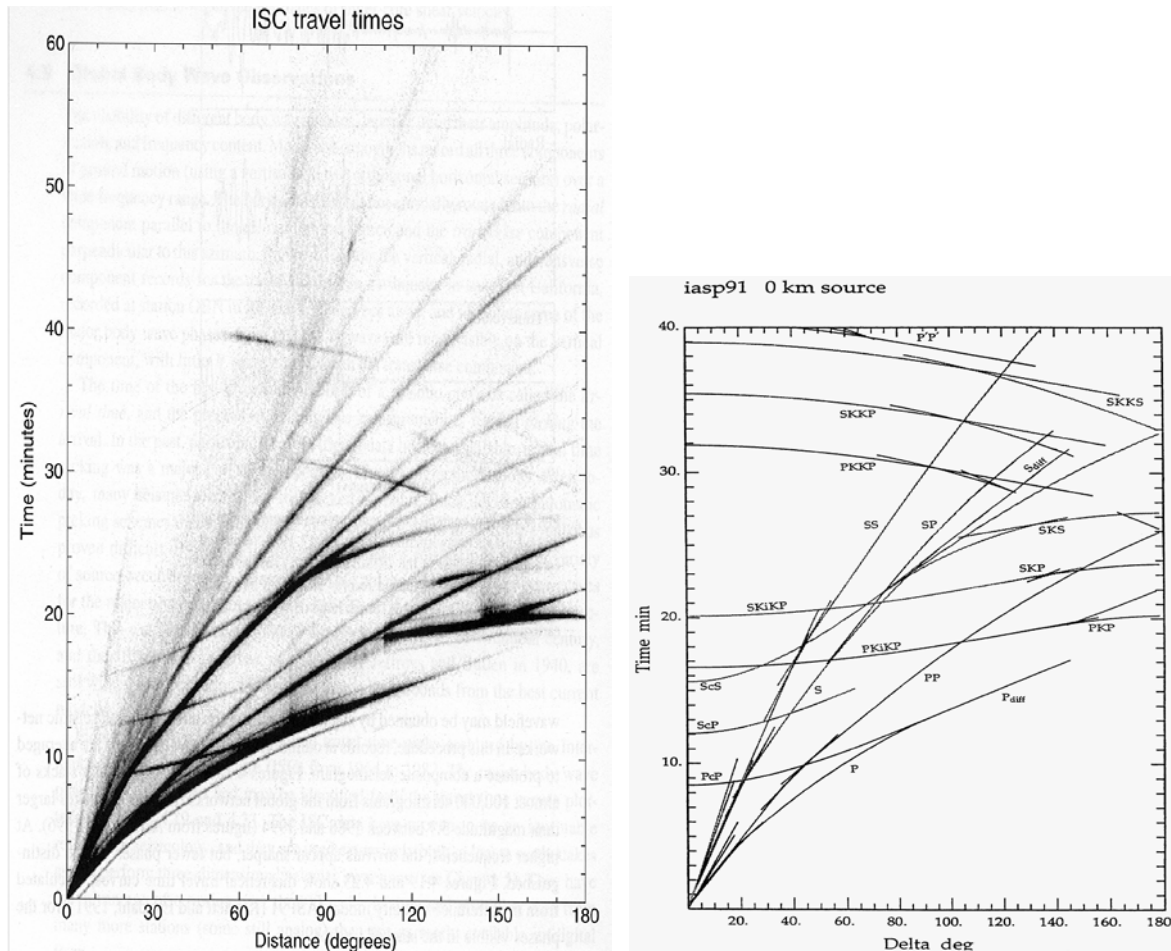


**Fig. 2.55** Long-period records of Pdif, PKPdf (with shortly after following PKPbc between 145° and 157° and PKPab (between 145 and 180°; all better discernable in the short-period records of Fig. 2.54), and PP (courtesy of S. Wendt, 2002).

The first discernable motion of a seismic phase in the record is called the *arrival time* and the measurement of it is termed *picking* of the arrival (see Chapter 11, section 11.2.2). Up to now, arrival time picking and reporting to international data centers is one of the major operations of data analysts or automatic analysis systems at seismic stations or network centers. By plotting the time differences between reported arrival times and calculated origin times over the epicenter distance, seismologists were able to construct travel-time curves for the major phases and to use them to infer the average radial velocity structure of the Earth (see 2.7).

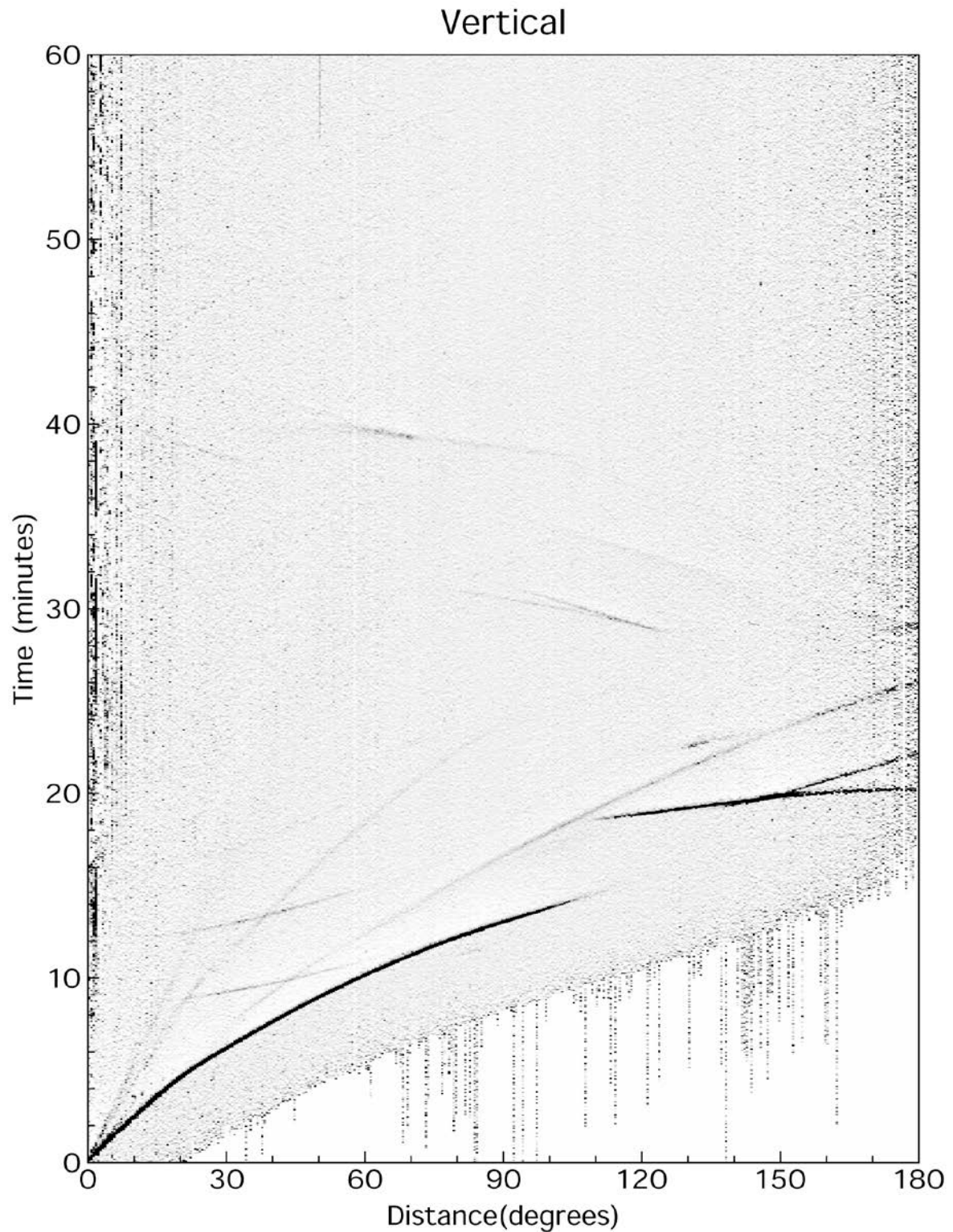


In Fig. 2.56 (left) more than five million travel-time picks, archived by the International Seismological Centre (ISC) for the time 1964 to 1987, have been plotted. Most time picks align nicely to travel-time curves, which match well with the travel-time curves theoretically calculated for major seismic phases on the basis of the IASP91 model (Fig. 2.56 right).



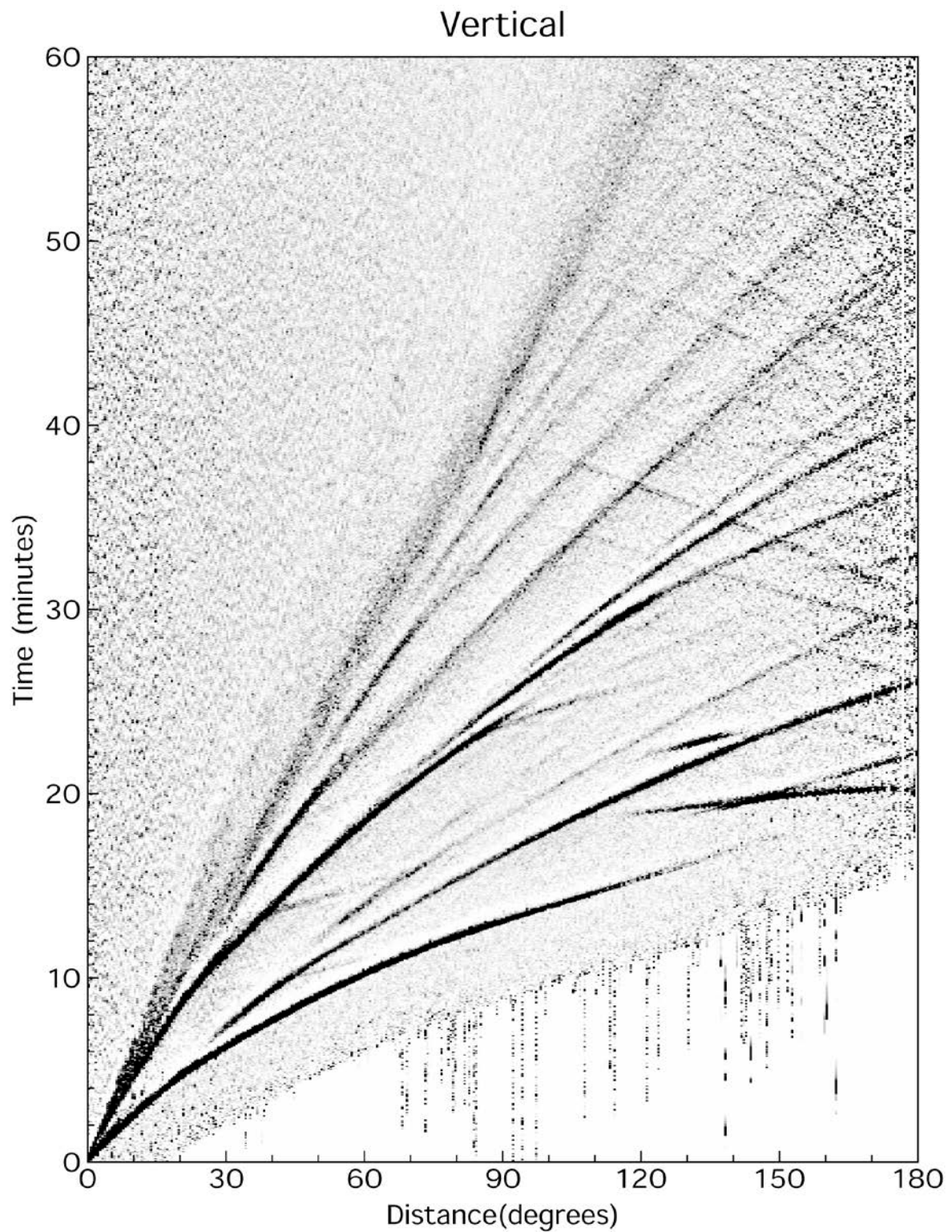
**Fig. 2.56 Left:** Travel-time picks collected by the ISC between 1964 and 1987 for events shallower than 50 km. (From Shearer, *Introduction to Seismology*, 1999; © with permission from Cambridge University Press). **Right:** IASP91 travel-time curves for surface focus (from Kennett, 1991).

An even more complete picture of the entire seismic wavefield may nowadays be obtained by stacking data from global digital seismic networks. Records at common source-receiver ranges have been averaged to produce a composite seismogram. Stacks of almost 100,000 seismograms from the global digital networks have been plotted by Astiz et al. (1996; Fig. 2.57 for short-period records with periods  $T < 2$  s and Fig. 2.58 for long-period records with  $T > 10$  s). Although the arrivals appear sharper at higher frequencies and late arriving reflected core phases P'P' (PKPPKP), PKKP as well as higher multiples of them are discernable in short-period records only, in total much less later phases can be distinguished in short-period records. The darker the dot “curves” appear against the gray background the larger is the relative frequency with which these phases can be observed above the noise level. Figs. 2.59 and 2.60 present the calculated travel-time curves for the IASP91 velocity model (Kennett and Engdahl, 1991) together with the phase nomenclature. The theoretical travel-time curves match excellent with the difference between automatically picked onset times and origin time.

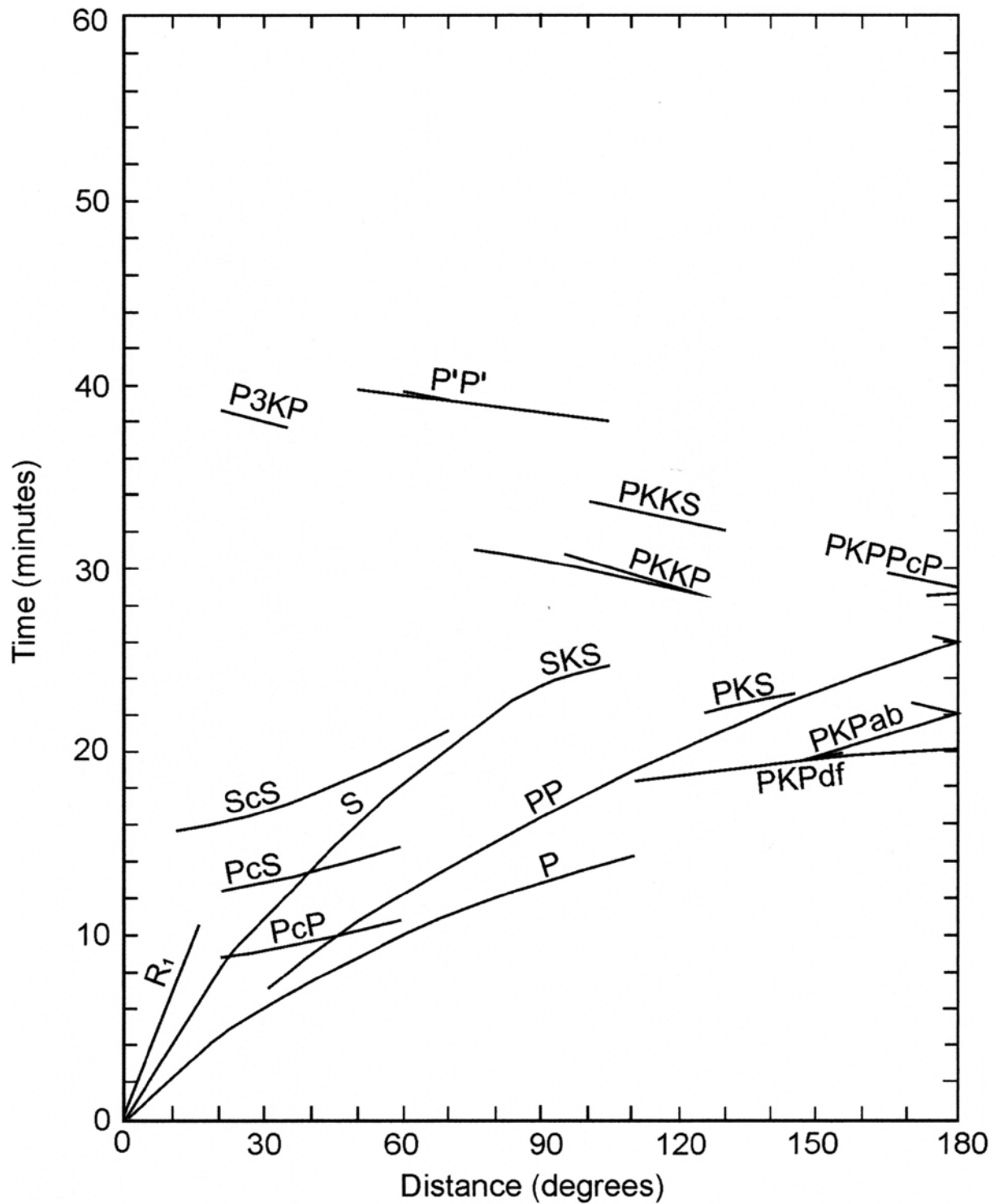


**Fig. 2.57** Stack of short-period filtered ( $T < 2$  s), vertical component data from the global networks between 1988 and 1994 (from Astiz et al., Global Stacking of Broadband Seismograms, *Seismological Research Letters*, Vol. 67, No. 4, p. 12, © 1996; with permission of Seismological Society of America).

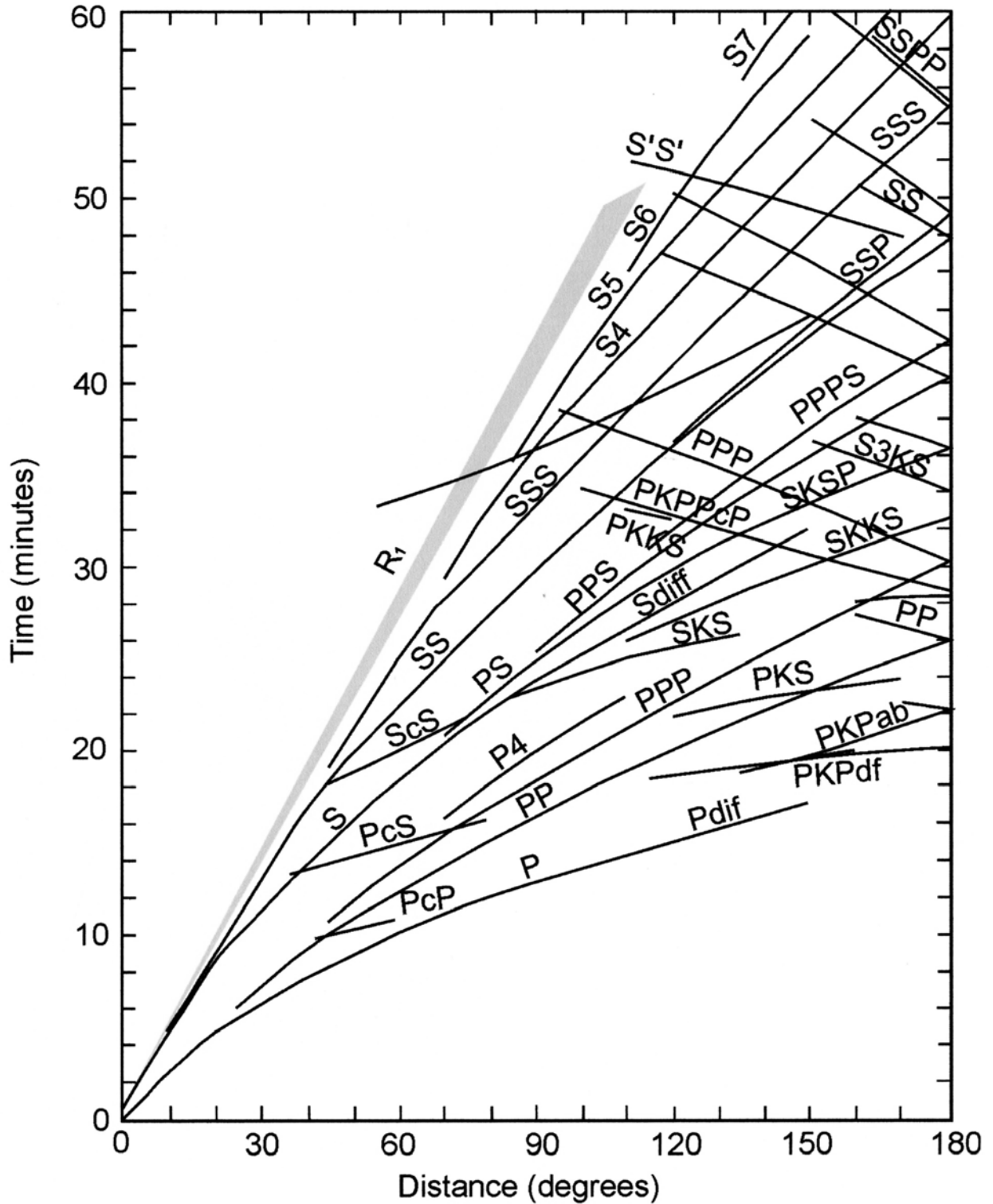




**Fig. 2.58** Stack of long-period ( $T > 10$  s), vertical component data from the global networks between 1988 and 1994. (from Astiz et al., Global Stacking of Broadband Seismograms, *Seismological Research Letters*, Vol. 67, No. 4, p. 14, © 1996; with permission of Seismological Society of America).

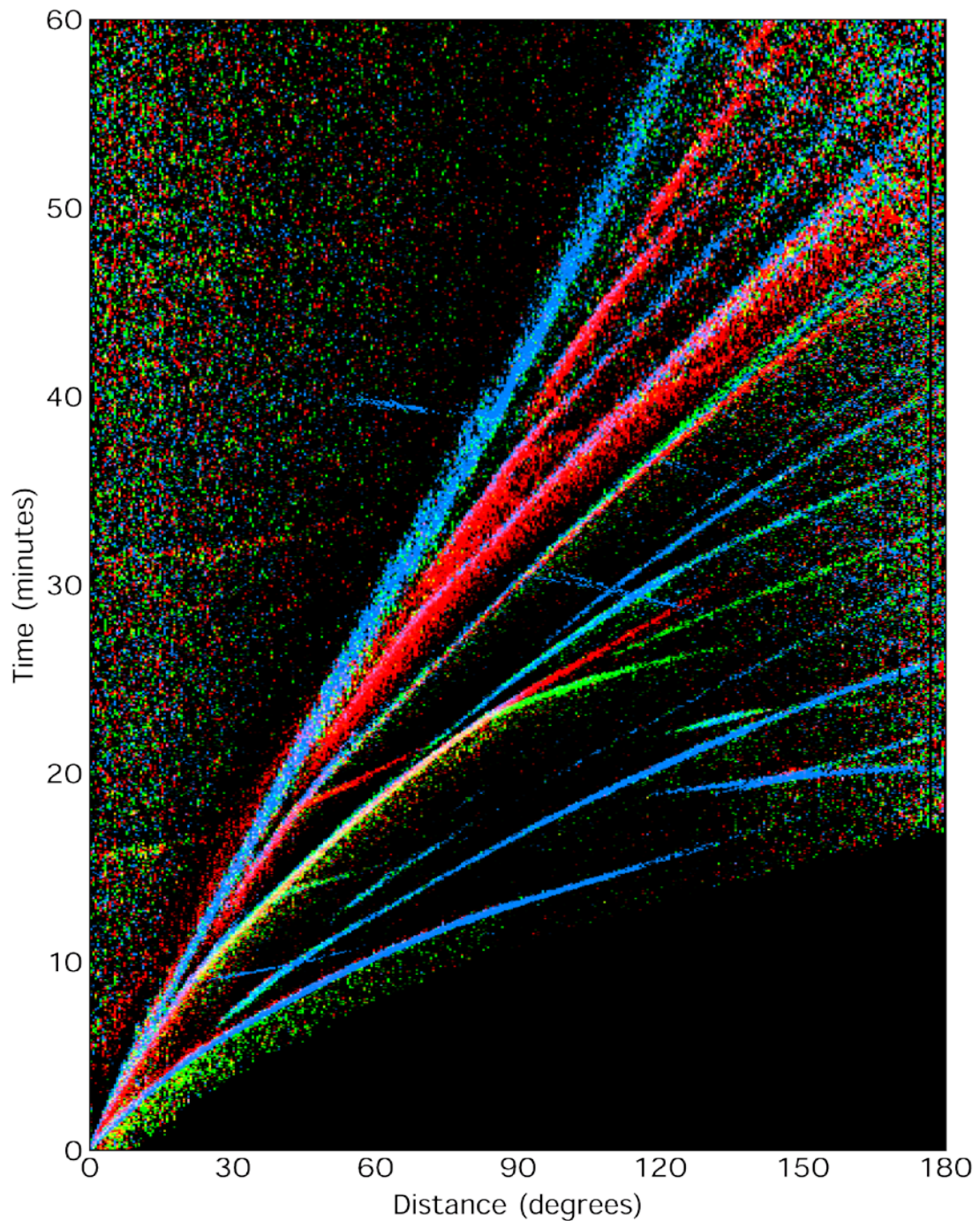


**Fig. 2.59** Overlay to Fig. 2.57 with the theoretical IASP91 travel-time curves for surface foci and the related phase nomenclature. Produce a 1:1 copy on a transparent sheet and match it with Fig. 2.57.



**Fig. 2.60** Overlay to Fig. 2.58 with the theoretical IASP91 travel-time curves for surface foci and the related phase nomenclature. Produce a 1:1 copy on a transparent sheet and match it with Fig. 2.58.

Complementary to Figs. 2.57 and 2.58, Fig. 2.61 reveals the different polarization of the various seismic phases.



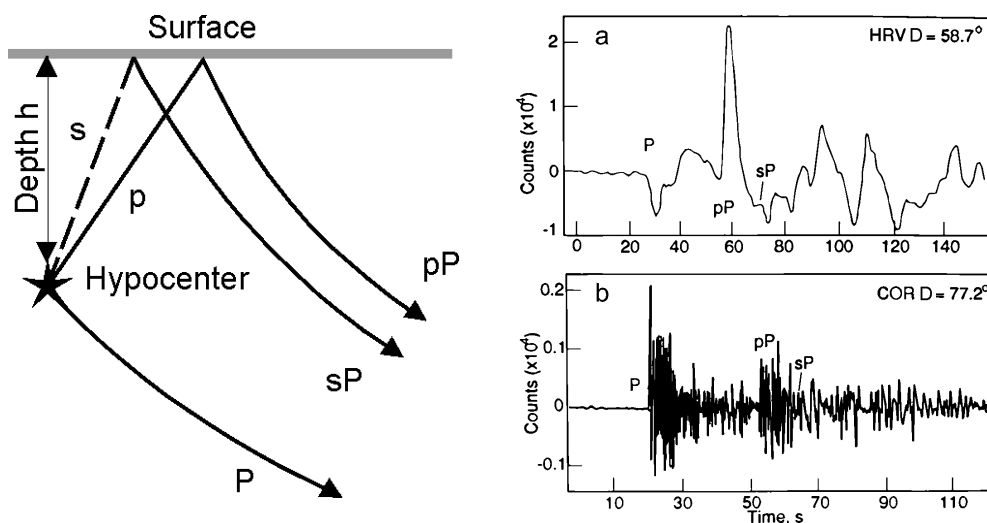
**Fig. 2.61** Global travel-time curves for shallow earthquakes as produced by stacking broadband seismograms. Seismic phases are shown in different colors depending on their polarization (**blue**: vertical motion; **green**: radial-horizontal; and **red**: transverse-horizontal) (courtesy of L. Astiz).

While all primary longitudinal phases and all from P or K to S converted phases, and vice versa, appear on vertical and radial-horizontal components only, multiple reflected S waves, which lose with each reflection more and more of their SV polarized energy due to conversion into P (or K at the CMB), become more and more transversely polarized. Primary

S, however, has significant energy on both horizontal components that are oriented either parallel to the backazimuth to the source (radial) or perpendicular to it (transverse). Direct P waves, polarized in the direction of ray propagation, have in the teleseismic range dominating vertical components because of their steep incidence angle, which gets smaller and smaller with increasing distance (see e.g., PKP phases). PP, P3 and higher multiples may, however, have significant energy in the radial component too. These examples illustrate that the visibility and discrimination of body wave phases in seismic records depends on their relative amplitude, polarization and frequency content. All of these criteria have to be taken into account, besides the differences in travel-times, when analyzing seismic records.

### 2.6.3 Depth phases

In the case of distant deep earthquakes the direct P wave that leaves the source downward will arrive at a teleseismic station first. It will be followed, depending on the source depth, up to about 4.5 min later, by other phases that have left the source upward (Fig. 2.62). These phases, also when associated with PP, S, SS and reflected or converted at the free surface of the Earth or at the ocean bottom (such as pP, sP, ' , sPP, pPKP, etc.), at the free surface of the ocean (e.g., pwP, swP), or from the inner side of the Moho (e.g., pm-P) are called *depth phases*. Their proper identification, onset-time picking and reporting is of crucial importance for reliable determination of source depth (see below and section 6.1 with Figure 7 in the Information Sheet IS 11.1). Differential travel-time tables for pP-P and sP-P are presented in Exercise EX 11.2. For the definition of these phases see also IS 2.1.

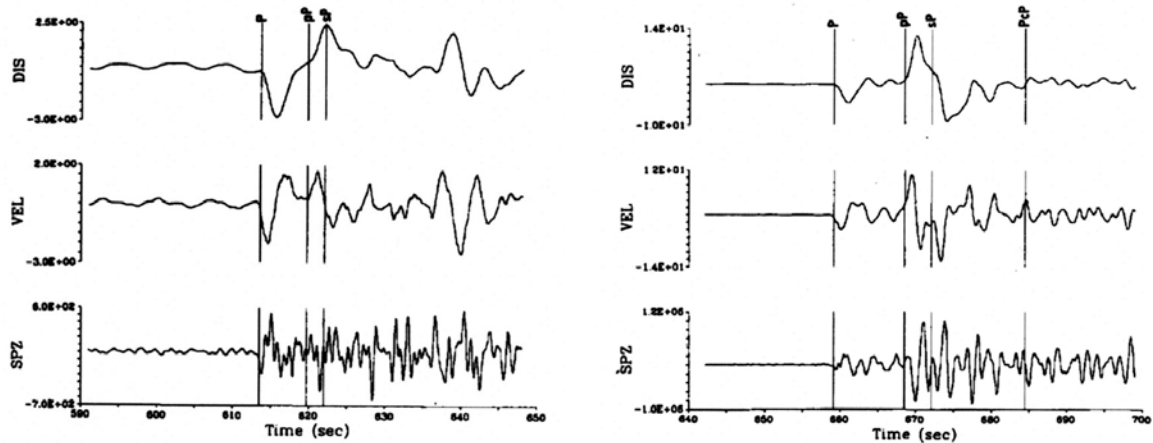


**Fig. 2.62 Left:** Different ray paths of a direct teleseismic P wave and of its depth phases. **Right:** Records of depth phases of the May 24, 1991 Peru earthquake (hypocenter depth  $h = 127$  km); a) broadband record and b) simulated short-period recording (the right figure is a corrected cutout of Fig. 6.4 of Lay and Wallace, *Modern Global Seismology*, p. 205, © 1995; with permission of Elsevier Science, USA).

Depth phases of P, PP, S, SS or of converted P or S phases have commonly smaller amplitudes because of reflection and/or conversion losses as well as extra attenuation along the additional segment of travel. For shallow earthquakes a clear discrimination between

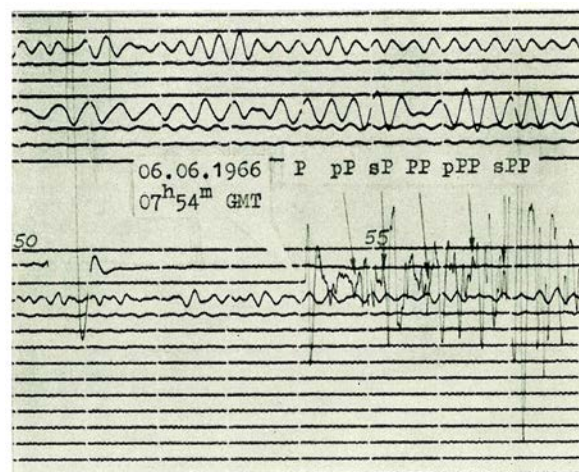


direct and depth phases is difficult and usually beyond the possibilities of routine seismogram analysis because of the closeness in time of the subsequent phase onsets and the superposition of their waveforms. They may, however, be discriminated by waveform modeling and matching for variable source depth and source mechanism (see subchapter 2.8, Fig. 2.85). Yet, in any event, for shallow earthquakes, short-period seismic records that emphasize on periods  $< 2\text{--}3\text{ s}$  are more suitable for identifying and time-picking of closely spaced depth phase onsets than broadband displacement or (somewhat better) velocity records (Fig. 2.63).



**Fig. 2.63** Appearance of P, pP and sP in displacement (DIS) and velocity (VEL) broadband as well as short-period vertical component records. The source depths of the earthquakes have been 18 km (left) and 40 km (right), respectively. Copied from Figures 1 and 5 in Choy and Engdahl (1987), *Phys. Earth Planet. Int.* **47** (© Elsevier Science Publ., 2012).

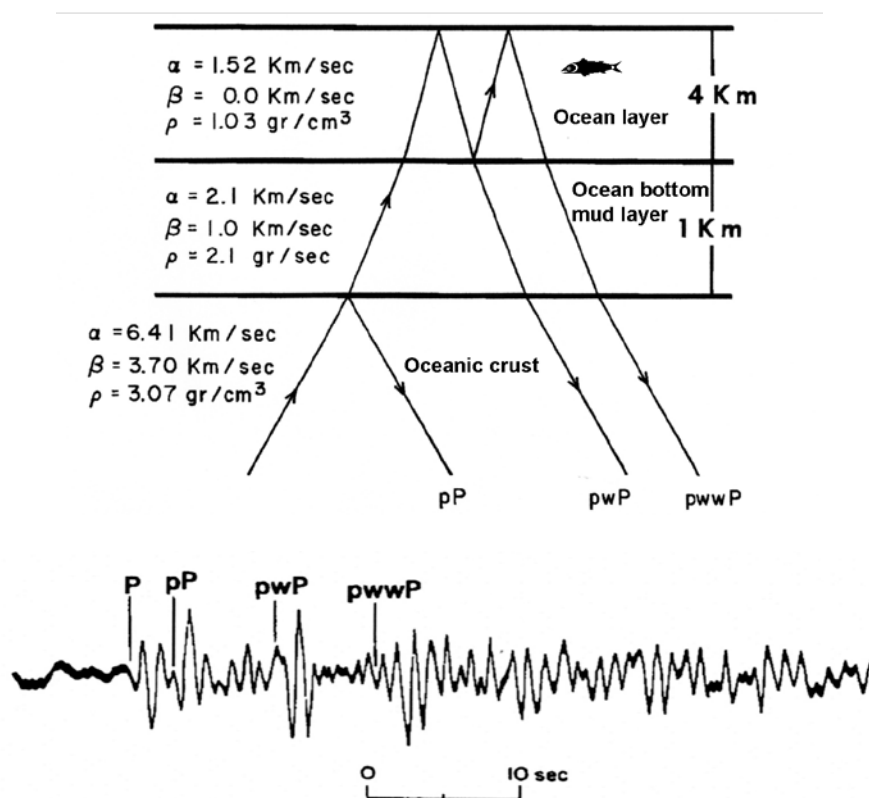
Yet, depth phases may sometimes also be larger than their primary phases, as confirmed by both observations (e.g., Fig. 2.62 upper right, Fig. 2.63 right and 2.64 below) and synthetic modeling (e.g., Langston and Helmberger, 1975).



**Fig. 2.64** Medium-period displacement-proportional record of P, PP and their depth phases by a Kirnos SKD seismograph ( $T_s = 20\text{ s}$ ;  $T_g = 1.25\text{ s}$ ;  $V_{\max} = 1000$  times) at station MOX, Germany, of an earthquake in the Afghanistan-USSR border region on 06 June 1966 ( $D = 44.3^\circ$ ,  $h = 220\text{ km}$ ). Copy of Fig. 7 in Bormann (1969).

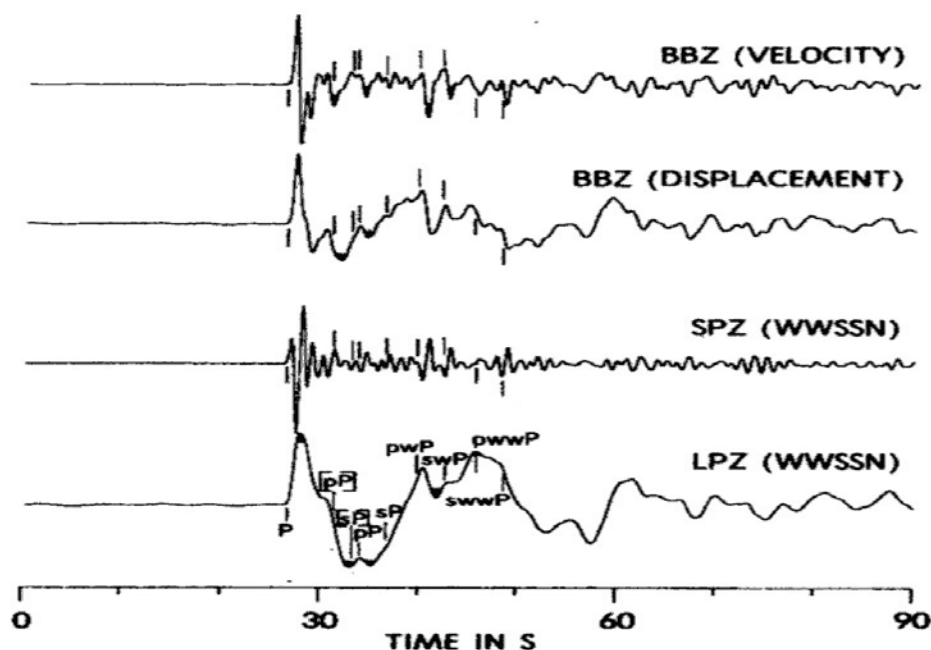
The great variability of amplitudes of both P and PP (also S and SS) and of their depth phases is a source-mechanism dependent function of take-off angle and azimuth and thus varies with distance and azimuth of the recording station in respect to the source radiation pattern.

Although the inclusion of correctly identified depth phases into the location procedure (besides S, PKiKP, PKPdf, PcP, ScP, if available) may significantly improve hypocenter location their proper identification becomes even more difficult if the reflection/conversion points of the up-going p or s rays are underneath an ocean with significant water depth. In this case significant reflection/conversion may occur both at the ocean bottom and at the ocean surface. While the former would be the common pP or sP phase, the latter requires a specifying phase name, namely pwP or swP (w = water). There maybe even double reflections within the water column, making, e.g., for a pwwP (see Fig. 2.65).



**Fig. 2.65 Top:** Average crust-ocean layer velocity model with seismic rays of P and its depth phases (amended after Mendiguren, 1971). **Bottom:** A short-period record example of these phases according to Engdahl and Billington (1986).

These ocean depth phases have first been described by Mendiguren (1971) and later in more detail by Yoshiu (1979) and Engdahl and Billington (1986). Since the water layer depth phases follow the primary depth phases usually within about 10-15 s, their waveforms are better discernable and their onset times more reliable to pick on short-period or velocity broadband records than on long-period or displacement broadband records (Fig. 2.66).



**Fig. 2.66** P and the depth-phase reflections from the oceanic sediment-basement interface ([pP] and [sP]), from the sediment-water interface (pP and sP), from the water surface (pwP and swP), as well as the multiples of the latter, once again reflected within the water layer (pwwP and swwP) (see Fig. 2.65 top), recorded with different instruments. Vertical bars mark the model-dependent theoretically calculated onset times of these phases. The figure was modified for better time-resolution from Figure 3b in Engdahl and Kind (1986), *Annales Geophysicae*, 4, (© European Geophysical Society, 2012).

According to the record examples shown, the water-layer depth phases may be even larger than the primary depth phases. Misinterpretations are therefore rather common. About one-third of the ISC identifications as pP, sP and PcP have been re-identified in a special analysis carried out by Engdahl et al. (1998). According to these authors quite many ISC pP, sP and partially also PcP phases have in fact been pwP and many sP onsets had been misinterpreted as pP. If a pwP phase produced by a water layer 4 km thick is misinterpreted as pP this may result (in the absence of local or regional P phases) in an overestimation of the focal depth by about 50 km. Therefore, Engdahl et al. (1998) developed an automatic phase identification algorithm, termed EHB for short, which aims at significantly improved hypocenter relocations. EHB is applied to the phase group which immediately follows the P phase at teleseismic distances, including pP, pwP, sP and PcP and allows the unambiguous identification of later phases such as S. In conjunction with there publication the authors relocated some 100.000 earthquakes. The EHB bulletin is now permanently updated and complemented and posted on the ISC website <http://www.isc.ac.uk/>.

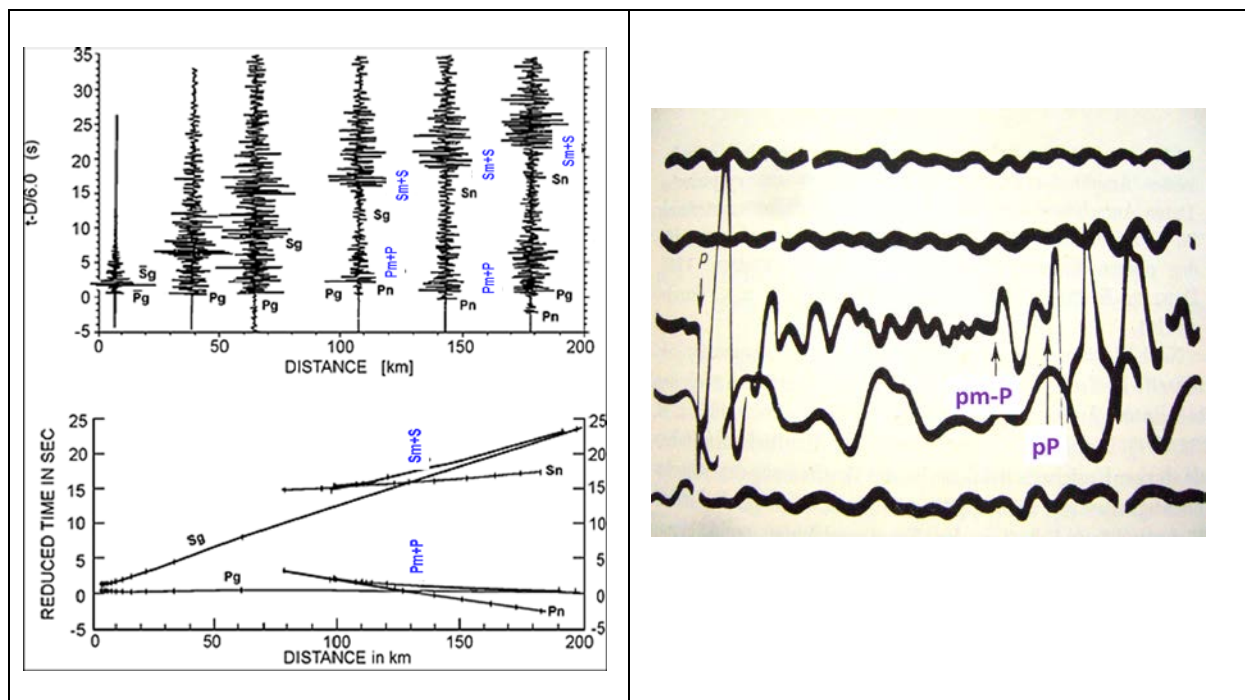
#### 2.6.4 $P_z \pm P$ , $S_z \pm S$ , $P_z \pm S$ , $S_z \pm P$ , etc.

Since the *impedance contrasts* at the free Earth surface, the ocean bottom and the core-mantle boundary are larger than at any other velocity-density discontinuity in the Earth, reflected and converted P and S phases from these discontinuities dominate the secondary body-wave onsets in seismograms. None the less, there are several other globally represented



discontinuities of significant contrast which give rise to additional reflections and conversions of P and S phases which may appear even in unprocessed seismograms as recognizable fore- or after-runner onsets to the primary phases P and S and thus complicate for analysts the proper phase identification and onset picking. The most important discontinuities in this regard are the Moho as well as the discontinuities bordering the upper mantle transition zone at about 410 km and 660 km depth. According to IS 2.1 as well as Storchak et al. (2003 and 2011) the unified nomenclature for such phases is:

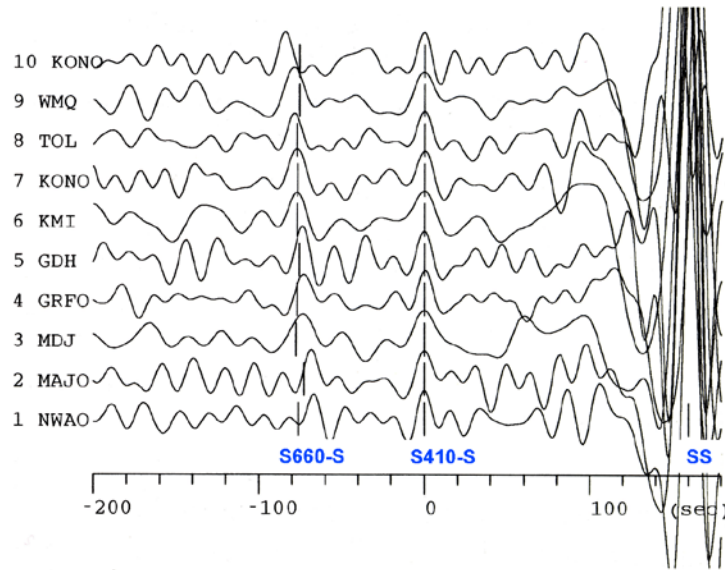
- **Pz+P or Sz+S:** P or reflection from the **outer** (top) side of a discontinuity at depth  $z$  outside of the core;  $z$  may be a positive numerical value in km, e.g., P660+P or S660+S, which are then reflections of P or S from the top of the 660 km discontinuity, or  $z$  may be a letter, e.g., m for Moho (see Pm+P and the bottom side reflection pm-P in Fig. 2.67).



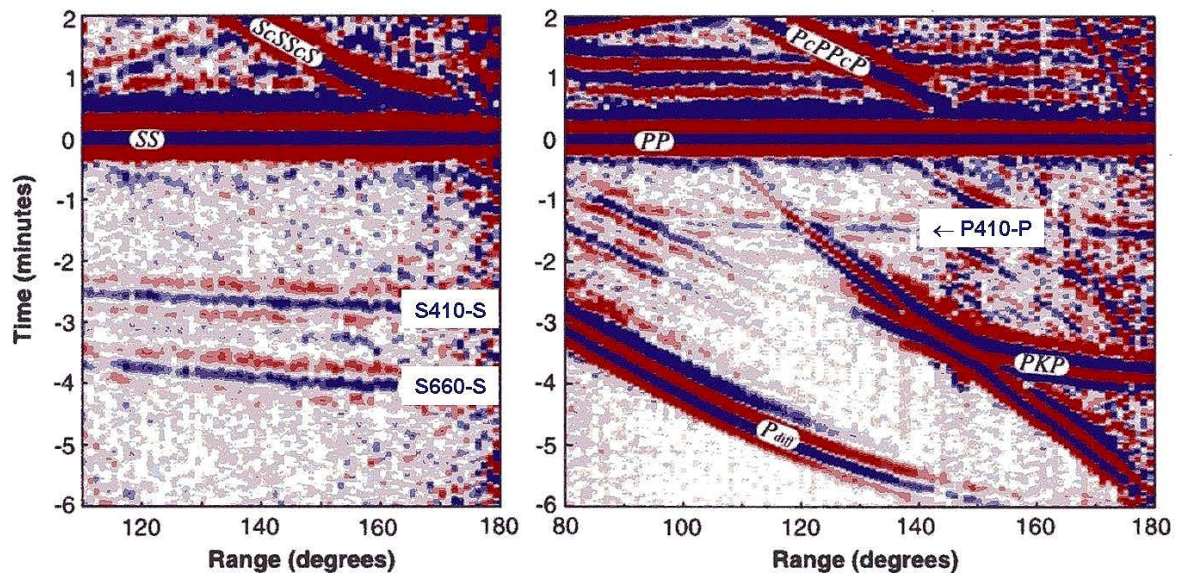
**Fig. 2.67 Left:** Travel-time curves and records of critically reflected crustal P and S waves from the upper side of the Moho: Pm+P and Sm+S, respectively. **Note:** The +sign is not obligatory to identify the phase as on upward reflection from the upper side of a discontinuity. Writing PmP and SmS instead is acceptable. See IS 2.1. Modified from Fig. 2.47;

**Right:** Depth phase pP from an earthquake at the Tyrrhenian Sea (13.04.1938,  $h = 256$  km) recorded at the station Sverdlovsk, Russia ( $D=33^\circ$ ), together with its smaller forerunner phase pm-P which is the downward reflection of the upcoming ray p on the inner (bottom) side of the Moho (see next bullet point). Modified Figure 108 in Savarenski and Kirnos (1960).

- **Pz-P or Sz-S:** P or S reflection from the **inner** (bottom) side of a discontinuity at depth  $z$  outside of the core, e.g., at the Moho (see Fig. 2.67 right) or in the Earth upper mantle such as P660-P or S410-S which are the P or S reflections from below the 660 km and 410 km discontinuity, respectively, and **precursory to PP or SS** (Fig. 2.68 and 2.69). **Note:** Writing the -sign is in these cases obligatory to identify them as a downward reflections from the bottom side of the discontinuity.



**Fig. 2.68** Aligned long-period records of SS phases (very large amplitudes on the outermost right) and of their much weaker precursors reflected from the inner (bottom) side of the discontinuities in the mantle at 410 km (**S410-S**) and 660 km (**S660-S**) depth from different earthquakes recorded at different seismic stations and distances: Trace 1: 1983 Oct 17 19:36:21.5 ; Trace 2: 1980 Jan 27 16:38:01.1; Trace 3: 1988 Feb 06 18:03:54.8; Trace 4: 1980 Oct 25 11:00:05.1; Trace 5: 1990 Feb 19 05:34:37.4; Trace 6: 1990 Dec 17 11:00:19.6; Trace 7: 1989 Apr 18 12:33:54.8; Trace 8: 1984 May 26 22:42:47.3; Trace 9: 1989 Dec 03 14:16:48.8; Trace 10: 1986 Jun 28 05:03:47.5. The vertical bars mark the theoretical travel times with respect to S410-S according to the global IASP91 model. Modified after Gossler and Kind (1996).

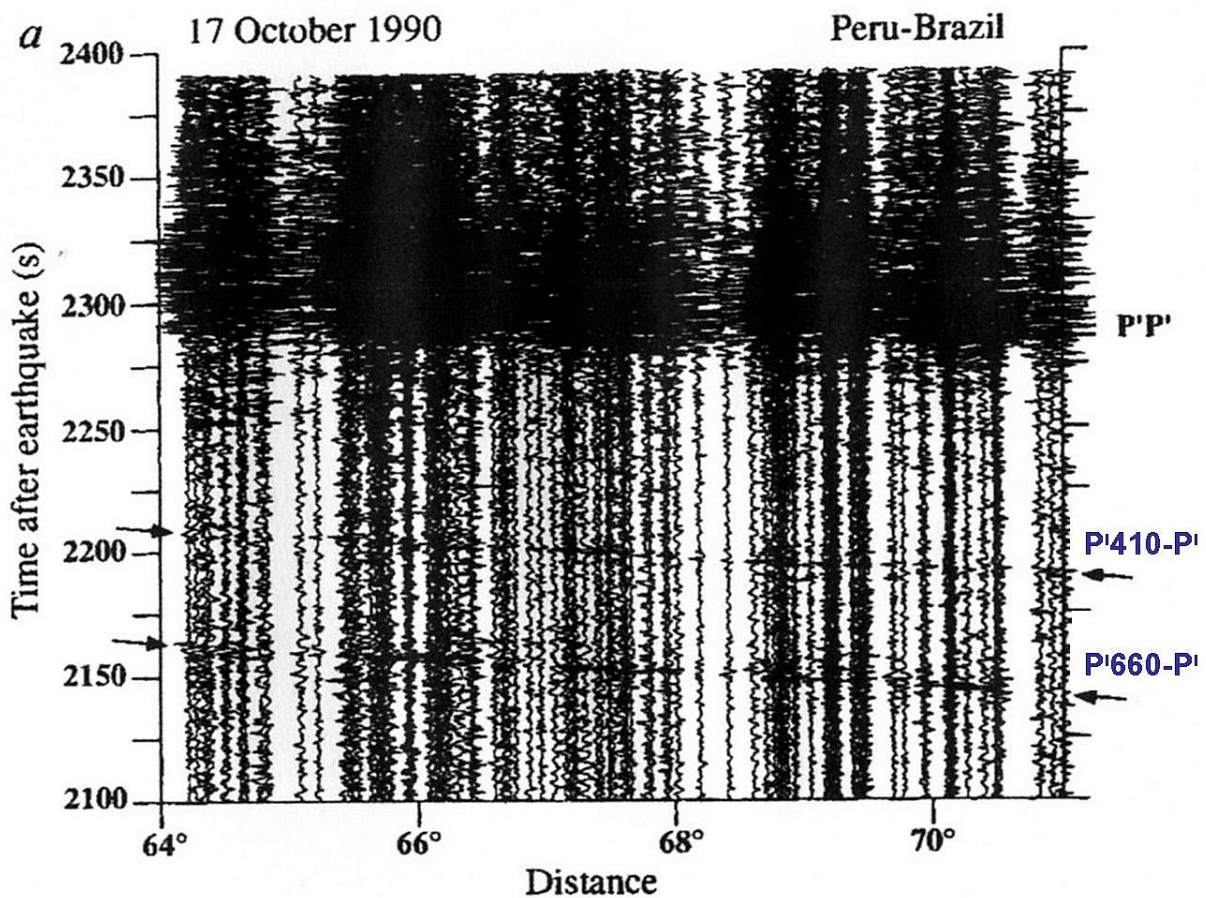


**Fig. 2.69 Left:** Similar to Fig. 2.68 but stacked and color-coded record section in a wide distance range, and aligned with respect to the arrival time of the free surface reflection of SS. Clearly visible are the fainter precursory arrivals of S660-S and S410-S, which arrive about 4 and 2.5 min ahead of the SS. **Right:** The same for the still weaker P410-P with respect to PP. Modified from Figure 1 in Shearer and Flanagan (1999), *Science*, **285**, p. 1546, in order to agree with IASPEI standard phase nomenclature (© Science).



- Accordingly,  $P_z+S$  and  $S_z+P$  or  $P_z-S$  and  $S_z-P$ , respectively, are P or S waves that have been converted into S or P when reflected from the outer (top) or inner (bottom) side, respectively, of a discontinuity at depth  $z$ .
- In an analog way the above notations apply also to  $PKP = P^I$  and  $SKS = S^I$  phases, such as  $P^I_z-P^I$  being a PKP phase reflected from the inner (top) side of a discontinuity at depth  $z$  outside the core, thus being precursory to  $P^I = PKP$ . A fine example is given in Fig. 2.70.

Usually, however, the SNR of such phases is too small to be recognized and picked correctly in routine seismogram analysis.



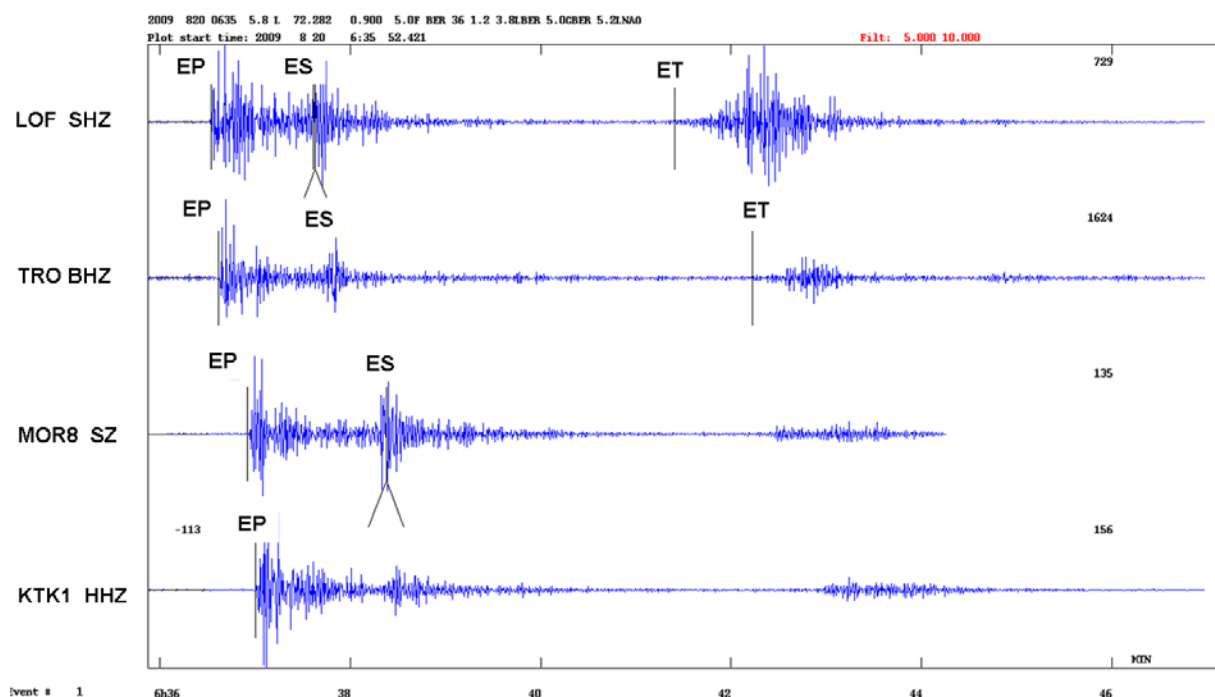
**Fig. 2.70**  $P^I P^I$  record section for a South American earthquake recorded by the California seismic array showing weak but distinct precursors  $P^I P^I$  and  $P^I P^I$ . Modified from Fig. 2a of Benz and Vidale (1993), *Nature*, **365**, p. 148, in order to agree with IASPEI standard phase nomenclature. (© Copyright Clearing Centre of Nature).

### 2.6.5 T phases

The letter *T* stands for *Tertiary wave*. This term has been introduced into seismology by Linehan (1940) in order to discriminate these waves from the much earlier arriving P (primary) and S (secondary) waves radiated by the seismic source that have travelled all their way through the Earth. T waves are generated by earthquakes, explosions or volcanic eruptions in or near the oceans, propagate in the oceans as an acoustic wave guided in the

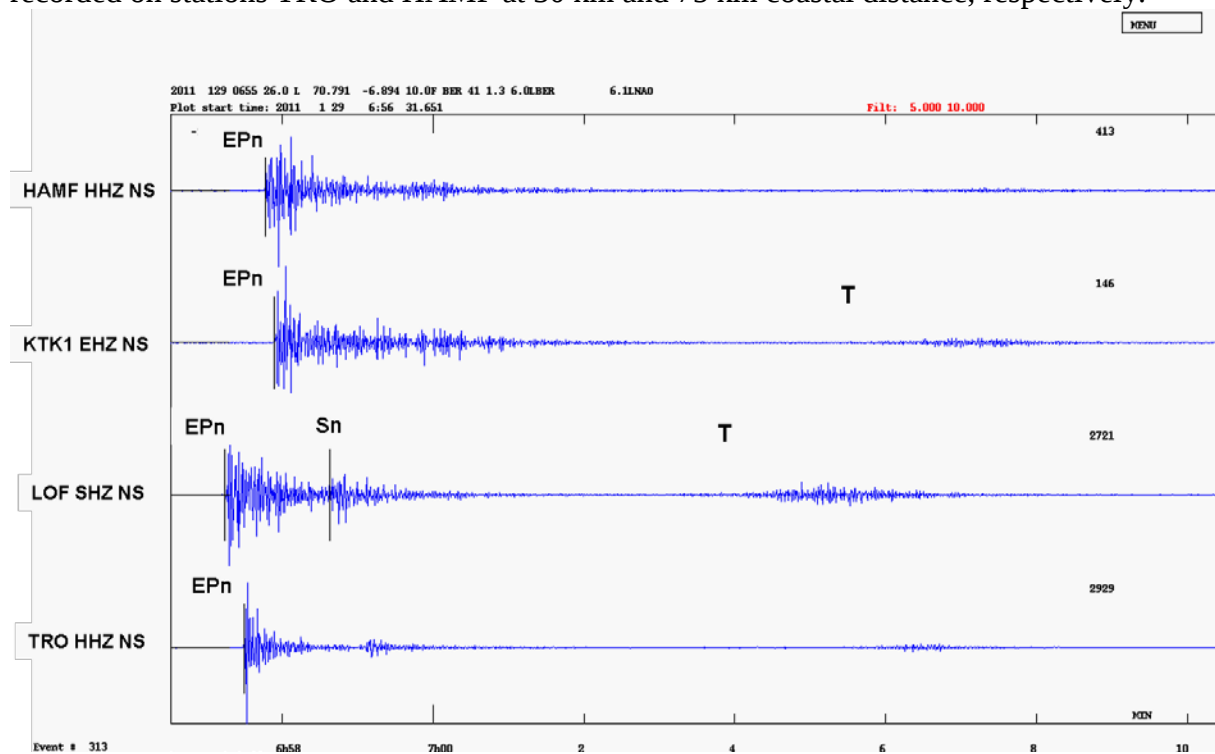
*SOFAR channel* (Bullen and Bolt, 1985) or by multiple reflections between the sea floor and the sea surface (Båth and Shahidi, 1974) and are converted back to seismic waves at the ocean-land boundary near the recording site. The SOFAR channel is a depth range of low acoustic velocity within the ocean water column which “traps” sound waves to propagate over very large distances with very low attenuation. Thus it is a very efficient *wave guide* for the seismic *T phase*. T waves are generally short-period waves because they can propagate efficiently only, if their wavelengths fit inside the width of the SOFAR channel. This is the case for frequencies above 2 Hz only (see Okal et al., 2003, and records in Figs. 2.71 and 2.72). T waves are best recorded by ocean-bottom seismometers (OBS) and hydrophone stations (see Chapter 7, section 7.5) and by near coastal inland stations. Typical for T phases is their gradual increase and subsequent decrease of the amplitudes of rather monochromatic oscillations with a total duration up to several minutes and hardly any recognizable onset time (see the markers ET set in Fig. 2.71).

Since T phases travel through the SOFAR channel with much lower velocity than the direct P and S waves radiated by the seismic source, the T phases arrive much later in the seismic record, the later the longer the SOFAR path. Therefore, they may be misinterpreted as independent seismic events. Figs. 2.71 and 2.72 present T-phase records at stations of the Norwegian National Seismic Network from two earthquakes in the Atlantic Ocean.



**Fig. 2.71** High-frequency filtered records (passband 5 to 10 Hz) at stations LOF, TRO, MOR8 and KTK1 in Norway of the shallow ( $h = 5$  km)  $M_l = 5.2$  Mohns Ridge earthquake on 20 Aug. 2009. Note the at least some 2 minutes long gradual T phase with no clearly recognizable onset ET which follows about 4 to 5 min after the first P onset. The epicenter distance of the stations and their distance from the coast (second number in brackets) are – from top to bottom - for LOF 730 km (15 km), TRO 710 km (30 km), MOR8 860 km (100 km) and KTK1 900 km (200 km). (Modified version of an original record plot kindly provided by Lars Ottemöller, 2011).

Although the amplitudes of the primary and secondary earthquake waves in all 4 records are about the same, the T-phase amplitudes decrease strongly with increasing distance from the coast due to the rather strong attenuation of the high-frequency waves travelling through the rather heterogeneous uppermost crust. This, however, is not a strict rule, obviously modified by differences in the efficiency of the conversion of the acoustic SOFAR channel waves into seismic waves at the coastal bounce point as well as differences in the attenuation conditions along the travel paths on land. In the second record example in Fig. 2.72, e.g., the T-phase amplitudes recorded on station KTK1 at a coastal distance of 220 km are larger than those recorded on stations TRO and HAMF at 30 km and 75 km coastal distance, respectively.



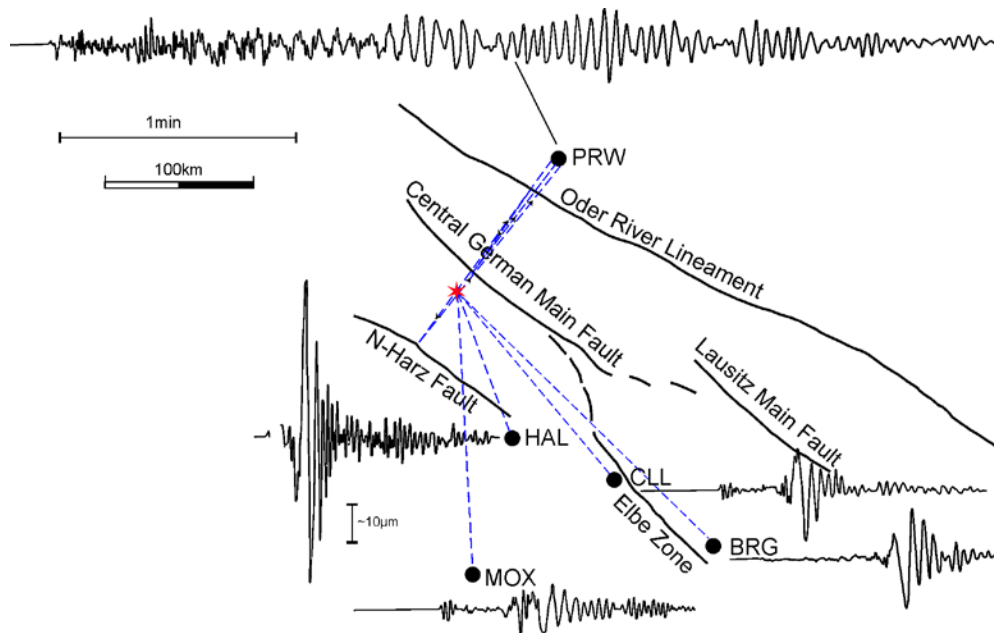
**Fig. 2.72** High-frequency filtered records (passband 5 to 10 Hz) at stations HAMF, KTK1, LOF, and TRO in Norway of the shallow ( $h = 10$  km)  $M_l = 6.1$  Jan Mayen region earthquake on 29 Jan. 2011. Note the up to about 4 min long T phase which follow about 6 to 8 min after the first P onset. The epicenter distance of the stations and their distance from the coast (second number in brackets) are – from top to bottom – for HAMF 1115 km (75 km), KTK1 1165 km (220 km), LOF 845 km (15 km), TRO 970 km (30 km). (Modified version of an original record plot kindly provided by Lars Ottemöller, 2011).

As compared to ordinary sub-marine and *tsunamigenic earthquakes*, high-frequency T waves are highly deficient (by 1.5 to 2.5 orders of magnitude) in records of very slow rupturing *tsunami earthquakes*. Therefore, Okal et al. (2003) suggested that computations of T-phase efficiency might become a valuable contribution to real-time tsunami warning in the far field.

## 2.6.6 Surface-wave fundamental and higher modes and the W phase

The basic types of horizontally propagating seismic surface waves (Rayleigh waves, Love waves, and their higher modes; see 2.3 in Chapter 2) remain more or less unchanged with growing distance. Surface waves, however, do not form seismic phases (wavelets) with well-

defined onsets and duration but rather dispersed wave trains (see Figs. 2.12, 2.13, 2.20). Due to the dispersion their duration increases with distance. Although it is very common to determine the magnitude of shallow seismic events by means of surface waves ( $M_{s\_20}$  and  $M_{s\_BB}$ ; see Chapter 3 and IS 3.3), the analysis of surface-wave group and phase velocities for investigating Earth structure and properties requires specific methods of analysis and is beyond the scope of seismological observatory practice. Moreover, the phenomenology of seismic surface-wave trains is sometimes complicated by near surface heterogeneities. E.g., surface wave trains of relatively high frequencies, as generated by shallow local events, may additionally be prolonged significantly due to lateral reverberations when propagating through strong lateral velocity contrasts in the crust (see Fig. 2.73). This phenomenon was used by Meier et al. (1997) to establish a tomography with reflected surface waves.

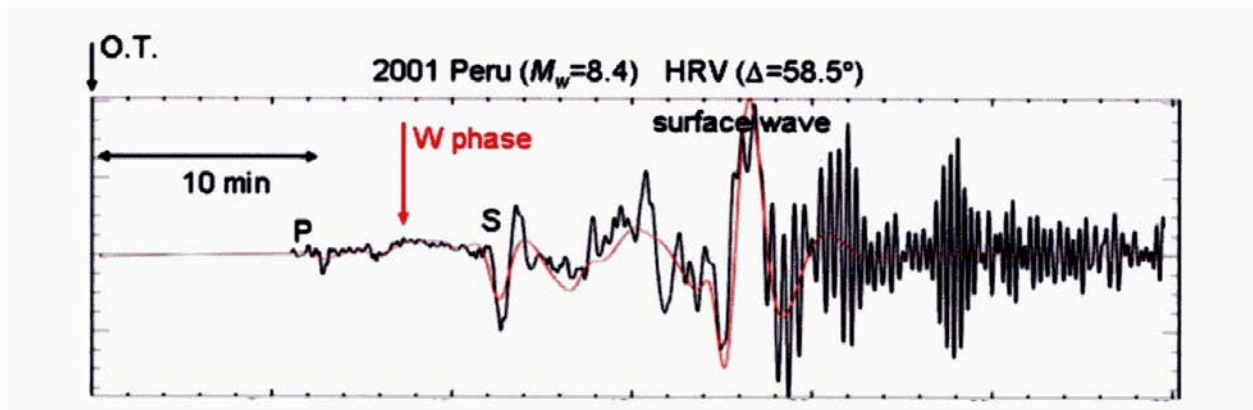


**Fig. 2.73** Ray paths of surface waves (broken lines) from a mining collapse (star) to several seismic stations in the eastern part of Germany. Note: Records at stations along travel paths that have not or only once crossed some of the main tectonic faults in the area, are rather short. They have only one surface-wave maximum. In contrast, at station PRW, which is at the same epicenter distance as HAL, the seismic record is about four times longer and shows four surface-wave groups due to multiple reflections at several pronounced fault systems (compiled from data provided by H. Neunhöfer (1985; and personal communication)).

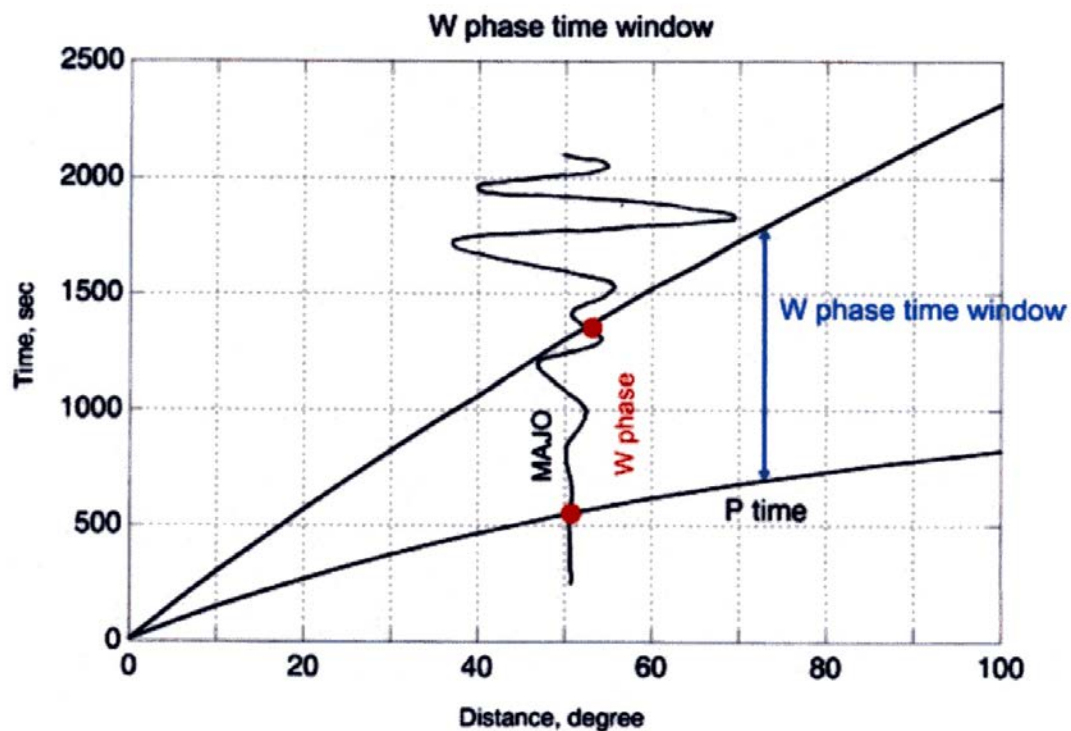
Recently, the analysis of the long-period *W phase*, which arrives before the onset of *S* (Fig. 2.74), has gained immense practical importance for speeding up seismic tsunami warning. Kanamori (1993) has been first to name and extensively document this phase. According to Kanamori and Rivera (2008) the *W phase* can be interpreted as superposition of the fundamental, first, second and third overtones of spheroidal modes or Rayleigh waves. Its group velocity varies between 4.5 and 9 km/s over a period range of 100 to 1000 s. The amplitudes of so long-period waves represent well the tsunami potential of great earthquakes. *W phase* inversion may yield reliable and non-saturating estimates of seismic moment and moment magnitude as well as of the source mechanism of large earthquakes within about 20 minutes of the earthquake occurrence (Fig. 2.75). Source inversion of the *W phase* has



recently been extended down to approximately  $M_w$  6. Real-time implementations are operational at the U.S. Geological Survey's NEIC (Hayes et al., 2009) as well as at the GFZ Potsdam and the Indonesian Tsunami Early Warning System (InaTEWS).



**Fig. 2.74** W phase record of the 2001  $M_w8.4$  Peruvian earthquake at HRV superimposed with the synthetic W phase trace (red) computed by mode summation using the Global Centroid Moment Tensor solution for this earthquake (copy of Figure 1 on page 223 of Kanamori and Rivera, 2008; © Geophysical Journal International).



**Fig. 2.75** Illustration of the distance-dependent time window within which W-phase observations are available for analysis after origin time. (Copy of Figure 15 on page 234 of Kanamori and Rivera, 2008; © Geophysical Journal International).

## 2.7 Global Earth models (E. R. Engdahl)

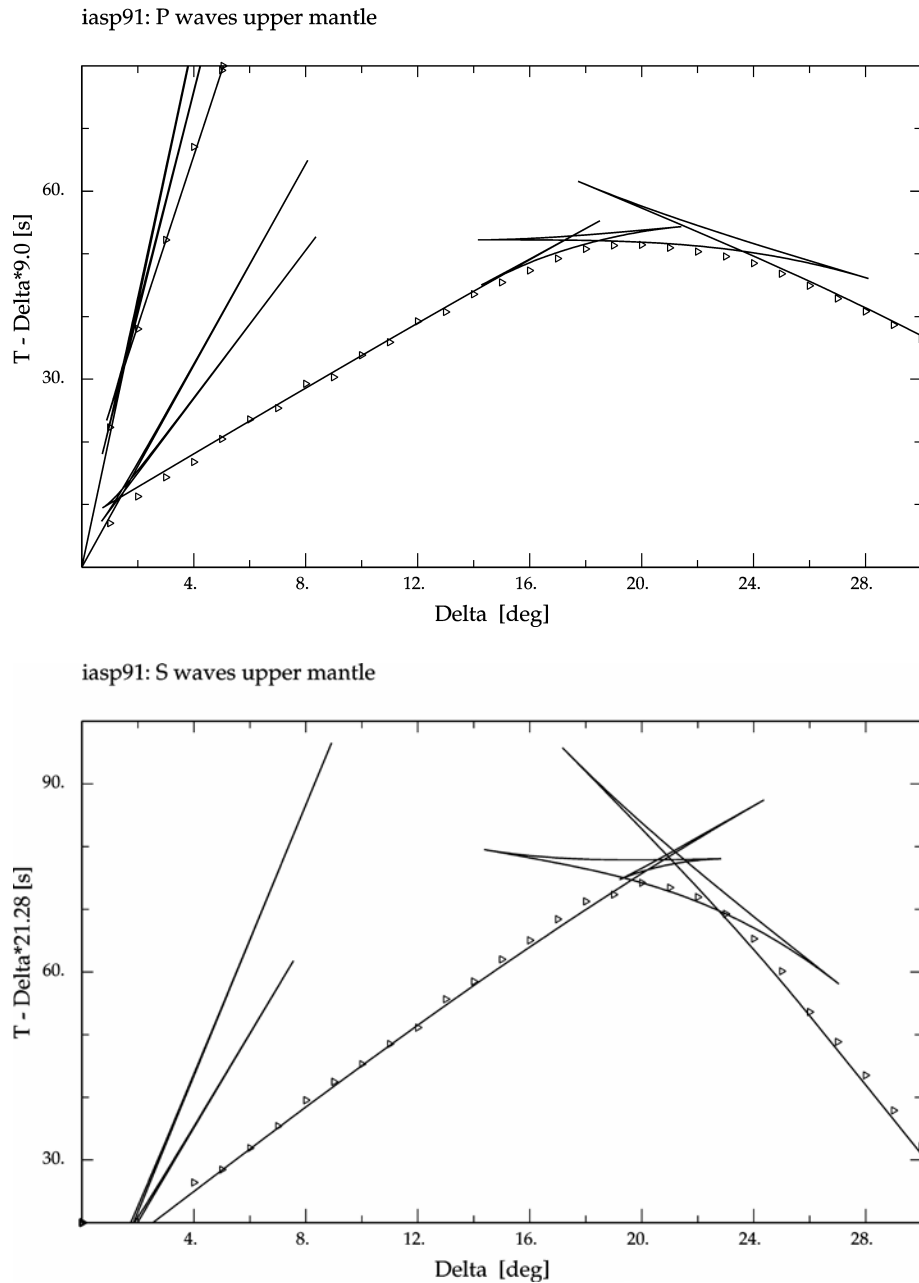
### 2.7.1 1-D models

In the first part of the 20th century travel-time models for seismic phases, empirically derived from historical data, were rudimentary at best. One of the earliest travel-time model, the Zoeppritz tables (Zoeppritz, 1907) were applied by Herbert Hall Turner in a version as published by Galitzin (1914) to locate earthquakes for the ‘Bulletin of the British Association of the Advancement of Science, Seismology Committee’ for the years 1914 until 1917. During the 1920s, Turner gradually expanded these tables for newly discovered phases and better phase observations, often suggested and derived by Beno Gutenberg. These Zoeppritz-Turner tables were in use to locate earthquakes for the International Seismological Summary (ISS) from 1918 to 1929. This situation greatly improved with the introduction of the Jeffreys-Bullen (J-B) tables (Jeffreys and Bullen, 1940), which provided a complete, remarkably accurate representation of P, S and other later-arriving phases. Like the Gutenberg-Richter travel-time tables, the J-B tables were developed in the 1930s using reported arrival times of seismic phases from a sparse global network of stations, many of which often had poor time-keeping. Once the travel times of the main phases had been compiled, smoothed empirical representations of these travel times were inverted using the Herglotz-Wiechert method to generate a velocity model. The travel times for other phases were then determined directly from the velocity model. As a testament to the careful work that went into producing the J-B tables, they were until only recently used by the International Seismological Centre (ISC) and by the U. S. Geological Survey National Earthquake Information Center (NEIC) for routine earthquake location. Both agencies now use the AK135 tables.

Although the limitations of the J-B tables were known for some time, it was not until the early 1980’s that a new generation of models was constructed in a completely different way. Instead of establishing smoothed, empirical representations of phase-travel times, inverse modeling was used to construct one-dimensional models for structure that fit phase travel times reported in the ISC Bulletin since 1964 and other parametric data. The Preliminary Reference Earth Model (PREM) of Dziewonski and Anderson (1981) was the most important member of this generation of new global 1-D models. However, PREM was constructed to fit both body-wave travel-time and normal-mode data, so it was not generally thought to be especially useful for earthquake location. In fact, soon afterwards Dziewonski and Anderson (1983) published a separate analysis of just P waves in an effort to produce an improved travel-time table.

In 1987 the International Association of Seismology and Physics of the Earth’s Interior (IASPEI) initiated a major international effort to construct new global travel-time tables for earthquake location and phase identification. As a result of this effort two models were developed: IASP91 (Kennett and Engdahl, 1991); and SP6 (Morelli and Dziewonski, 1992). Although differences in predicted travel times between these two models were small, some effort was still required to reconcile the travel times of some important, well-observed seismic phases before either of these models could be used by the ISC and NEIC for routine earthquake location. The upper mantle part of the IASP91 model was fitted to summary P and S wave travel-times, binned in 1° intervals of epicenter distance, published by Dziewonski and Anderson (1981, 1983) (Fig. 2.76).

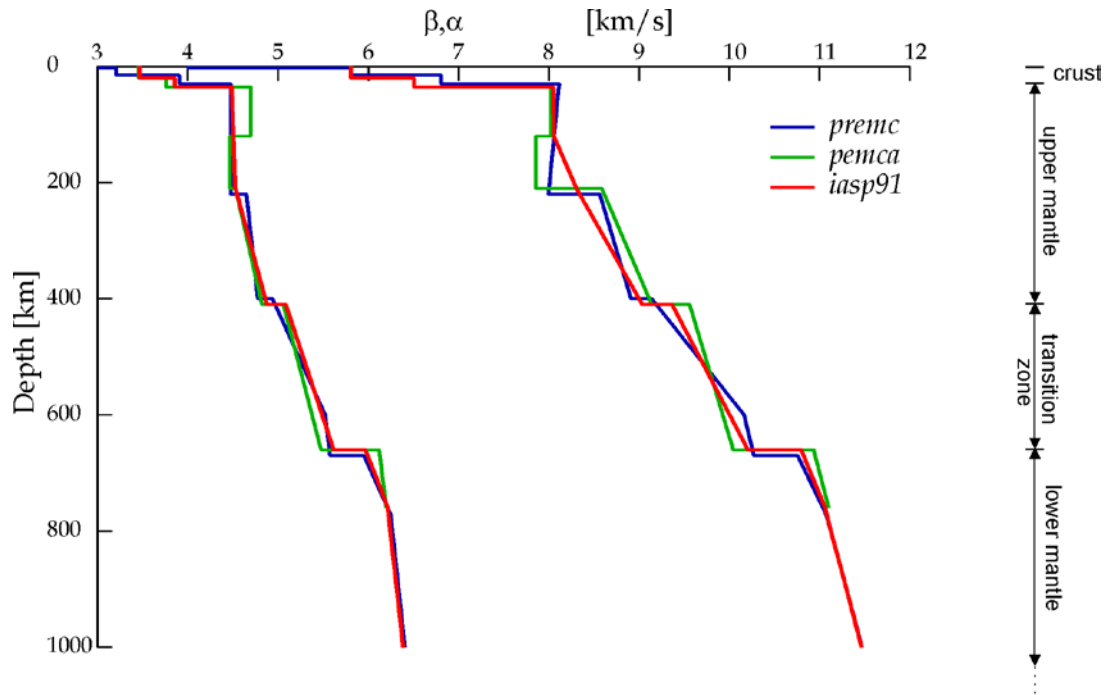




**Fig. 2.76** Fitting of IASP91 upper mantle travel times as a function of epicenter distance to the summary first-arrival travel times of P (top) and S waves (bottom) according to Dziewonski and Anderson (1981, 1983) in time-reduced presentation (from Kennett and Engdahl, 1991).

As shown in Fig. 2.77, the IASP91 upper mantle differed substantially from PREM and, in particular, IASP91 had no mantle low-velocity zone for either P or S waves. Although this did run counter to the prevailing ideas about upper mantle structure, it did have a practical advantage for locating events because the upper mantle travel times in IASP91 were not discontinuous. Characteristics of the main upper mantle discontinuities were also different from previous models. In IASP91 the 210 km discontinuity was essentially absent. The 410 km and 660 km discontinuity velocity jumps in IASP91 were slightly greater in amplitude than in PREM. Path coverage was generally more uniform in the lower mantle, so these parts of the IASP91 P and S models were considered to be more representative of the average

Earth. P structure was reasonably well constrained, except near the core-mantle boundary, but the complication of interfering phases put a practical limit on the amount and quality of data constraining S structure. Nevertheless, IASP91 seems to have done a reasonably good job of representing teleseismic travel times, as indicated by the analysis of arrival-time data from well-constrained explosions and earthquakes (Kennett and Engdahl, 1991).



**Fig. 2.77** Comparison of upper mantle velocity models for IASP91, PEMCA, and PREM (from Kennett and Engdahl, 1991). Left:  $\beta$  - speed of S wave; right:  $\alpha$  - speed of P wave.

Morelli and Dziewonski (1993) developed an alternative model (SP6) using the same model parameterization and upper mantle model as Kennett and Engdahl (1991). In their approach, they solved for multiple source-region station corrections averaged over  $5^\circ$  areas to account for lateral heterogeneity in an approximate manner. They then derived new sets of summary travel times for lower mantle and core P and S phases binned in  $1^\circ$  intervals of epicenter distance, and inverted those summary times for 1-D P and S velocity models. Although lower mantle P and S in the resulting model was generally comparable to IASP91, the models differed in that SP6 had slightly lower velocity gradients with depth and correspondingly higher velocity jumps at the 660-km discontinuity. Moreover, SP6 had incorporated a pronounced negative velocity gradient in the D'' region, a layer 100-150 km thick just above the core-mantle boundary. The SP6 model fitted the S data and all core-phase observations significantly better than IASP91. The differences in the seismic velocities between the models were significant for the core, owing to the addition of substantial core phase data in the construction of SP6.

The most significant differences between these new models and the older J-B travel-time model are in the upper mantle and core. The upper mantle is highly heterogeneous. Hence, velocities and major discontinuities in the upper mantle of recent models such as IASP91 and SP6 are set at values, which give an effective average representation of velocities for waves

traveling out to  $25^\circ$  (see Kennett and Engdahl, 1991). The core models for IASP91 and SP6 predict more accurately than the J-B model the observed travel times of later-arriving core phases bottoming in the lowermost part of the outer core.

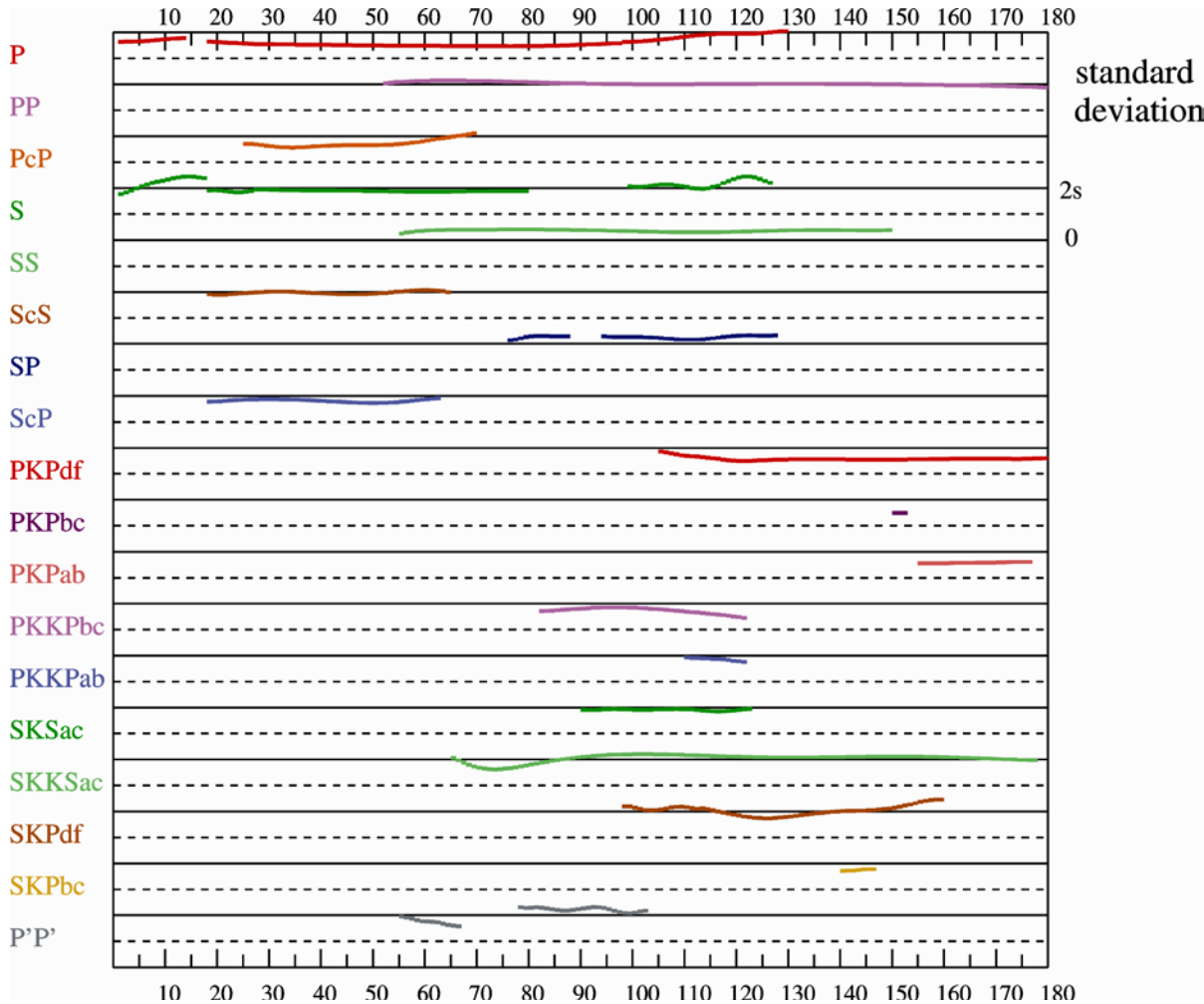
These models also resolve a long-standing problem in that the relocation of nuclear tests using the J-B travel-time model results in incorrect estimates of the origin times of nuclear explosions by about -1.8 sec. This error will propagate into all derived travel times and may affect the procedure of phase association. Kennett and Engdahl (1991) resolved this error in the absolute travel time (or "baseline" error) by fitting the IASP91 model to the mean teleseismic residual estimated from the origin times and hypocenters reported for explosions and well-constrained earthquakes by "test event" contributors. As a result, the times of teleseismic P and S waves for the IASP91 model now appear to be in better agreement with the travel time data than the times predicted by the J-B model. The IASP91 model has been adopted as the global reference model for the International Data Centre in Vienna established under the 1996 Comprehensive Nuclear-Test-Ban Treaty (CTBT).

Subsequently, Kennett et al. (1995) began with the best characteristics of the IASP91 and SP6 models and sought to enhance the data quality by improving the locations of a carefully selected set of geographically well-distributed events. The basic strategy was to use a location algorithm developed by Engdahl et al. (1998) with a IASP91 model modified to conform to the SP6 core to relocate events and improve phase identifications using only first arriving P phases and re-identified depth phases (pP, pwP and sP). The resulting set of smoothed empirical relations between travel time and epicenter distance for a wide range of re-identified seismic phases was then used to construct an improved reference model for the P and S radial velocity profile of the Earth (AK135). A composite residual plot (Fig. 2.78) shows that the model AK135 provides a very good fit to the empirical times of 18 seismic phases. The baseline and trend of S is well presented and most core phase times are quite well matched. Thus, for improved global earthquake location and phase association, there has been convergence on effective global, radially symmetric P- and S-velocity Earth models that provide a good average fit to smoothed empirical travel times of seismic phases.

The primary means of computing travel times from such models is based on a set of algorithms (Buland and Chapman, 1983) that provide rapid calculation of the travel times and derivatives of an arbitrary set of phases for a specified source depth and epicenter distance. In the mantle, AK135 differs from IASP91 only in the velocity gradient for the D" layer and in the baseline for S wave travel times (about -0.5 sec). Significant improvement in core velocities relative to earlier model fits was also realized. Inner core anisotropy, as discussed in the literature, is not yet accounted for in any of the newer 1-D Earth models. However there are so few reported arrivals of PKPdf at large distances along the spin axis of the Earth that the effects of this anisotropy in earthquake location are negligible.

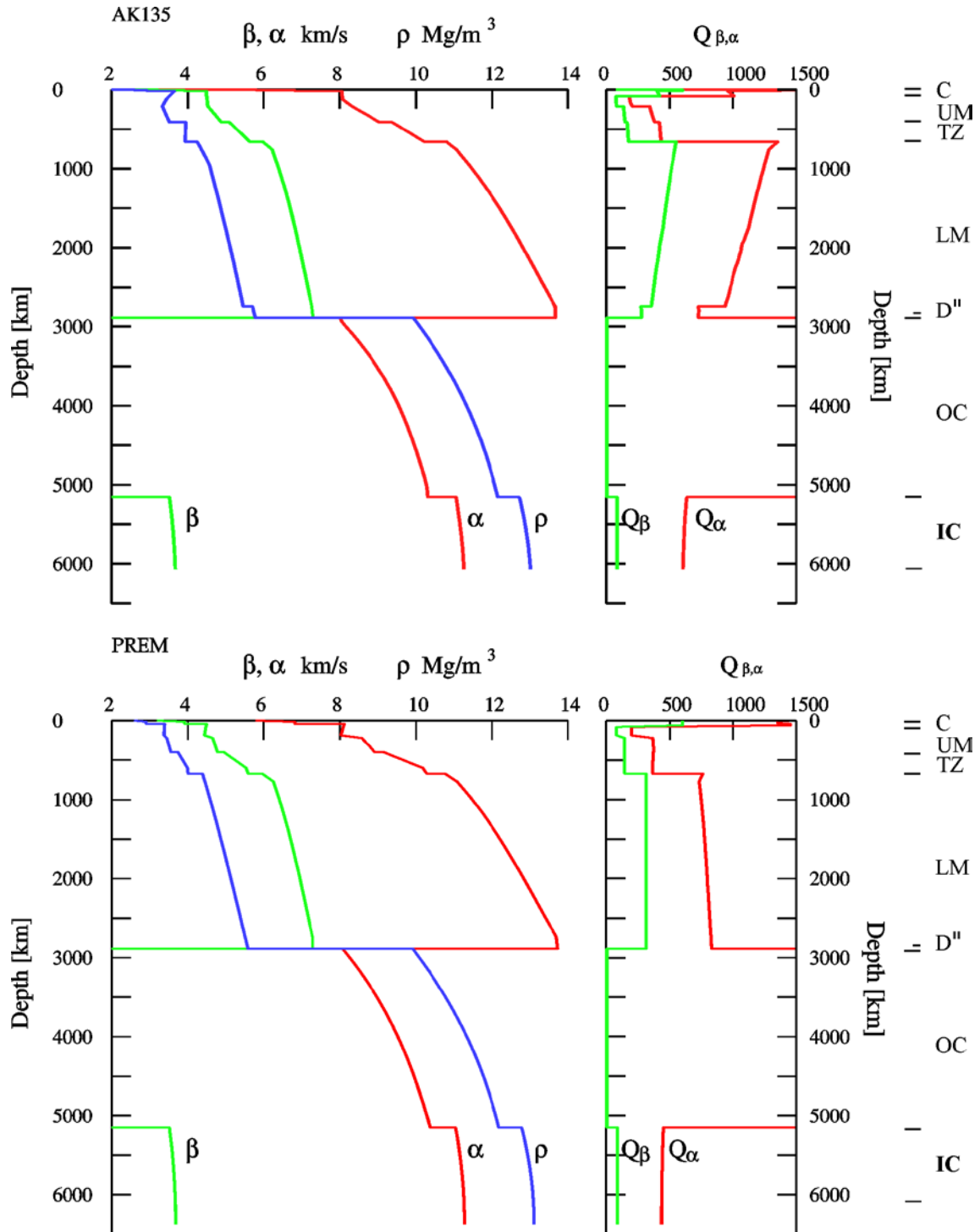
The model AK135 has since been used for further reprocessing of the arrival time information (Engdahl et al., 1998). The reprocessed data set and the AK135 reference model have formed the basis of much recent work on high-resolution travel-time tomography to determine three-dimensional variations in seismic wave speed (e.g., Bijwaard et al., 1998). However, it is important to recognize that none of these models can properly account for the effect of lateral heterogeneities in the Earth on teleseismic earthquake location. Most deeper than normal earthquakes occur in or near subducted lithosphere where aspherical variations in seismic wave velocities are large (i.e., on the order of 5-10%). Such lateral variations in seismic velocity, the uneven spatial distribution of seismological stations, and the specific choice of

seismic data used to determine the earthquake hypocenter can easily combine to produce bias in teleseismic earthquake locations of up to several tens of kilometers (Engdahl et al., 1998). For a review of recent advances in teleseismic event location, with the primary emphasis on applications using one-dimensional velocity models such as AK135, the reader is referred to Thurber and Engdahl (2000). The most accurate earthquake locations are best determined using a regional velocity model with phase arrival times from a dense local network, which may differ significantly (especially in focal depth) from the corresponding teleseismic locations.



**Fig. 2.78** Composite display of the estimates of standard deviations for the empirical travel times used in the construction of the AK135 velocity model (Kennett et al., 1995).

The AK135 wave speed reference model is shown in Fig. 2.79. However, though the P- and S-wave speeds are well constrained by high-frequency seismic phases, more information is needed to provide a full model for the structure of the Earth.



**Fig. 2.79** Radial symmetric reference models of the Earth. **Top:** AK135 (seismic wave speeds according to Kennett et al., 1995), attenuation parameters and density according to Montagner and Kennett (1996); **Bottom:** PREM (Dziewonski and Anderson, 1981).  $\alpha$  - and  $\beta$ : P- and S-wave velocity, respectively;  $\rho$  - density,  $Q_\alpha$  and  $Q_\beta = Q_\mu$  - “quality factor”  $Q$  for P and S waves. Note that wave attenuation is proportional to  $1/Q$ . The abbreviations on the outermost right stand, within the marked depth ranges, for: C – crust, UM – upper mantle, TZ – transition zone, LM – lower mantle, D''-layer, OC – outer core, IC – inner core.

In particular, any reference model should also include the density and inelastic attenuation distributions in the Earth. Work by Montagner and Kennett (1996) provided these parameters which, although known less precisely than the seismic velocities, are needed because it makes the model suitable for use as a reference to compute synthetic seismograms (see 2.8) without requiring additional assumptions. Nevertheless, the primary use of AK135 (and IASP91) remains earthquake location and phase identification. The PREM model of Dziewonski and Anderson (1981), also shown for comparison in Fig. 2.79, forms the basis for many current studies on global Earth's structure using quantitative exploitation of seismic waveforms at longer periods. It is the objective of the 'Working Group on Reference Earth models' in the 'IASPEI Commission on Earth Structure and Geodynamics' to retrieve a new 1-D reference Earth model for many depth-depending parameters which is also in agreement with observations of the Earth's normal modes.

The IASPEI 1991 Seismological Tables (Ed. Kennett, 1991) are now out of print. The more recent global P- and S-wave velocity and density model AK135, and the related body wave travel-time tables and plots are available via <http://rses.anu.edu.au/pub/ak135> and can be downloaded or printed in postscript. Additionally, software for travel-time routines and for corrections of the ellipticity of the Earth can be obtained via <http://rses.anu.edu.au/seismology/ttsoft.html>.

### **2.7.2 3-D models and their use in event location**

To first order, the structure of the Earth is dominated by its radial structure. However, important deviations between observed and theoretical travel times of seismic waves with respect to 1D-reference models cannot be explained by these simple 1D models. These differences, though relatively small (<10%) are due to the existence of lateral heterogeneities of physical properties within the different layers of the Earth. The determination of 3D models of seismic parameters of the mantle (also coined tomographic models) consists in imaging the structure of convection in order to provide clues to the nature of the driving mechanism and to investigate mantle convective processes, the engine of plate tectonics. Thanks to the intensive use of computers which can handle very large data sets, the techniques of seismic tomography can image the 3D deep structure of the Earth and provide unavoidable constraints of possible models of mantle convection and makes it possible to test different competing hypotheses. However, the main limitation of seismic tomography is that it only provides an instantaneous image of our planet Earth and it is necessary to confront seismic results with other earth science fields providing information on the history of our planet.

Seismic wave propagation is very sensitive to temperature, chemical, and mineralogical anomalies. These anomalies reflect the active dynamics of the Earth. It was observed for many decades that seismic velocity below old continents is much larger than the velocity beneath young oceans. This difference between tectonic provinces can be explained by a different history: cratons older than 1 billion years had a lot of time to get colder, whereas below oceans, the younger and warmer lithosphere is permanently recycled within the mantle. An increase in temperature is usually translated into decrease in seismic velocity, whereas an increase in pressure is translated into velocity increase. At a given depth (approximately constant pressure), the lateral variations of seismic velocity can be primarily interpreted in terms of temperature variations. A large variety of tomographic techniques similar to ultra-sonography or scanner have been developed for getting the 3D structure of the Earth, notwithstanding the geological nature of the surface.

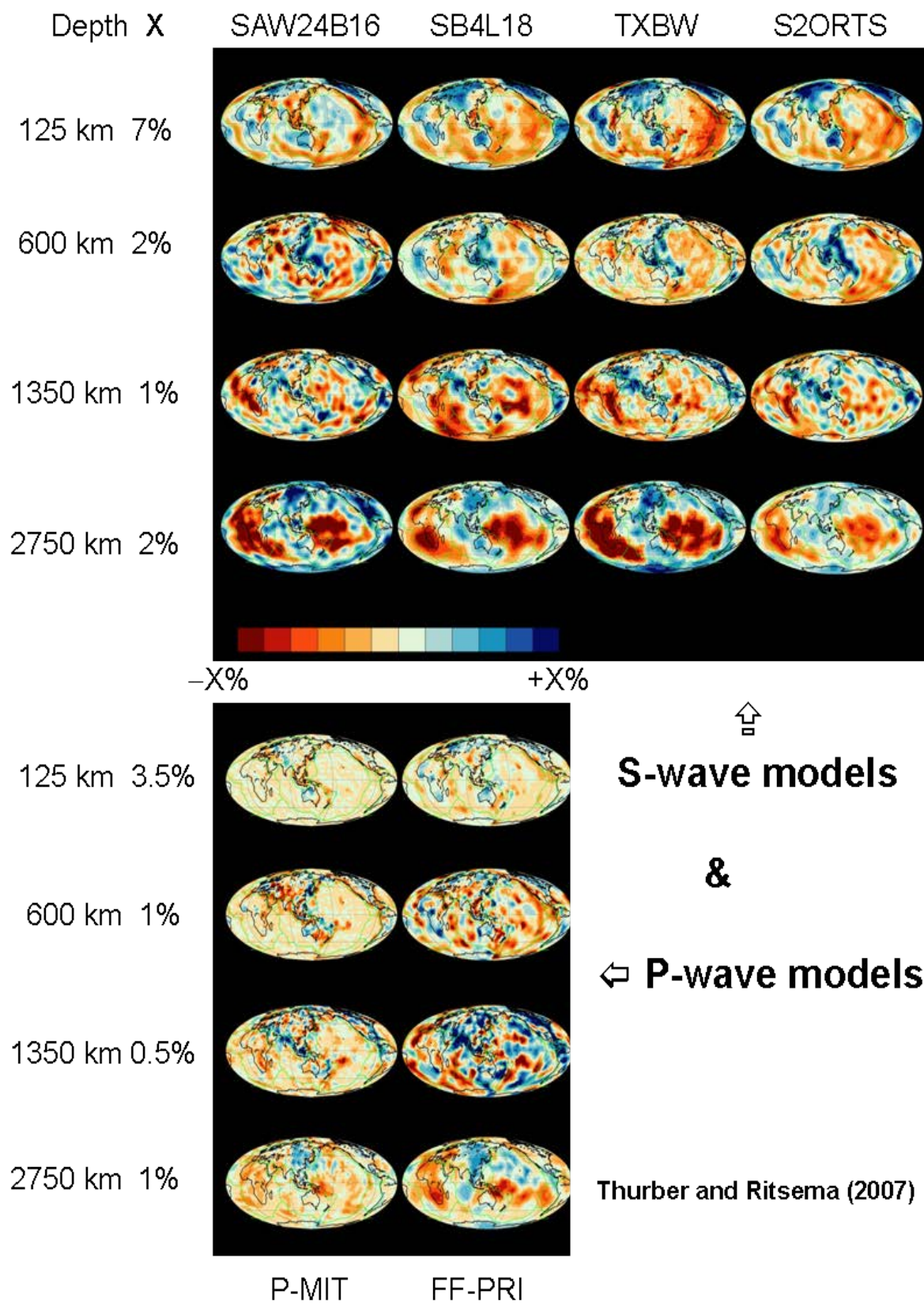
The first 3D models of the Earth's mantle were proposed by the end of the 1970s (for example, Dziewonski et al., 1977), but the true revolution of seismic tomography took place in the middle 1980s when a complete set of tomographic models was proposed from the surface down to the center of the Earth. The present lateral resolution at global scale for the whole Earth is approximately 1,000km. Tomographic models (Fig. 2.80) fundamentally renewed our vision of earth dynamics, and shed new light on deep processes. At 100km depth, the distribution of seismic velocity anomalies reflects the geological surface structure. Almost all plate boundaries are slow since they are often associated with active volcanism. In oceanic zones, seismic velocity increases with the age of the sea floor confirming the cooling of plates with age. This remarkable agreement between observed seismic structure and structure deduced from geology was the first proof of the validity of the tomographic approach. At larger depths, the correlation between surface geology and seismic structure progressively vanishes and the amplitude of seismic anomalies decreases. The continental roots can be found down to 250-300km but mid-ocean ridges are no longer visible. In contrast, subducting slabs associated with high velocities can be traced down very deeply. The transition zone (410-660km) is still one of the less well-resolved depth ranges. It is characterized by weak heterogeneities but the velocity distribution is dominated by a simple scheme, with high velocities around the Pacific ocean and slow velocities below central Pacific and Africa, the so-called “degree 2” pattern (Masters et al., 1982).

The behaviour of slabs in regional studies does not seem to be uniform (Fig. 2.81). Some slabs seem to fall down to the core-mantle boundary (van der Hilst, et al., 1997), whereas others, e.g. beneath Japan (Fukao et al, 2001) are stagnant in the transition zone. Mantle plumes and the associated hotspots might initiate continental breakup, inducing the opening of new ridges of oceans, and might dramatically affect biodiversity. After plate tectonics, plume tectonics? There is still a controversy on the origin of plumes. It is likely that they do not have a single origin, but can originate from the different boundary layers within the Earth, in the asthenosphere, transition zone and D''-layer. There are still lively debates about the origin at depth of mantle plumes and two opposite extreme models (plate model and plume model) have been proposed. At the base of the mantle, the D''-layer is still very mysterious since it is the probable graveyard of slabs and a good candidate for being the matrix of some mantle plumes giving birth to hotspots at the surface of the Earth.

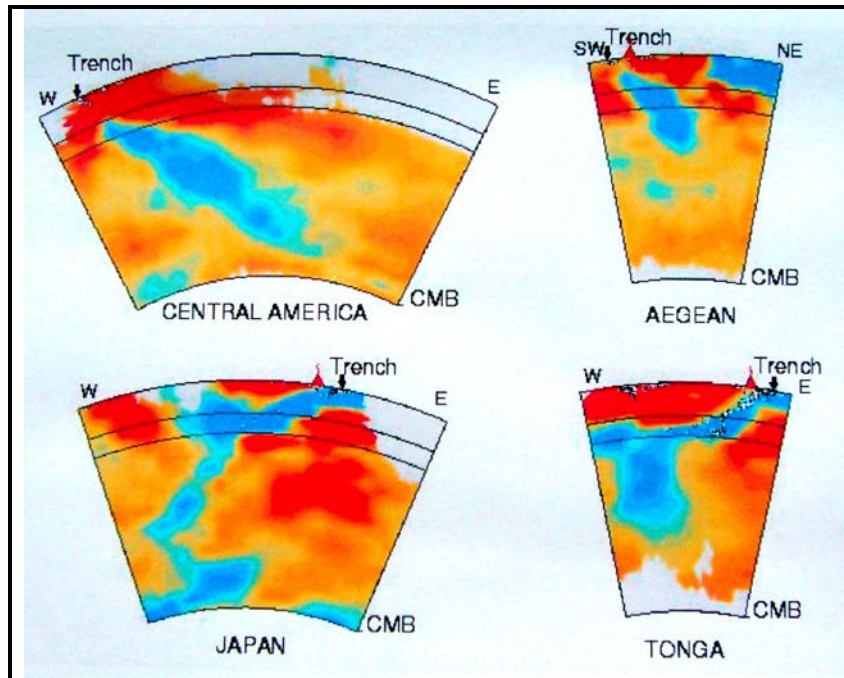
For sake of mathematical simplicity, it was usually assumed that the propagating elastic medium is isotropic, in spite of the large amplitude of anisotropy in several depth ranges. And only 1-D reference models of attenuation are incorporated in seismic modelling. Global tomographic models improved over years not only by an increase in the number of data but also by more general parameterizations. They are now including anisotropy, general slight anisotropy (for example, Montagner and Tanimoto, 1991), and anelasticity. Whereas isotropic heterogeneities enable one to map hot and cold regions within the Earth, seismic anisotropy gives access to the convective flow.

Impressive progress through global tomography, geochemistry and mineral physics has been made during the last twenty years in our understanding of global geodynamics, demonstrating how active and turbulent our planet is. All layers are heterogeneous, interact and exchange matter. The geosciences community must improve the lateral resolution and the quality of the 3D images and incorporate on a routine basis, anisotropy and anelasticity. There is a real need for a 3D seismic reference earth model in agreement with geological, mineralogical data, and fluid dynamics modelling.





**Fig. 2.80** Maps through S-velocity models SAW24B16, SB4L18, TXBW, and S2ORTS and P-velocity models P-MIT and FF-PRI at depths of 125, 600, 1350, and 2750 km (compiled by P. Bormann according to data published by Thurber and Ritsema, 2007).



**Fig. 2.81** Tomographic images of slabs (Top: van der Hilst et al., 1997) and plumes (Bottom: Montelli et al., 2004). **Blue colors** and **red colors** indicate positive and negative velocity perturbations to reference mantle velocities, respectively.

Currently almost all published earthquake catalogues have applied traditional, iterative linearized inversion schemes and 1-D Earth models to obtain event location and uncertainty parameters. Although there is a considerable effort to apply 3-D Earth models for travel time predictions, these methods and Earth models have yet to find their way into routine production of earthquake catalogs. Nevertheless, progress is being made.

In the local distance range ( $0^{\circ}$ – $2.5^{\circ}$ ) it has been shown (Bondar et al, 2004) that, using an accurate 1-D local velocity model and Pg and Sg arrival times recorded by a dense network of stations at distances less than about 250 km, one can achieve a location accuracy of 5 km or better at the 95% confidence level. However, at regional ( $2.5^{\circ}$ – $20^{\circ}$ ) distances lateral heterogeneity in the Earth's crust and uppermost mantle can severely affect Pn travel times to stations in that distance range, leading to location errors of 25 km or more when using a 1-D model. This has been a challenge to the test ban monitoring community, which has led to new developments in the use of 3-D models and azimuthally based station corrections to reduce location errors to as small as 5 km when using regional station arrival times to locate events in Eurasia.

When stations at teleseismic distances are used for location, P waves are travelling primarily in the Earth's lower mantle, with only 1-2% lateral heterogeneity, and location errors can be reduced to on average 15 km for earthquakes occurring in continental regions. However, in subduction zones location errors can be as large as several tens of km because of the effects of high velocity (+8 to 10%) downgoing slabs in the mantle. Only now are 3-D global models of regional and lower mantle structure being developed that can account for these effects, but for location applications these models require intensive ray tracing using high-speed computers for travel time prediction and phase identification.

## 2.8 Synthetic seismograms and waveform modeling

(R. Kind and P. Bormann)

A good measure of the advancement made by a scientific discipline is its ability to predict the phenomena with which it is dealing. One of the goals of seismology, as stated already over a hundred years ago by Emil Wiechert, is to understand every wiggle on the seismogram. This requires, as sketched in Fig. 1.1 of Chapter 1, an understanding and quantitative modeling of the contributions made to the seismic record (the output) by the various subsystems of the complex information chain: the source effects (input), the propagation effects (medium), the influence of the seismograph (sensor) and of the data processing. It is possible nowadays to model each of these effects quite well mathematically and thus to develop procedures for calculating *synthetic seismograms*. While the modeling of the seismometer response (see Chapter 5) and of the source effects (see 3.5 and IS 3.1) have been outlined in more detail in this Manual, it is beyond the scope of a handbook on observatory practice to go into the depth of wave propagation theory. Here we have to refer to pertinent textbooks such as Aki and Richards (1980 and 2002), Kennett (1983, 2001 and 2002), Lay and Wallace (1995), Dahlen and Tromp (1998) or, for some condensed introduction, to Shearer (1999). Below we will only sketch some of the underlying principles, refer to some fundamental approaches, discuss their potential and shortcomings and give a few examples of synthetic seismogram calculation and waveform modeling for near and teleseismic events.

Based on advanced theoretical algorithms and the availability of powerful and fast computers the calculation of synthetic seismograms for realistic Earth models is becoming more and more a standard procedure both in research and in advanced observatory routines. Such calculations, based on certain model assumptions and parameter sets for the source, propagation path and sensor/recorder are sometimes referred to as the solution of the *direct* or *forward problem* whereas the other way around, namely, to draw inferences from the observed data itself on the effects and relevant parameters of propagation path and source is termed the *inverse problem* (see Fig. 1.1). With the exception of a few specialized cases of direct analytical solutions to the inverse problem (such as using the Wiechert-Herglotz inversion [Eqs. (2.21) and (2.22)] for calculating the velocity-depth distribution of the medium from the observed travel-time curves), most inverse problems are solved by comparing synthetic data with observed ones. The model parameters are then changed successively in an iterative process until the differences between the observed and the synthetic data reach a minimum. The procedure of comparing synthetic and observed seismograms is known as *waveform modeling*. It can be used in routine practice for better identification of seismic phases and more reliable onset-time picking in case of noisy data. Additionally, more and more advanced seismological data centers, such as NEIC, now make use of waveform fitting for fast seismic moment tensor and other source parameter solutions, such as source depth (see 3.5.6.1).

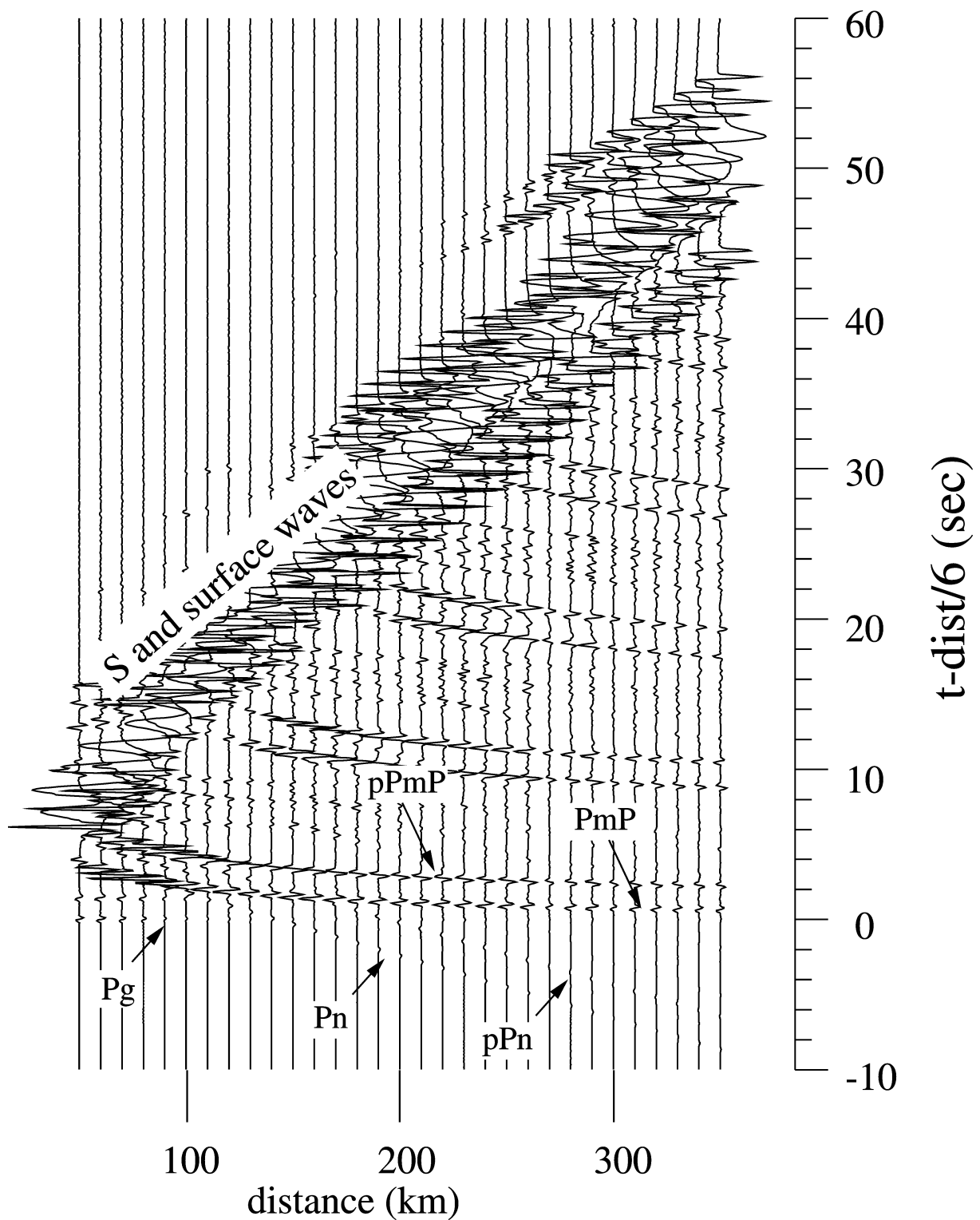
The underlying mathematical tool for constructing synthetic seismograms is the *linear filter theory*. The seismogram is thus treated as the output of a sequence of linear filters, each accounting for relevant aspects of the seismic source, propagation path and sensor/recorder. Accordingly, the seismogram  $u(t)$  can be written as the result of convolution of three basic filters, namely:

$$u(t) = s(t) * g(t) * i(t), \quad (2.37)$$

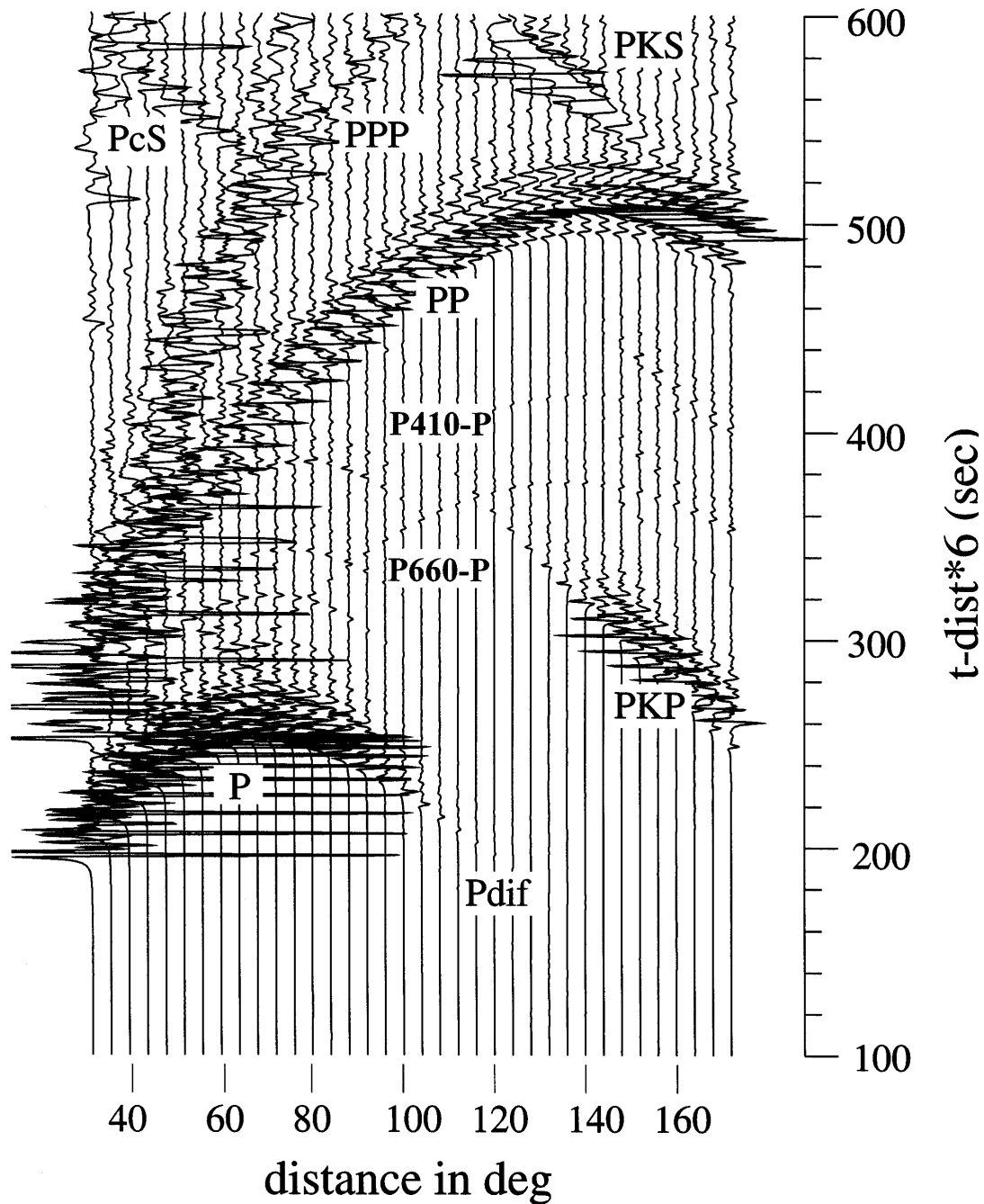
where  $s(t)$  is the signal from the seismic source,  $g(t)$  is the propagation filter, and  $i(t)$  is the overall instrument response. These basic filters can in fact be broken down into various sub-filters, each accounting for specific effects of the source (such as source radiation directivity, source-time function), the propagation medium (such as structure and attenuation) or the instrument (such as sensor and recorder). This makes it possible to study in detail the effects of a specific parameter or process on the character of the seismogram, e.g., the effects of the shape and bandwidth of the seismograph response on the recording (see 4.2) or of the source depth, rupture orientation or time-history of the rupture process on the signal shape (see pp. 400-412 in Lay and Wallace, 1995). With respect to the propagation term in Eq. (2.37) it may be modelled on the basis of a full wave-theoretical approach, solving Eq. (2.5) for 1-D media consisting of stacks of homogeneous horizontal layers. The complete response of such series of layers may be described by matrixes of their reflection and transmission coefficients and a so-called *propagator algorithm* (Thomson, 1950 and Haskell, 1953) or by generalized reflection and transmission coefficients for the entire stack as in the reflectivity method by Fuchs and Müller (1971), Kennett (1983), Müller (1985). Another, ray theoretical approach (e.g., Červený et al., 1977; Červený, 2001) is possible when assuming that variations in the elastic parameters of the media are negligible over a wavelength and thus these gradient terms tend to zero at high frequencies. While pure ray tracing allows one only to model travel-times, the assumption of so-called "*Gaussian beams*", i.e., "ray tubes" with a Gaussian bell-shaped energy distribution, permits the modeling of both travel-times and amplitudes and thus to calculated complete synthetic seismograms also for non-1-D structures.

While a decade ago limited computer power allowed one to model realistically only relatively long-period teleseismic records, it is now possible to compute complete short-period seismograms of up to about 10 Hz or even higher frequencies. Several program packages (e.g. Fuchs and Müller, 1971; Kind, 1978; Kennett, 1983; Müller 1985; Sandmeier, 1990; Wang, 1999) permit one to compute routinely for given source parameters and, based on 1-D Earth models, synthetic seismograms for both near field and teleseismic events.

Two examples of synthetic seismogram sections in reduced travel-time presentation are shown below. Fig. 2.82 shows records for the local/regional distance range between 50 and 350 km with P, S and surface waves in the frequency range between about 0.5 and 2 Hz. Fig. 2.83 compiles synthetic records for longitudinal and some converted phases with frequencies between about 0.1 and 0.3 Hz in the teleseismic distance range between 32° and 172°. The earth-flattening approximation of Müller (1977) is used to transform the flat layered model into a spherical model. This approximation does not permit calculation of phases travelling close to the center of the Earth. The theoretical record sections are noise-free and have simpler waveforms than most real seismograms, owing to the assumption of a simple source function. Fig. 2.82 does not show signal-generated codas of scattered waves that are so typical for short-period records of local events.



**Fig. 2.82** Synthetic seismogram sections in the distance range 50-350 km, calculated for a hypothetical explosive source at 6 km depth in a homogenous single layer crustal model of 30 km thickness. For the calculation the program by Kind (1978) was used.



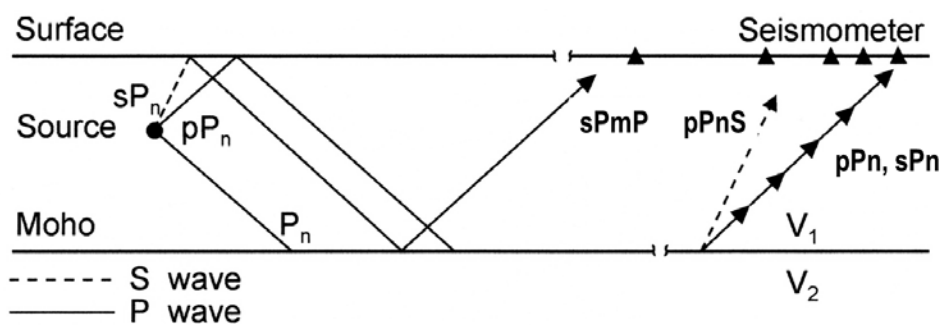
**Fig. 2.83** Long-period synthetic seismic record section for the epicenter distance range 36°-166°, assuming a surface explosion and wave-propagation through the IASP91 model (Kennett and Engdahl, 1991). For the calculation the program by Kind (1978) was used.

The synthetic record sections shown in Figs. 2.82 and 2.83 provide some general insights into basic features of seismograms in these two distance ranges such as:

- The overcritical Moho reflections PmP have the largest amplitudes in the P-wave part of near seismic recordings, with maximum amplitudes near the critical point around 70 km;
- Pg is the first arrival up to about 140 km (for a crustal thickness of 30 km) with amplitudes decaying rapidly with distance in this simple model example;

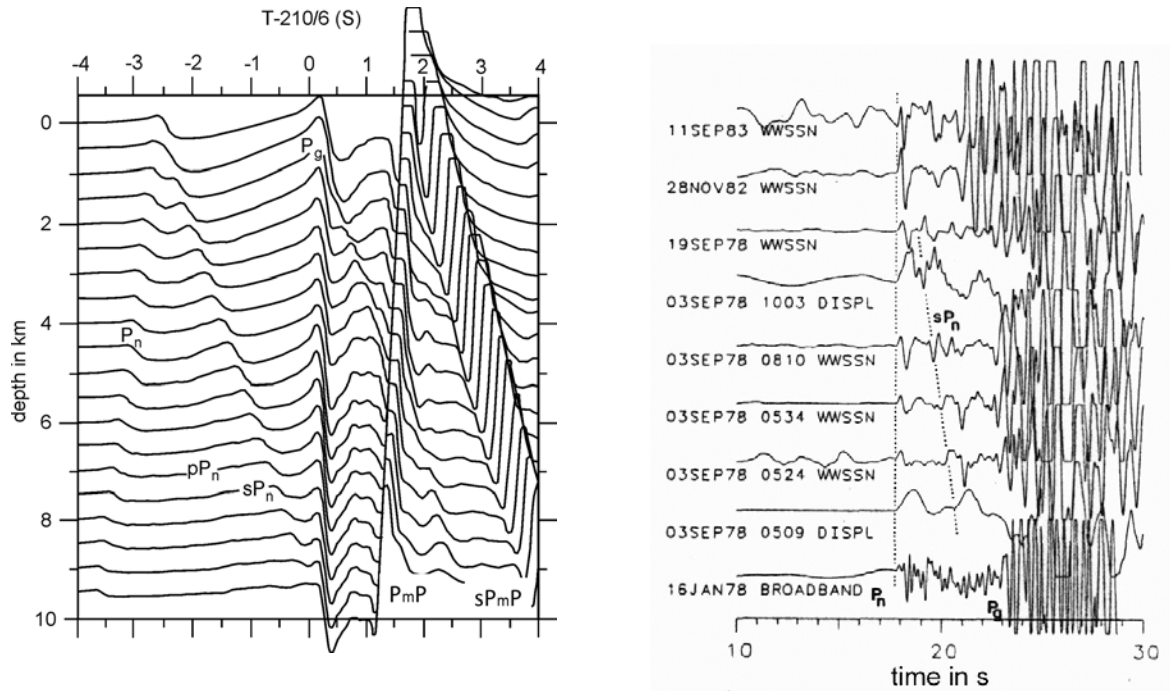
- Since the travel-time curve of PmP approaches that of Pg asymptotically for larger distances, it may be difficult to separate Pg from Pm in real Earth for distances larger than about 100 km (see Fig. 2.67);
- Pn takes over as first arrival beyond about 140 km with generally rather weak amplitudes and higher apparent velocity;
- Sg (and in case of shallow events also surface waves, e.g., Rg) has (have) much larger amplitudes than the various types of direct, refracted or reflected P waves in records of local/regional events;
- The core shadow due to the strongly reduced P-wave velocities in the outer core is indeed clearly developed at epicenter distances between about 100° and 140°, however, long-period diffracted P waves may still be observable as relatively weak first arrivals up to 120° and more;
- PP is the first strong wave arrival in the core shadow range and, if Pdif or the weak inner-side reflections of P from the 660km or 410 km discontinuities (phase names P660-P and P440-P, respectively) are buried in the noise, PP can easily be misinterpreted as P-wave first arrival;
- The caustic of PKP around 145° produces very strong amplitudes comparable to those of P between about 50° to 70°;
- The branching of PKP into three travel-time branches beyond the caustic is well reproduced in the synthetic seismograms;
- Converted core reflections (PcS) and converted core refractions (PKS) may be rather strong secondary later arrivals in the P-wave range between about 35°-55° and in the core-shadow range between about 120°-140°, respectively.

The following figures illustrate the potential of waveform modeling. Depth phases are not only very useful for determining the focal depth from teleseismic records, they are also frequently observed at regional distances and permit accurate depth determinations. Fig. 2.84 shows the ray paths for the phases Pn, pPn, sPn, pPnS and sPmP in a single layer crust from an event at depth  $h$ , as recorded in the distance range beyond 150 km, when Pn appears as the first arrival. Fig. 2.85 (left) shows the theoretical seismograms for these phases at a distance of 210 km and as a function of source depth. It is easy to identify the depth phases. Fig. 2.85 (right) presents a compilation of the summation traces of all available vertical component records of the Gräfenberg array stations for the 1978 Swabian Jura (Germany) earthquake (September 3, 05:09 UT;  $M_L = 6.0$ ) and for several of its aftershocks. All these events have been recorded at an epicenter distance of about 210 km. Depth phases sPn were observed in most records. From the correlation of sPn in neighboring traces it becomes obvious that the source depth migrated within 5 hours from the main shock at  $h = 6.5$  km to a depth of only about 2-3 km for the aftershock at 10:03 UT.



**Fig. 2.84** Ray paths of local depth phases (for synthetics and records see Fig. 2.85).



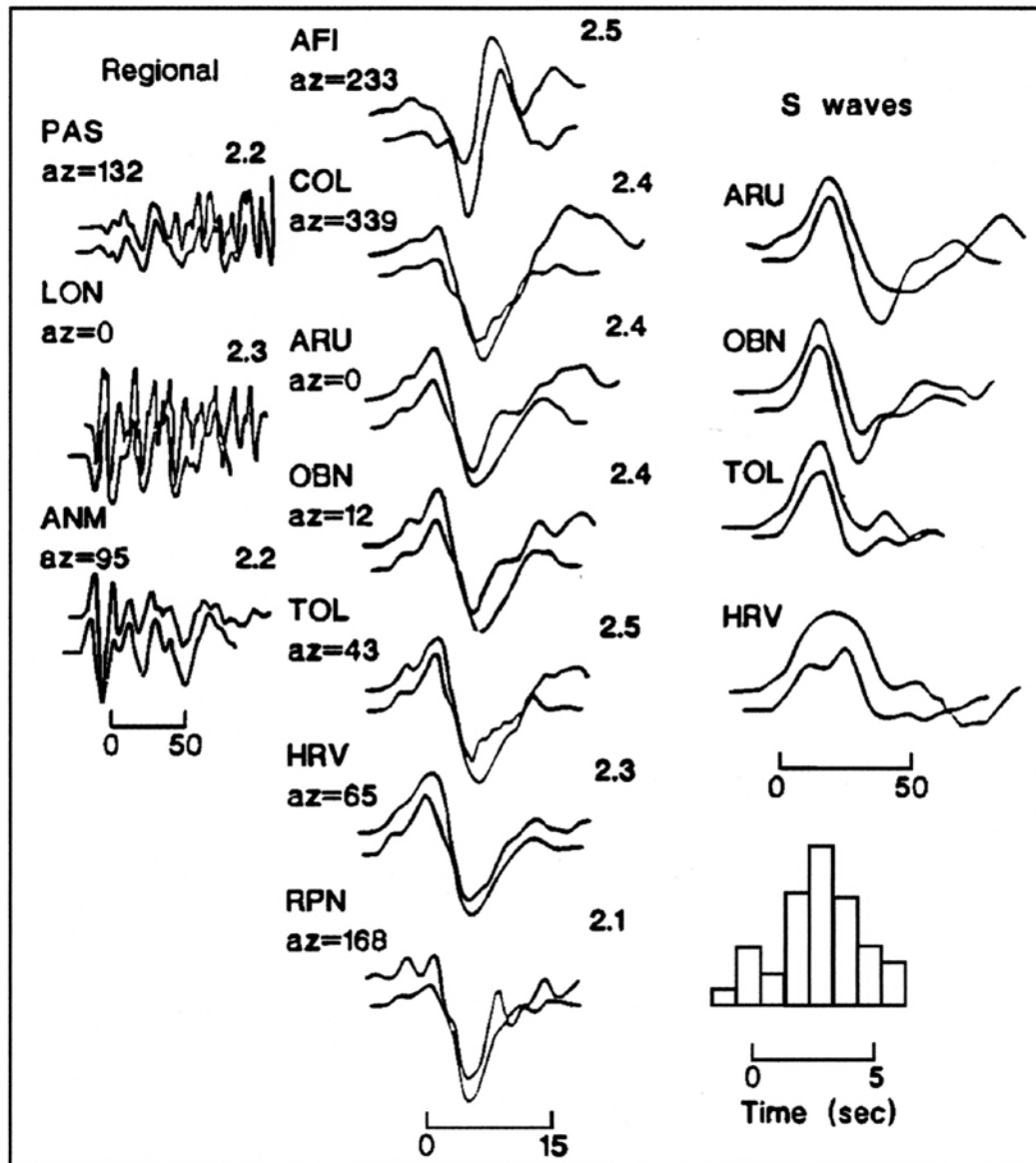


**Fig. 2.85** **Left:** Theoretical seismograms in reduced travel-time presentation at 210 km epicenter distance as function of source depth for a single-layer crust of 30 km thickness. A clear depth phase  $sP_n$  is recognizable between  $P_n$  and  $P_g$ ; **right:** Gräfenberg records of Swabian Jura events in southern Germany. Epicenter distance is 210 km. Between the  $P_n$  and  $P_g$  arrival, a clear depth phase  $sP_n$  can be observed. These observations indicate that after the main shock on September 3 at 05:09 the aftershocks migrated from 6.5 km depth to 2-3 km depth within 5 hours (from Kind, 1985).

Langston and Helmberger (1975) studied the influence of hypocenter depth  $h$ , type of source mechanism, source-time function and of stress drop on seismic waveforms. The superposition of  $P$ ,  $pP$  and  $sP$ , which follow close one after another in the case of crustal earthquakes, make it difficult to separate these individual phases properly in more long-period teleseismic records and to pick the onset times of the depth phases reliably. However, because of the pronounced changes in the waveform of this  $P$ -wave group as a function of depth, one may be able to constrain also the source depth of distant earthquakes rather well by waveform modeling with an accuracy of about 5 km. On the other hand, one should be aware that there is a strong trade-off between source depth and the duration of the source-time function. A deeper source with source function of shorter duration may be similar to a shallower source with a longer source function. For simple sources, broadband data may help to overcome much of this trade-off. For complex source functions, however, these may trade-off with differences in source depth if only data from single stations are available. Using data from several stations instead could reduce this problem.

Generally, waveform modeling is much more powerful than first-motion focal mechanism determinations (see 3.4) in constraining fault orientation. Even with only a few stations and limited azimuthal coverage around the source superior results may be achieved. This is of particular importance for a fast determination of source parameters. Additionally, by comparing predicted and observed amplitudes of waveforms, the seismic moment can be determined rather reliably (see 3.5). Fig. 2.86 shows an example of waveform modeling in the

teleseismic distance range for records of the 1989 Loma Prieta earthquake in different azimuth around the source. From the best fitting synthetics, the source-time function, fault strike  $\phi$ , dip  $\delta$ , rake  $\lambda$  and seismic moment  $M_0$  were estimated. However, Basham and Kind (1986) could show that even with the broadband data from only one teleseismic station good estimates of fault depth, strike, dip and rake could be derived from waveform modeling.



**Fig. 2.86** Results of waveform modeling for the 1989 Loma Prieta earthquake. Depicted are the pairs of observed (top trace) and synthetic waveforms (bottom trace) for long-period Pn (left column), teleseismic P (middle column) and SH waves (right column). The time function used is shown at the lowermost right side. From the inversion of these data the following source parameters were determined:  $\phi = 128^\circ \pm 3^\circ$ ,  $\delta = 66^\circ \pm 4^\circ$ ,  $\lambda = 133^\circ \pm 7^\circ$ , and the moment  $M_0 = 2.4 \times 10^{19}$  Nm (reproduced from Wallace et al., 1991, A broadband seismological investigation of the 1989 Loma Prieta, California, earthquake: Evidence for deep slow slip? Bull. Seism. Soc. Am., Vol. 81, No. 5, Fig. 2, page 1627; 1991; © Seismological Society of America).

## Acknowledgments

The authors thank P. Shearer, P. Malischewsky and J. Schweitzer for careful proof-reading of the first edition and P. Malischewsky also of the amended second edition of this Chapter, and many valuable suggestions. Thanks go also to M. Baumbach, B. L. N. Kennett, D. Di Giacomo, K. Klinge, P. Malischewsky, H. Neunhöfer, P. Richards, J. Ritsema, B. Schurr, J. Schweitzer S. Wendt and K. Wylegalla for making some figures or related data for their production available.

## Recommended overview readings

Aki and Richards (1980 and 2002)  
Bullen and Bolt (1985)  
Chapman (2002)  
Cormier (2009)  
Di Giacomo and Bormann (2011)  
Kennett (2001 and 2002)  
Lay and Wallace (1995)  
Lognonne and Clevede (2002)  
Romanovic and Mitchell (2007)  
Sato et al. (2002)  
Shearer (1999)  
Stein and Wysession, M. (2003)

## References

- Aki, K., and Richards, P. G. (1980). *Quantitative seismology – theory and methods*, Volume II, Freeman and Company, ISBN 0-7167-1059-5(v. 2), 609-625.
- Aki, K., and Richards, P. G. (2002). *Quantitative seismology*. Second Edition, ISBN 0-935702-96-2, University Science Books, Sausalito, CA., xvii + 700 pp.
- Aki, K., and Richards, P. G. (2009). *Quantitative seismology*. 2. ed., corr. Print., Sausalito, Calif., Univ. Science Books, 2009, XVIII + 700 pp., ISBN 978-1-891389-63-4 (see [http://www.ldeo.columbia.edu/~richards/Aki\\_Richards.html](http://www.ldeo.columbia.edu/~richards/Aki_Richards.html)).
- Anderson, D. L., Kanamori, H., Hart, R. S., and Liu, H. (1976). The Earth as a seismic absorption band. *Science*, **196**, 1104-1106.
- Anderson, D. L., and Given, J.W. (1982). Absorption band Q model for the Earth. *J. Geophys. Res.*, **87**, 3893-3904.
- Astiz, L., Earle, P., and Shearer, P. (1996). Global stacking of broadband seismograms. *Seism. Res. Lett.*, **67**, 8-18.
- Båth, M., and Shahidi, M. (1974). T-phase from Atlantic earthquakes. *Pure Appl. Geophys. (PAGEOPH)*, **92**, 74-114.
- Basham, P. W., and Kind, R. (1986). GRF broad-band array analysis of the Miramichi, New Brunswick earthquake sequence. *Journal of Geophysics-Zeitschrift für Geophysik* **60**, 2, 120-128.
- Beauduin, R., Lognonné, P., Montagner, J. P., Cacho, S., Karczewski, J. F., and Morand, M. (1996). The effects of the atmospheric pressure changes on seismic signals or How to improve the quality of a station. *Bull. Seism. Soc. Am.*, **86**, 6, 1760-1769.

- Ben-Menahem, A. and S.J. Singh (2000). *Seismic Waves and Sources*. 2<sup>nd</sup> edition, Dover Publications, New York.
- Benndorf, H. (1905). Über die Art der Fortpflanzung der Erdbebenwellen im Erdinneren. 1. Mitteilung. *Sitzungsberichte der Kaiserlichen Akademie in Wien*. Mathematisch-Naturwissenschaftliche Klasse 114, Mitteilungen der Erdbebenkommission, Neue Folge **29**, 1-42.
- Benndorf, H. (1906). Über die Art der Fortpflanzung der Erdbebenwellen im Erdinneren. 2. Mitteilung. *Sitzungsberichte der Kaiserlichen Akademie in Wien*. Mathematisch-Naturwissenschaftliche Klasse 115, Mitteilungen der Erdbebenkommission, Neue Folge **31**, 1-24.
- Benz, H. M., and Vidale, J. E. (1993). Sharpness of upper-mantle discontinuities determined from high-frequency reflections. *Nature*, **365**, 147-150.
- Benz, H. M., Vidale, J. E., and Mori, J. (1994). Using regional seismic networks to study the Earth's deep interior. *EOS, Trans. Am. Geophys. Un.* **75**, 225, 229.
- Berckhemer, H., Kampfmann, W., Aulbach, E., and Schmeling, H. (1982). Shear module and Q of forsterite and dunite near partial melting from forced-oscillation experiments. *Phys. Earth Planet. Inter.*, **29**, 30-41.
- Bijwaard, H., Spakman, W., and Engdahl, E. R. (1998). Closing the gap between regional and global travel time tomography. *J. Geophys. Res.*, **103**, 30055-30078.
- Boatwright, J. and Fletcher, J. B., 1984. The partition of radiated energy between P and S waves. *Bull. Seismol. Soc. Am.*, **74**: 361-376.
- Boatwright, J. and Quin, H., 1986. The seismic radiation from a 3-D dynamic model of a complex rupture process. Part I: confined ruptures. In: S. Das, J. Boatwright and C. H. Scholz (eds.). *Earthquake source mechanics*. Geophys. Monograph., Am. Geophysical Union, **37**, 97/109.
- Bock, G., Grünthal, G., and Wylegalla, K. (1996). The 1985/86 Western Bohemia earthquakes: Modelling source parameters with synthetic seismograms. *Tectonophysics*, **261**, 139-146.
- Bondár I., S., Myers, C., Engdahl, E. R , and Bergman, E. A. (2004). Epicenter accuracy based on seismic network criteria. *Geophys. J. Int.*, **156**, 483-496.
- Bormann, P. (1969). Some features of seismic body-wave onsets and their consideration in improving source location (in German with extended English abstract: Einige Merkmale von seismischen Raumwelleneinsätzen und deren Berücksichtigung zur Verbesserung der Herdortung;). In: *Veröff. d. Inst. für Geodynamik Jena*, Akademie-Verlag, Berlin, Reihe A, Heft **14**, 21-32.
- Bormann, P., Burghardt, P. T., Makeyeva, L. I., and Vinnik, L. P. (1993). Teleseismic shear-wave splitting and deformations in Central Europe. *Phys. Earth Plant. Int.*, **78**, 157-166.
- Buland, R., and Chapman, C. (1983). The computation of seismic travel times. *Bull. Seism. Soc. Am.*, **73**, 1271-1302.
- Bullen, K. E., and Bolt, B. A. (1985). *An introduction to the theory of seismology*. Fourth Edition, Cambridge University Press, 4<sup>th</sup> revised ed., xvii+499 pp., ISBN: 0-521-28389-2, 451-454, 499 pp..
- Cara, M. (2002). Seismic anisotropy. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 875-885.

- Castagna, J. P., Batzle, M. L., and Eastwood, R. L. (1985). Relationships between compressional-wave and shear-wave velocities in clastic silicate rocks. *Geophysics*, **50**(4), 571-581.
- Červený, V. (2001). *Seismic ray theory*. ISBN 0-521-36671-2, Cambridge University Press, New York, VII + 713 pp.
- Červený, V., Molotkov, I. A., and Pšenčík, I. (1977). *Ray methods in seismology*. Univerzita Karlova, Prague, 214 pp.
- Chapman, C. H. (2002). Seismic ray theory and finite frequency extensions. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 103-123.
- Choy, G. L., and Richards, P. G. (1975). Pulse distortion and Hilbert transformation in multiply reflected and refracted body waves. *Bull. Seism. Soc. Am.*, **65**, 1, 55-70.
- Choy, G. L., and Cormier, V. F. (1986). Direct measurement of the mantle attenuation operator from broadband P and S waves. *J. Geophys. Res.*, **91**, 7326-7342.
- Choy, G. L., and Engdahl, E. R. (1987). Analysis of broadband seismograms from selected IASPEI events. *Phas. Earth Planet. Int.*, **47**, 80-92.
- Cormier, V. F. (2009). Seismic viscoelastic attenuation. In: Gupta, H. (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 1279-1290; doi: 10.1007/978-90-481-8702-7\_55.
- Curtis, A., and Snieder, R. (2002). Probing the Earth's with seismic tomography. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 861-874.
- Dahlen, F. A., and Tromp, J. (1998). *Theoretical global seismology*. Princeton University Press.
- Der, Z. A., McElfresh, Th. W., and O'Donnell, A. (1982). An investigation of the regional variations and frequency-dependence of anelastic attenuation in the mantle under the United States in the 0.5-4 Hz band. *Geophys. J. R. astr. Soc.*, **69**, 68-99.
- Di Giacomo, D., and Bormann, P. (2011). Earthquake energy. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 233-236; doi: 10.1007/978-90-481-8702-7.
- Duda, S. J., and Yanovskaya, T. B. (1993). Spectral amplitude-distance curves for P-waves: effects of velocity and Q-distribution. *Tectonophysics*, **217**, 255-265.
- Dziewonski, A. M., and Anderson, D. L. (1981). Preliminary reference Earth model. *Phys. Earth Planet. Inter.*, **25**, 297-356.
- Dziewonski, A. M., and Anderson, D. L. (1983). Travel times and station corrections for P waves at teleseismic distances. *J. of Geophys. Res.*, **88**, 3295-3314.
- Dziewonski, A. M., Hager, B. H., and O'Connell, R. (1977). Large-scale heterogeneities in the lower mantle. *J. Geophys. Res.*, **82**, 239-255.
- Engdahl, E. R., and Billington, S. (1986). Focal depth determination of central Aleutian earthquakes. *Bull. Seism. Soc. Am.*, **76**, 77-93.
- Engdahl, E. R., and Kind, R. (1986). Interpretation of broadband seismograms from central Aleutian earthquakes. *Annales Geophysicae*, **4**, B(3), 233-240.

- Engdahl, E. R., Van der Hilst, R. D., and Buland, R. P. (1998). Global teleseismic earthquake relocation with improved travel times and procedures for depth determination. *Bull. Seism. Soc. Am.*, **88**, 722-743.
- Fuchs, K., and Müller, G. (1971). Computation of synthetic seismograms with the reflectivity method and comparison to observations. *Geophys. J. R. astr. Soc.*, **23**, 417-433.
- Fukao, Y., Widiyantoro, S., and Obayashi, M. (2001). Stagnant slabs in the upper and lower mantle transition region. *Rev. Geophys.*, **39**, 291-323.
- Galitzin, B. B. (1914). Vorlesungen über Seismometrie. *Verlag Teubner*, Berlin 1914, VIII + 538 pp.
- Gilbert, F., and Dziewonski, A. M. (1975). An application of normal mode theory to the retrieval of structural parameters and source mechanisms from seismic spectra. *Philos. Trans. R. Soc. London, Ser. A* **278**, 187-269.
- Gossler, J., and Kind, R. (1996). Seismic evidence for very deep roots of continents. *Earth and Planet. Sci. Lett.*, **138**, 1-13.
- Haskell, N. A. (1953). The dispersion of surface waves on multilayered media. *Bull. Seism. Soc. Am.*, **43**, 17-43.
- Hayes, G. P., Rivera, L., and Kanamori, H. (2009). Source inversion of the W-phase: Real-time implementation and extension to low magnitudes. *Seism. Res. Lett.*, **80**(5), 817-822.
- IASPEI (2011) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf)
- IASPEI (2011). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf)
- Jeffreys, H., and Bullen, K. E. (1940, 1948, 1958, 1967, and 1970). *Seismological Tables*. British Association for the Advancement of Science, Gray Milne Trust, London, 50 pp.
- Kanamori, H. (1993). W phase. *Geophys. Res. Lett.*, **20**(16), 1691-1694.
- Kanamori, H., and Rivera, L. (2008). Source inversion of W phase: speeding up seismic tsunami warning. *Geophys. J. Int.*, **175**, 222-238.
- Kennett, B. L. N. (1983). Seismic wave propagation in stratified media. ISBN 0-521-23933-8 *Cambridge University Press*, Cambridge, viii + 242 pp.
- Kennett, B. L. N. (1988). Systematic approximations to the seismic wavefields. In: Doornbos (1988a), 237-259.
- Kennett, B. L. N. (Ed.) (1991). *IASPEI 1991 Seismological Tables*. Research School of Earth Sciences, Australian National University, 167 pp.
- Kennett, B. L. N. (2001). *The seismic wavefield. Volume I: Introduction and theoretical development*. Cambridge University Press, Cambridge, x + 370 pp.
- Kennett, B. L. N. (2002). *The seismic wavefield. Volume II: Interpretation of seismograms on regional and global scales*. Cambridge University Press, Cambridge, x + 534 pp, ISBN 0-521-00665-1, 163-170.
- Kennett, B. L. N., and Engdahl, E. R. (1991). Travel-times for global earthquake location and phase identification. *Geophys. J. Int.*, **105**, 429-465.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, **122**, 108-124.

- Kim, W.-Y., Simpson, D. W., and Richards, P. G., 1993. Discrimination of earthquakes and explosions in the Eastern United States using regional high-frequency data. *Geophysical Research Letters*, **20**, 1507–1510.
- Kind, R., (1978). The reflectivity method for a buried source. *J. Geophys.*, **44**, 603-612.
- Kind, R., (1985). The reflectivity method for different source and receiver structures and comparison with GRF data. *J. Geophys.*, **58**, 146-152.
- Kulháněk, O. (1990). Anatomy of seismograms. *Developments in Solid Earth Geophysics* **18**, Elsevier, Amsterdam, 78 pp.
- Langston, Ch. A., and Helmberger, D. V. (1975). A procedure for modeling shallow dislocation sources. *Geophys. J. R. astr. Soc.*, **42**, 117-130.
- Lapwood, E. R., and Usami, T. (1981). *Free oscillations of the Earth*. Cambridge University Press, Cambridge, UK.
- Lay, T., and Wallace, T. C. (1995). *Modern global seismology*. ISBN 0-12-732870-X, Academic Press, 521 pp.
- Lillie, R. J. (1999). *Whole Earth Geophysics*. Prentice-Hall Inc., New Jersey, ISBN 0-13 4990517-2, 361 pp.
- Linehan, D. (1940). Earthquakes in the West Indian region. *Trans. Am. Geophys. Union, EOS* **30**, 229-232.
- Liu, H.-P., Anderson, D. L., and Kanamori, H. (1976). Velocity dispersion due to anelasticity; implications for seismology and mantle composition. *Geophys. J. R. Astron. Soc.*, **47**, 41-58.
- Lognonné, P., and Clévéde, E. (2002). Normal modes of the Earth and planets. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 125-147.
- Malischewsky, P. (1987). *Surface waves and discontinuities*. Elsevier, Amsterdam, ISBN 0-444-98959-5, 229 pp.
- Malischewsky Auning, P. G. (2004). A note on Rayleigh-wave velocities as a function of the material parameters. *Geofísica Internacional*, **43**, 507-509.
- Malischewsky Auning, P. G., Lomnitz, C., Wuttke, F., Saragoni, R. (2006). Prograde Rayleigh-wave motion in the valley of Mexico. *Geofísica Internacional*, **45**, 149-162.
- Malischewsky, P. G., Scherbaum, F., Lomnitz, C., Tran Thanh Tuan, Wuttke, F., Shamir, G. (2008). The domain of existence of prograde Rayleigh-wave particle motion for simple models. *Wave Motion*, **45**, 556-564.
- Malischewsky, P. (2011). Seismic waves and surface waves: past and present. *Geofísica Internacional*, **SO-4**, 485-493.
- Marquering, H., Dahlen, F. A., and Nolet, G. (1999). Three-dimensional sensitivity kernels for finite-frequency traveltimes: the banana-doughnut paradox. *Geophys. J. Int.*, **137**, 805-815.
- Masters, G., Jordan, T. H., Silver, P. G., and Gilbert F. (1982). Aspherical Earth structure from fundamental spheroidal mode data. *Nature*, **298**, 609–613.
- Meier, T., Malischewsky, P. G., and Neunhöfer, H. (1997). Reflection and transmission of surface waves at a vertical discontinuity and imaging of lateral heterogeneity using reflected fundamental Rayleigh waves. *Bull. Seism. Soc. Am.*, **87**, 1648-1661.
- Mendiguren, J. A. (1971). Focal mechanism of a shock in the middle of Nazca plate. *J. Geophys. Res.*, **76**, 3861-3879.



- Minshull, T. A. (2002). Seismic structure of the oceanic crust and passive continental margins. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 911-924.
- Montagner, J.-P., and Tanimoto, T. (1991). Global upper mantle tomography of seismic velocities and anisotropies. *J. Geophys. Res.*, **96**, 20,337–20,351.
- Montagner, J.-P., and Kennett, B. L. N. (1996). How to reconcile body-wave and normal-mode reference Earth models? *Geophys. J. Int.*, **125**, 229-248.
- Montelli, R., Nolet, G., Dahlen, F., Masters, G., Engdahl, E. R., and Hung, S. (2004). Finite-frequency tomography reveals a variety of plumes in the mantle. *Science*, **303**, 338–343.
- Mooney, W. D., Prodehl, C., and Pavlenkova, N. I. (2002). Seismic velocity structure of the continental lithosphere from controlled source data. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 887-912.
- Morelli, A., and Dziewonski, A. M. (1993). Body wave traveltimes and a spherically symmetric P- and S-wave velocity model. *Geophys. J. Int.*, **112**, 178-194.
- Müller, G. (1977). Earth flattening approximation for body waves derived from geometric ray theory - improvements, corrections and range of applicability. *J. Geophys.*, **44**, 429-436.
- Müller, G. (1985). The reflectivity method: A tutorial. *J. Geophys.*, **58**, 153-174.
- Neunhöfer, H. (1985). Primary and secondary effects of surface wave propagation and their geophysical causes (in German). *Veröff. Zentralinstitut Phys. Erde* **No. 85**, AdW der DDR, Potsdam, 111 pp.
- Nolet, G. (1987). *Seismic tomography: With applications in global seismology and exploration geophysics*. Reidel, Dordrecht, 386 pp.
- Nolet, G. (2008). *A breviary of seismic tomography: Imaging the interior of the Earth and Sun*. Cambridge University Press, 344 pp.
- Okal, E. A. (1992a). Use of the mantle magnitude  $M_m$  for the reassessment of the moment of historical earthquakes. I: Shallow events. *Pageoph*, **139**, 1, 17-57.
- Okal, E. A. (1992b). Use of the mantle magnitude  $M_m$  for the reassessment of the moment of historical earthquakes. II: Intermediate and deep events. *Pageoph*, **139**, 1, 59-85.
- Okal, E. A., and Talandier, J. (1989).  $M_m$ : A variable-period mantle magnitude. *J. Geophys. Res.*, **94**, 4169-4193.
- Okal, E. A., Alsset, P.-J., Hyvernaud, O., and Schindelé, F. (2003). The deficient T waves of tsunami earthquakes. *Geophys. J. Int.*, **152**, 416-432.
- Payo, G. (1986). *Introducción al análisis de sismogramas*. Ministerio de la Presidencia, Instituto Geográfico Nacional, Madrid, Monographías No. 3, 125 pp.
- Pickett, G. R. (1963). Acoustic character logs and their applications in formation evaluation. *J. Petr. Tech.*, **15**, 650-667.
- Rautian, T., and V. I. Khalturin (1978). The use of coda for determination of the earthquake source spectrum. *Bull. Seism. Soc. Am.*, **68**, 904-922.
- Rautian, T., Khalturin, V., Fujita, K., Mackey, K., and Kendall, A. (2007). Origins and methodology of the Russian energy K-class system and its relationship to magnitude scales. *Seism. Res. Lett.*, **78**, 579-590.
- Rayleigh, Lord (J. W. S.) (1887). On waves propagated along the plane surface of an elastic solid. *Proc. London Math. Soc.*, **17**, 4-11.

- Richards, P. G., and Wu, Z. (2011). Seismic monitoring of nuclear explosions. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 1144-1156; doi: 10.1007/978-90-481-8702-7.
- Romanowicz, B. (2002). Inversion of surface waves: A review. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 149-173.
- Romanowicz, B., Mitchell, B. (2007). Deep Earth structure: Q of the Earth from crust to core. In: Schubert, G. (Ed.), *Treatise on Geophysics*, **1**, Elsevier, 731-774.
- Sandmeier, K.-J., (1990). *Untersuchung der Ausbreitungseigenschaften seismischer Wellen in geschichteten und streuenden Medien*. PhD Thesis, University of Karlsruhe
- Sawarenski, E. F., Kirnos, D. P. (1960). *Elemente der Seismologie und Seismometrie*. Berlin, Akad.-Verl., XIII, 512 pp.
- Shearer, P. M. (1999). *Introduction to seismology*. Cambridge University Press, 260 pp.
- Shearer, P. M., and Flanagan, M. P. (1999). Seismic velocity and density jumps across the 410- and 660-km discontinuities. *Science*, **285**, 1545-1548; doi: 10.1126/science. 285. 5433.1545.
- Sheriff, R. E., and Geldart, L. P. (1995). *Exploration seismology*. Cambridge University Press, Second Edition, 592 pp.
- Silver, P. G., and Chan, W. W. (1991). Shear-wave splitting and mantle deformation. *J. Geophys. Res.*, **96**, 16429-16454.
- Sinha, N. K. (1987). Effective Poisson's ratio of isotropic ice. *Proceed. 6<sup>th</sup> International Offshore Mechanics and Arctic Engineering Symposium*. Houston, IRC Paper No. 1472, Vol. IV, 189-195.
- Song, X. (2002). The Earth's core. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*. Academic Press, Amsterdam, 925-933.
- Stein, S., and Wysession, M. (2003). *An introduction to seismology, earthquakes and Earth structure*. Blackwell Publishing Ltd., Oxford (UK), 498 pp.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2003). The IASPEI Standard Seismic Phase List. *Seism. Res. Lett.*, **74**, 761-772.
- Storchak, D. A., Schweitzer, J. and Bormann, P. (2011). Seismic phase names: IASPEI standards. In: Gupta, H. (2011). *Encyclopedia of Solid Earth Geophysics*, Vol. 2, Springer, 1162-1173.
- Thomson, S. T. (1950). Transmission of elastic waves through a stratified solid medium. *J. Appl. Phys.*, **21**, 89-93.
- Thurber, C. H., and Engdahl, E. R. (2000). Advances in global seismic event location. In: in Thurber, T. C., and Rabinowitz, N. (2000), 3-22.
- Thurber, T. C., and Ritsema, J. (2007). Theory and observation – seismic tomography and inverse methods. In: Schubert, G. (ed.). *Treatise on Geophysics*, Elsevier Ltd., Oxford, Vol. 1, 323-360.
- Ulug, A. and Berckhemer, H. (1984). Frequency dependence of Q for seismic body waves in the Earth's mantle. *J. Geophys.*, **56**, 9-19.
- van der Hilst, R.D., Widiyantoro, S., and Engdahl, R. (1997). *Nature*, **386**, 578-584.
- Vinnik, L. P., Farra, F., and Romanowicz, B. (1989a). Azimuthal anisotropy in the Earth from observations of SKS at Geoscope and NARS broadband station. *Bull. Seism. Soc. Am.*, **79**, 1542-1558.

- Vinnik, L. P., Kind, R., Kosarev, G. L., and Makeyeva, L. I. (1989b). Azimuthal anisotropy in the lithosphere from observations of long period S waves. *Geophys. J. Int.*, **99**, 549-559.
- Wang, R. (1999). A simple orthonormalization method for stable and efficient computation of Green's functions. *Bull. Seism. Soc. Am.*, **89**, 3, 733-741.
- Warburton, R. J., and Goodkind, J. M. (1977). The influence of barometric pressure variations on gravity. *Geophys. J. R. astr. Soc.* **48**, 281-292.
- Wiechert, E., und Geiger, L. (1910). Bestimmung des Weges der Erdbebenwellen im Erdinnern. *Physikalische Zeitschrift*, **11**, 294-311.
- Yoshii, T. (1979). A detailed cross-section of the deep seismic zone beneath northeastern Honshu, Japan. *Tectonophysics*, **55**, 349-360.
- Zoeppritz, K. (1907). Über Erdbebenwellen. II. Laufzeitkurven. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*, Mathematisch-physikalische Klasse, 529-549.

| Topic       | Global 1-D Earth models                                                                                                                                                                  |
|-------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Compiled by | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, Potsdam, Telegrafenberg, D-14473 Potsdam, Germany);<br>E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version     | October 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_DS_2.1">10.2312/GFZ.NMSOP-2_DS_2.1</a>                                                                                   |

Below, data and background information on the most frequently used 1-D Earth reference models in global seismology have been compiled by using Appendix 1 of Shearer (1999), Kennett (1991) and personal information received from B. Kennett (2002).

## 1 PREM Model

For many years the most widely used 1-D model of seismic velocities in the Earth has been the Preliminary Reference Earth Model (PREM) of Dziewonski and Anderson (1981). This model was designed to fit a variety of different data sets, including free oscillation center frequency measurements, surface-wave dispersion observations, travel-time data for a number of body-wave phases, and basic astronomical data (Earth's radius, mass, and moment of inertia). Table 1 summarizes, as functions of depth and Earth's radius, the PREM velocities  $v_p$  and  $v_s$  for P and S waves, the density  $\rho$ , the shear and bulk quality factors,  $Q_\mu$  and  $Q_\kappa$ , and the pressure  $P$ . Note, that density and attenuation ( $\sim 1/Q$ ) are known less precisely than the seismic velocities but these parameters are required for computing synthetic seismograms. In order to simultaneously fit Love- and Rayleigh-wave observations, PREM is transversely isotropic between 80 and 220 km depth in the upper mantle. Transverse isotropy is a spherically symmetric form of anisotropy in which SH and SV waves travel at different speeds. Table 1, however, lists only values from an isotropic version of PREM. The true PREM model is also specified in terms of polynomials between node points. Linear interpolation between the values given in Table 1 will produce only approximate results. All current Earth models have values that are reasonably close to PREM. The largest differences are in the upper mantle where PREM shows a discontinuity at 220 km which is not found in most other models. Fig. 2.53 in Chapter 2 depicts PREM together with the more recent model AK135 (see Table 3 below).

**Table 1 Preliminary Reference Earth Model (isotropic version)**

| Depth (km) | Radius (km) | $v_p$ (km/s) | $v_s$ (km/s) | $\rho$ (g/cm <sup>3</sup> ) | $Q_\mu$ | $Q_\kappa$ | $P$ (GPa) |
|------------|-------------|--------------|--------------|-----------------------------|---------|------------|-----------|
| 0.0        | 6371.0      | 1.45         | 0.00         | 1.02                        | 0.0     | 57823.0    | 0.0       |
| 3.0        | 6368.0      | 1.45         | 0.00         | 1.02                        | 0.0     | 57823.0    | 0.0       |
| 3.0        | 6368.0      | 5.80         | 3.20         | 2.60                        | 600.0   | 57823.0    | 0.0       |
| 15.0       | 6356.0      | 5.80         | 3.20         | 2.60                        | 600.0   | 57823.0    | 0.3       |
| 15.0       | 6356.0      | 6.80         | 3.90         | 2.90                        | 600.0   | 57823.0    | 0.3       |
| 24.4       | 6346.6      | 6.80         | 3.90         | 2.90                        | 600.0   | 57823.0    | 0.6       |
| 24.4       | 6346.6      | 8.11         | 4.49         | 3.38                        | 600.0   | 57823.0    | 0.6       |
| 71.0       | 6300.0      | 8.08         | 4.47         | 3.38                        | 600.0   | 57823.0    | 2.2       |
| 80.0       | 6291.9      | 8.08         | 4.47         | 3.37                        | 600.0   | 57823.0    | 2.5       |
| 80.0       | 6291.0      | 8.08         | 4.47         | 3.37                        | 80.0    | 57823.0    | 2.5       |
| 171.0      | 6200.0      | 8.02         | 4.44         | 3.36                        | 80.0    | 57823.0    | 5.5       |
| 220.0      | 6151.0      | 7.99         | 4.42         | 3.36                        | 80.0    | 57823.0    | 7.1       |
| 220.0      | 6151.0      | 8.56         | 4.62         | 3.44                        | 143.0   | 57823.0    | 7.1       |
| 271.0      | 6100.0      | 8.66         | 4.68         | 3.47                        | 143.0   | 57823.0    | 8.9       |

**Table 1 (continued)**

| Depth<br>(km) | Radius<br>(km) | $v_p$<br>(km/s) | $v_s$<br>(km/s) | $\rho$<br>(g/cm <sup>3</sup> ) | $Q_\mu$ | $Q_\kappa$ | $P$<br>(GPa) |
|---------------|----------------|-----------------|-----------------|--------------------------------|---------|------------|--------------|
| 371.0         | 6000.0         | 8.85            | 4.75            | 3.53                           | 143.0   | 57823.0    | 12.3         |
| 400.0         | 5971.0         | 8.91            | 4.77            | 3.54                           | 143.0   | 57823.0    | 13.4         |
| 400.0         | 5971.0         | 9.13            | 4.93            | 3.72                           | 143.0   | 57823.0    | 13.4         |
| 471.0         | 5900.0         | 9.50            | 5.14            | 3.81                           | 143.0   | 57823.0    | 16.0         |
| 571.0         | 5800.0         | 10.01           | 5.43            | 3.94                           | 143.0   | 57823.0    | 19.9         |
| 600.0         | 5771.0         | 10.16           | 5.52            | 3.98                           | 143.0   | 57823.0    | 21.0         |
| 600.0         | 5771.0         | 10.16           | 5.52            | 3.98                           | 143.0   | 57823.0    | 21.0         |
| 670.0         | 5701.0         | 10.27           | 5.57            | 3.99                           | 143.0   | 57823.0    | 23.8         |
| 670.0         | 5701.0         | 10.75           | 5.95            | 4.38                           | 312.0   | 57823.0    | 23.8         |
| 771.0         | 5600.0         | 11.07           | 6.24            | 4.44                           | 312.0   | 57823.0    | 28.3         |
| 871.0         | 5500.0         | 11.24           | 6.31            | 4.50                           | 312.0   | 57823.0    | 32.8         |
| 971.0         | 5400.0         | 11.42           | 6.38            | 4.56                           | 312.0   | 57823.0    | 37.3         |
| 1071.0        | 5300.0         | 11.58           | 6.44            | 4.62                           | 312.0   | 57823.0    | 41.9         |
| 1171.0        | 5200.0         | 11.78           | 6.50            | 4.68                           | 312.0   | 57823.0    | 46.5         |
| 1271.0        | 5100.0         | 11.88           | 6.56            | 4.73                           | 312.0   | 57823.0    | 51.2         |
| 1371.0        | 5000.0         | 12.02           | 6.62            | 4.79                           | 312.0   | 57823.0    | 55.9         |
| 1471.0        | 4900.0         | 12.16           | 6.67            | 4.84                           | 312.0   | 57823.0    | 60.7         |
| 1571.0        | 4800.0         | 12.29           | 6.73            | 4.90                           | 312.0   | 57823.0    | 65.5         |
| 1671.0        | 4700.0         | 12.42           | 6.78            | 4.95                           | 312.0   | 57823.0    | 70.4         |
| 1771.0        | 4600.0         | 12.54           | 6.83            | 5.00                           | 312.0   | 57823.0    | 75.4         |
| 1871.0        | 4500.0         | 12.67           | 6.87            | 5.05                           | 312.0   | 57823.0    | 80.4         |
| 1971.0        | 4400.0         | 12.78           | 6.92            | 5.11                           | 312.0   | 57823.0    | 85.5         |
| 2071.0        | 4300.0         | 12.90           | 6.97            | 5.16                           | 312.0   | 57823.0    | 90.6         |
| 2171.0        | 4200.0         | 13.02           | 7.01            | 5.21                           | 312.0   | 57823.0    | 95.8         |
| 2271.0        | 4100.0         | 13.13           | 7.06            | 5.26                           | 312.0   | 57823.0    | 101.1        |
| 2371.0        | 4000.0         | 13.25           | 7.10            | 5.31                           | 312.0   | 57823.0    | 106.4        |
| 2471.0        | 3900.0         | 13.36           | 7.14            | 5.36                           | 312.0   | 57823.0    | 111.9        |
| 2571.0        | 3800.0         | 13.48           | 7.19            | 5.41                           | 312.0   | 57823.0    | 117.4        |
| 2671.0        | 3700.0         | 13.60           | 7.23            | 5.46                           | 312.0   | 57823.0    | 123.0        |
| 2741.0        | 3630.0         | 13.68           | 7.27            | 5.49                           | 312.0   | 57823.0    | 127.0        |
| 2771.0        | 3600.0         | 13.69           | 7.27            | 5.51                           | 312.0   | 57823.0    | 128.8        |
| 2871.0        | 3500.0         | 13.71           | 7.26            | 5.56                           | 312.0   | 57823.0    | 134.6        |
| 2891.0        | 3480.0         | 13.72           | 7.26            | 5.57                           | 312.0   | 57823.0    | 135.8        |
| 2891.0        | 3480.0         | 8.06            | 0.00            | 9.90                           | 0.0     | 57823.0    | 135.8        |
| 2971.0        | 3400.0         | 8.20            | 0.00            | 10.03                          | 0.0     | 57823.0    | 144.2        |
| 3071.0        | 3300.0         | 8.36            | 0.00            | 10.18                          | 0.0     | 57823.0    | 154.8        |
| 3171.0        | 3200.0         | 8.51            | 0.00            | 10.33                          | 0.0     | 57823.0    | 165.2        |
| 3271.0        | 3100.0         | 8.66            | 0.00            | 10.47                          | 0.0     | 57823.0    | 175.5        |
| 3371.0        | 3000.0         | 8.80            | 0.00            | 10.60                          | 0.0     | 57823.0    | 185.7        |
| 3471.0        | 2900.0         | 8.93            | 0.00            | 10.73                          | 0.0     | 57823.0    | 195.8        |
| 3571.0        | 2800.0         | 9.05            | 0.00            | 10.85                          | 0.0     | 57823.0    | 205.7        |
| 3671.0        | 2700.0         | 9.17            | 0.00            | 10.97                          | 0.0     | 57823.0    | 215.4        |
| 3771.0        | 2600.0         | 9.28            | 0.00            | 11.08                          | 0.0     | 57823.0    | 224.9        |
| 3871.0        | 2500.0         | 9.38            | 0.00            | 11.19                          | 0.0     | 57823.0    | 234.2        |
| 3971.0        | 2400.0         | 9.48            | 0.00            | 11.29                          | 0.0     | 57823.0    | 243.3        |
| 4071.0        | 2300.0         | 9.58            | 0.00            | 11.39                          | 0.0     | 57823.0    | 252.2        |

**Table 1 (continued)**

| Depth<br>(km) | Radius<br>(km) | $v_p$<br>(km/s) | $v_s$<br>(km/s) | $\rho$<br>(g/cm <sup>3</sup> ) | $Q_\mu$ | $Q_\kappa$ | $P$<br>(GPa) |
|---------------|----------------|-----------------|-----------------|--------------------------------|---------|------------|--------------|
| 4171.0        | 2200.0         | 9.67            | 0.00            | 11.48                          | 0.0     | 57823.0    | 260.8        |
| 4271.0        | 2100.0         | 9.75            | 0.00            | 11.57                          | 0.0     | 57823.0    | 269.1        |
| 4371.0        | 2000.0         | 9.84            | 0.00            | 11.65                          | 0.0     | 57823.0    | 277.1        |
| 4471.0        | 1900.0         | 9.91            | 0.00            | 11.73                          | 0.0     | 57823.0    | 284.9        |
| 4571.0        | 1800.0         | 9.99            | 0.00            | 11.81                          | 0.0     | 57823.0    | 292.3        |
| 4671.0        | 1700.0         | 10.06           | 0.00            | 11.88                          | 0.0     | 57823.0    | 299.5        |
| 4771.0        | 1600.0         | 10.12           | 0.00            | 11.95                          | 0.0     | 57823.0    | 306.2        |
| 4871.0        | 1500.0         | 10.19           | 0.00            | 12.01                          | 0.0     | 57823.0    | 312.7        |
| 4971.0        | 1400.0         | 10.25           | 0.00            | 12.07                          | 0.0     | 57823.0    | 318.9        |
| 5071.0        | 1300.0         | 10.31           | 0.00            | 12.12                          | 0.0     | 57823.0    | 324.7        |
| 5149.5        | 1221.5         | 10.36           | 0.00            | 12.17                          | 0.0     | 57823.0    | 329.0        |
| 5149.5        | 1221.5         | 11.03           | 3.50            | 12.76                          | 84.6    | 57823.0    | 329.0        |
| 5171.0        | 1200.0         | 11.04           | 3.51            | 12.77                          | 84.6    | 57823.0    | 330.2        |
| 5271.0        | 1100.0         | 11.07           | 3.54            | 12.82                          | 84.6    | 57823.0    | 335.5        |
| 5371.0        | 1000.0         | 11.11           | 3.56            | 12.87                          | 84.6    | 57823.0    | 340.4        |
| 5471.0        | 900.0          | 11.14           | 3.58            | 12.91                          | 84.6    | 57823.0    | 344.8        |
| 5571.0        | 800.0          | 11.16           | 3.60            | 12.95                          | 84.6    | 57823.0    | 348.8        |
| 5671.0        | 700.0          | 11.19           | 3.61            | 12.98                          | 84.6    | 57823.0    | 352.2        |
| 5771.0        | 600.0          | 11.21           | 3.63            | 13.01                          | 84.6    | 57823.0    | 355.4        |
| 5871.0        | 500.0          | 11.22           | 3.64            | 13.03                          | 84.6    | 57823.0    | 358.0        |
| 5971.0        | 400.0          | 11.24           | 3.65            | 13.05                          | 84.6    | 57823.0    | 360.2        |
| 6071.0        | 300.0          | 11.25           | 3.66            | 13.07                          | 84.6    | 57823.0    | 361.8        |
| 6171.0        | 200.0          | 11.26           | 3.66            | 13.08                          | 84.6    | 57823.0    | 363.0        |
| 6271.0        | 100.0          | 11.26           | 3.67            | 13.09                          | 84.6    | 57823.0    | 363.7        |
| 6371.0        | 0.0            | 11.26           | 3.67            | 13.09                          | 84.6    | 57823.0    | 364.0        |

## 2 IASP91 velocity model

According to Kennett (1991) and Kennett and Engdahl (1991), the IASP91 model is a parameterized velocity model, in terms of normalized radius. It has been constructed to be a summary of the travel-time characteristics of the main seismic phases.

The crust consists of two uniform layers with discontinuities at 20 and 35 km. Between 35 and 760 km the velocities in each layer are represented by a linear gradient in radius. The major mantle discontinuities are set at 410 km and 660 km. The upper mantle model is designed for the specific purpose of representing the 'average' observed times for P and S waves out to 30° as well as providing a tie to teleseismic times. Since the distribution of seismic sources and recording stations is far from uniform, the IASP91 model will include geographical bias as well as the constraints imposed by the specific parameterization.

The distribution of P- and S-wave velocities,  $v_p$  and  $v_s$ , in the lower mantle is represented by a cubic radius between 760 km and 2740 km. The velocities in the lowermost mantle are taken as a linear gradient in radius down to the core-mantle boundary at 3482 km. In the core and inner core the velocity functions are specified as quadratic polynomials in radius.

Table 2.1 presents the parameterized form and Table 2.2 the tabulated form of the IASP91.

**Table 2.1 Parameterized form of the IASP91 model**

( $x$  = normalised radius  $r/a$  where  $a = 6371$  km)

| Depth<br>(z km) | Radius<br>(r km) | $v_p$<br>(km/s) | $v_s$<br>(km/s) |
|-----------------|------------------|-----------------|-----------------|
| 6371-5153.9     | 0-1217.1         | 11.24094        | 3.56454         |
|                 |                  | $-4.09689 x^2$  | $-3.45241 x^2$  |
| 5153.9-2889     | 1217.1-3482      | 10.03904        | 0               |
|                 |                  | $3.75665 x$     |                 |
|                 |                  | $-13.67046 x^2$ |                 |
| 2889-2740       | 3482-3631        | 14.49470        | 816616          |
|                 |                  | $-1.47089 x$    | $-1.58206 x$    |
| 2740-760        | 3631-5611        | 25.1486         | 12.9303         |
|                 |                  | $-41.1538 x$    | $-21.2590 x$    |
|                 |                  | $+51.9932 x^2$  | $+27.8988 x^2$  |
|                 |                  | $-26.6083 x^3$  | $-14.1080 x^3$  |
| 760-660         | 5611-5711        | 25.96984        | 20.76890        |
|                 |                  | $-16.93412 x$   | $-16.53147 x$   |
| 660-410         | 5711-5961        | 29.38896        | 17.70732        |
|                 |                  | $-21.40656 x$   | $-13.50652$     |
| 410-210         | 5961-6161        | 30.78765        | 15.24213        |
|                 |                  | $-23.25415 x$   | $-11.08552$     |
| 210-120         | 6161-6251        | 25.41389        | 5.75020         |
|                 |                  | $-17.69722 x$   | $-1.27420$      |
| 120-35          | 6251-6336        | 8.78541         | 6.706231        |
|                 |                  | $-0.74953 x$    | $-2.248585$     |
| 35-20           | 6336-6351        | 6.50            | 3.75            |
| 20-0            | 6351-6371        | 5.80            | 3.36            |



**Table 2.2 The IASP91 velocity model**

| Depth<br>(km) | Radius<br>(km) | $v_p$<br>(km/s) | $v_s$<br>(km/s) |
|---------------|----------------|-----------------|-----------------|
| 6371.00       | 0.             | 11.2409         | 3.5645          |
| 6271.00       | 100.000        | 11.2399         | 3.5637          |
| 6171.00       | 200.000        | 11.2369         | 3.5611          |
| 6071.00       | 300.000        | 11.2319         | 3.5569          |
| 5971.00       | 400.000        | 11.2248         | 3.5509          |
| 5871.00       | 500.000        | 11.2157         | 3.5433          |
| 5771.00       | 600.000        | 11.2046         | 3.5339          |
| 5671.00       | 700.000        | 11.1915         | 3.5229          |
| 5571.00       | 800.000        | 11.1763         | 3.5101          |
| 5471.00       | 900.000        | 11.1592         | 3.4956          |
| 5371.00       | 1000.00        | 11.1400         | 3.4795          |
| 5271.00       | 1100.00        | 11.1188         | 3.4616          |
| 5171.00       | 1200.00        | 11.0956         | 3.4421          |
| 5153.90       | 1217.10        | 11.0914         | 3.4385          |
| 5153.90       | 1217.10        | 10.2578         | 0.              |
| 5071.00       | 1300.00        | 10.2364         | 0.              |
| 4971.00       | 1400.00        | 10.2044         | 0.              |
| 4871.00       | 1500.00        | 10.1657         | 0.              |
| 4771.00       | 1600.00        | 10.1203         | 0.              |
| 4671.00       | 1700.00        | 10.0681         | 0.              |
| 4571.00       | 1800.00        | 10.0092         | 0.              |
| 4471.00       | 1900.00        | 9.9435          | 0.              |
| 4371.00       | 2000.00        | 9.8711          | 0.              |
| 4271.00       | 2100.00        | 9.7920          | 0.              |
| 4171.00       | 2200.00        | 9.7062          | 0.              |
| 4071.00       | 2300.00        | 9.6136          | 0.              |
| 3971.00       | 2400.00        | 9.5142          | 0.              |
| 3871.00       | 2500.00        | 9.4082          | 0.              |
| 3771.00       | 2600.00        | 9.2954          | 0.              |
| 3671.00       | 2700.00        | 9.1758          | 0.              |
| 3571.00       | 2800.00        | 9.0496          | 0.              |
| 3471.00       | 2900.00        | 8.9166          | 0.              |
| 3371.00       | 3000.00        | 8.7768          | 0.              |
| 3271.00       | 3100.00        | 8.6303          | 0.              |
| 3171.00       | 3200.00        | 8.4771          | 0.              |
| 3071.00       | 3300.00        | 8.3171          | 0.              |
| 2971.00       | 3400.00        | 8.1504          | 0.              |
| 2889.00       | 3482.00        | 8.0087          | 0.              |
| 2889.00       | 3482.00        | 13.6908         | 7.3015          |
| 2871.00       | 3500.00        | 13.6866         | 7.2970          |
| 2771.00       | 3600.00        | 13.6636         | 7.2722          |
| 2740.00       | 3631.00        | 13.6564         | 7.2645          |
| 2740.00       | 3631.00        | 13.6564         | 7.2645          |
| 2671.00       | 3700.00        | 13.5725         | 7.2302          |
| 2571.00       | 3800.00        | 13.4531         | 7.1819          |
| 2471.00       | 3900.00        | 13.3359         | 7.1348          |

Table 2.2 (continued)

| Depth<br>(km) | Radius<br>(km) | $v_p$<br>(km/s) | $v_s$<br>(km/s) |
|---------------|----------------|-----------------|-----------------|
| 2371.00       | 4000.00        | 13.2203         | 7.0888          |
| 2271.00       | 4100.00        | 13.1055         | 7.0434          |
| 2171.00       | 4200.00        | 12.9911         | 6.9983          |
| 2071.00       | 4300.00        | 12.8764         | 6.9532          |
| 1971.00       | 4400.00        | 12.7607         | 6.9078          |
| 1871.00       | 4500.00        | 12.6435         | 6.8617          |
| 1771.00       | 4600.00        | 12.5241         | 6.8147          |
| 1671.00       | 4700.00        | 12.4020         | 6.7663          |
| 1571.00       | 4800.00        | 12.2764         | 6.7163          |
| 1471.00       | 4900.00        | 12.1469         | 6.6643          |
| 1371.00       | 5000.00        | 12.0127         | 6.6101          |
| 1271.00       | 5100.00        | 11.8732         | 6.5532          |
| 1171.00       | 5200.00        | 11.7279         | 6.4933          |
| 1071.00       | 5300.00        | 11.5761         | 6.4302          |
| 971.00        | 5400.00        | 11.4172         | 6.3635          |
| 871.00        | 5500.00        | 11.2506         | 6.2929          |
| 771.00        | 5600.00        | 11.0756         | 6.2180          |
| 760.00        | 5611.00        | 11.0558         | 6.2095          |
| 760.00        | 5611.00        | 11.0558         | 6.2095          |
| 671.00        | 5700.00        | 10.8192         | 5.9785          |
| 660.00        | 5711.00        | 10.7900         | 5.9500          |
| 660.00        | 5711.00        | 10.2000         | 5.6000          |
| 571.00        | 5800.00        | 9.9010          | 5.4113          |
| 471.00        | 5900.00        | 9.5650          | 5.1993          |
| 410.00        | 5961.00        | 9.3600          | 5.0700          |
| 410.00        | 5961.00        | 9.0300          | 4.8700          |
| 371.00        | 6000.00        | 8.8877          | 4.8021          |
| 271.00        | 6100.00        | 8.5227          | 4.6281          |
| 210.00        | 6161.00        | 8.3000          | 4.5220          |
| 210.00        | 6161.00        | 8.3000          | 4.5180          |
| 171.00        | 6200.00        | 8.1917          | 4.5102          |
| 120.00        | 6251.00        | 8.0500          | 4.5000          |
| 120.00        | 6251.00        | 8.0500          | 4.5000          |
| 71.00         | 6300.00        | 8.0442          | 4.4827          |
| 35.00         | 6336.00        | 8.0400          | 4.4700          |
| 35.00         | 6336.00        | 6.5000          | 3.7500          |
| 20.00         | 6351.00        | 6.5000          | 3.7500          |
| 20.00         | 6351.00        | 5.8000          | 3.3600          |
| 0.            | 6371.00        | 5.8000          | 3.3600          |

### 3 Model AK135

The AK135 velocity model has been augmented with a density and Q model by combining the study of travel times with those of free oscillations. This velocity and density model is the product of two pieces of work.

(1) The velocity model below 120 km depth comes from the work of Kennett et al. (1995). The original continental structure for the uppermost layering is given in Tab. 3.1 below. This model probably gives a reasonable representation of spherically averaged structure below 760 km depth. The upper mantle, as in IASP91, is an artificial construct which gives a good fit to the ensemble of observed travel times out to 30 degrees. The structure in D" should also be regarded as representative.

The representation of the velocity model is via point-wise values in velocity and linear interpolation in radius is used as the basis of the travel-time calculations. Note that AK135, unlike IASP91, is not a parameterized model. Any suitable interpolation scheme may be used where appropriate. A software conversion is available for reading velocity models into the IASP travel-time software and ellipticity corrections have been constructed for all the phases represented by that software. The software can be obtained from <http://rses.anu.edu.au/seismology/ttsoft.html>.

(2) Modified density and Q models come from a study by Montagner and Kennett (1996). This study introduces a density model and Q to the velocity distribution from the travel-time work to try to fit observations of free oscillation frequencies. An averaged uppermost structure is imposed on the AK135 velocities. The version of the model represented here is isotropic, even though the paper investigates the inclusion of anisotropy as well.

The complex density structure in the upper mantle with a density inversion reflects the absence of a low velocity zone in the wave-speed model. For a spherical average either a low shear-wave zone or a low density zone is needed to match the free oscillation frequencies. The Q values are those needed to bring the 1 Hz travel-time velocities into a match with the free oscillations and also give a good fit to observed Q values for the normal modes.

NB: The upper mantle density model should be treated with caution and may well change with further work.

**Table 3 AK135 velocity model for travel times**

#### 3.1 Continental structure

| Depth<br>(km) | $v_p$<br>(km/s) | $v_s$<br>(km/s) |
|---------------|-----------------|-----------------|
| 0.000         | 5.8000          | 3.4600          |
| 20.000        | 5.8000          | 3.4600          |
| 20.000        | 6.5000          | 3.8500          |
| 35.000        | 6.5000          | 3.8500          |
| 35.000        | 8.0400          | 4.4800          |
| 77.500        | 8.0450          | 4.4900          |
| 120.000       | 8.0500          | 4.5000          |

### 3.2 Average structure

| Depth<br>(km) | Density<br>(g/cm <sup>3</sup> ) | v <sub>p</sub><br>(km/s) | v <sub>s</sub><br>(km/s) | Q <sub>α</sub> | Q <sub>μ</sub> |
|---------------|---------------------------------|--------------------------|--------------------------|----------------|----------------|
| 0.00          | 1.0200                          | 1.4500                   | 0.0000                   | 57822.00       | 0.00           |
| 3.00          | 1.0200                          | 1.4500                   | 0.0000                   | 57822.00       | 0.00           |
| 3.00          | 2.0000                          | 1.6500                   | 1.0000                   | 163.35         | 80.00          |
| 3.30          | 2.0000                          | 1.6500                   | 1.0000                   | 163.35         | 80.00          |
| 3.30          | 2.6000                          | 5.8000                   | 3.2000                   | 1478.30        | 599.99         |
| 10.00         | 2.6000                          | 5.8000                   | 3.2000                   | 1478.30        | 599.99         |
| 10.00         | 2.9200                          | 6.8000                   | 3.9000                   | 1368.02        | 599.99         |
| 18.00         | 2.9200                          | 6.8000                   | 3.9000                   | 1368.02        | 599.99         |
| 18.00         | 3.6410                          | 8.0355                   | 4.4839                   | 950.50         | 394.62         |
| 43.00         | 3.5801                          | 8.0379                   | 4.4856                   | 972.77         | 403.93         |
| 80.00         | 3.5020                          | 8.0400                   | 4.4800                   | 1008.71        | 417.59         |
| 80.00         | 3.5020                          | 8.0450                   | 4.4900                   | 182.03         | 75.60          |
| 120.00        | 3.4268                          | 8.0505                   | 4.5000                   | 182.57         | 76.06          |
| 165.00        | 3.3711                          | 8.1750                   | 4.5090                   | 188.72         | 76.55          |
| 210.00        | 3.3243                          | 8.3007                   | 4.5184                   | 200.97         | 79.40          |
| 210.00        | 3.3243                          | 8.3007                   | 4.5184                   | 338.47         | 133.72         |
| 260.00        | 3.3663                          | 8.4822                   | 4.6094                   | 346.37         | 136.38         |
| 310.00        | 3.4110                          | 8.6650                   | 4.6964                   | 355.85         | 139.38         |
| 360.00        | 3.4577                          | 8.8476                   | 4.7832                   | 366.34         | 142.76         |
| 410.00        | 3.5068                          | 9.0302                   | 4.8702                   | 377.93         | 146.57         |
| 410.00        | 3.9317                          | 9.3601                   | 5.0806                   | 413.66         | 162.50         |
| 460.00        | 3.9273                          | 9.5280                   | 5.1864                   | 417.32         | 164.87         |
| 510.00        | 3.9233                          | 9.6962                   | 5.2922                   | 419.94         | 166.80         |
| 560.00        | 3.9218                          | 9.8640                   | 5.3989                   | 422.55         | 168.78         |
| 610.00        | 3.9206                          | 10.0320                  | 5.5047                   | 425.51         | 170.82         |
| 660.00        | 3.9201                          | 10.2000                  | 5.6104                   | 428.69         | 172.93         |
| 660.00        | 4.2387                          | 10.7909                  | 5.9607                   | 1350.54        | 549.45         |
| 710.00        | 4.2986                          | 10.9222                  | 6.0898                   | 1311.17        | 543.48         |
| 760.00        | 4.3565                          | 11.0553                  | 6.2100                   | 1277.93        | 537.63         |
| 809.50        | 4.4118                          | 11.1355                  | 6.2424                   | 1269.44        | 531.91         |
| 859.00        | 4.4650                          | 11.2228                  | 6.2799                   | 1260.68        | 526.32         |
| 908.50        | 4.5162                          | 11.3068                  | 6.3164                   | 1251.69        | 520.83         |
| 958.00        | 4.5654                          | 11.3897                  | 6.3519                   | 1243.02        | 515.46         |
| 1007.50       | 4.5926                          | 11.4704                  | 6.3860                   | 1234.54        | 510.20         |
| 1057.00       | 4.6198                          | 11.5493                  | 6.4182                   | 1226.52        | 505.05         |
| 1106.50       | 4.6467                          | 11.6265                  | 6.4514                   | 1217.91        | 500.00         |
| 1156.00       | 4.6735                          | 11.7020                  | 6.4822                   | 1210.02        | 495.05         |
| 1205.50       | 4.7001                          | 11.7768                  | 6.5131                   | 1202.04        | 490.20         |
| 1255.00       | 4.7266                          | 11.8491                  | 6.5431                   | 1193.99        | 485.44         |
| 1304.50       | 4.7528                          | 11.9208                  | 6.5728                   | 1186.06        | 480.77         |
| 1354.00       | 4.7790                          | 11.9891                  | 6.6009                   | 1178.19        | 476.19         |
| 1403.50       | 4.8050                          | 12.0571                  | 6.6285                   | 1170.53        | 471.70         |
| 1453.00       | 4.8307                          | 12.1247                  | 6.6554                   | 1163.16        | 467.29         |
| 1502.50       | 4.8562                          | 12.1912                  | 6.6813                   | 1156.04        | 462.96         |
| 1552.00       | 4.8817                          | 12.2558                  | 6.7070                   | 1148.76        | 458.72         |
| 1601.50       | 4.9069                          | 12.3181                  | 6.7323                   | 1141.32        | 454.55         |
| 1651.00       | 4.9321                          | 12.3813                  | 6.7579                   | 1134.01        | 450.45         |

**Table 3.2 (continued)**

| Depth<br>(km) | Density<br>(g/cm <sup>3</sup> ) | v <sub>p</sub><br>(km/s) | v <sub>s</sub><br>(km/s) | Q <sub>α</sub> | Q <sub>μ</sub> |
|---------------|---------------------------------|--------------------------|--------------------------|----------------|----------------|
| 1700.50       | 4.9570                          | 12.4427                  | 6.7820                   | 1127.02        | 446.43         |
| 1750.00       | 4.9817                          | 12.5030                  | 6.8056                   | 1120.09        | 442.48         |
| 1799.50       | 5.0062                          | 12.5638                  | 6.8289                   | 1108.58        | 436.68         |
| 1849.00       | 5.0306                          | 12.6226                  | 6.8517                   | 1097.16        | 431.03         |
| 1898.50       | 5.0548                          | 12.6807                  | 6.8743                   | 1085.97        | 425.53         |
| 1948.00       | 5.0789                          | 12.7384                  | 6.8972                   | 1070.38        | 418.41         |
| 1997.50       | 5.1027                          | 12.7956                  | 6.9194                   | 1064.23        | 414.94         |
| 2047.00       | 5.1264                          | 12.8524                  | 6.9416                   | 1058.03        | 411.52         |
| 2096.50       | 5.1499                          | 12.9093                  | 6.9625                   | 1048.09        | 406.50         |
| 2146.00       | 5.1732                          | 12.9663                  | 6.9852                   | 1042.07        | 403.23         |
| 2195.50       | 5.1963                          | 13.0226                  | 7.0069                   | 1032.14        | 398.41         |
| 2245.00       | 5.2192                          | 13.0786                  | 7.0286                   | 1018.38        | 392.16         |
| 2294.50       | 5.2420                          | 13.1337                  | 7.0504                   | 1008.79        | 387.60         |
| 2344.00       | 5.2646                          | 13.1895                  | 7.0722                   | 999.44         | 383.14         |
| 2393.50       | 5.2870                          | 13.2465                  | 7.0932                   | 990.77         | 378.79         |
| 2443.00       | 5.3092                          | 13.3017                  | 7.1144                   | 985.63         | 375.94         |
| 2492.50       | 5.3313                          | 13.3584                  | 7.1368                   | 976.81         | 371.75         |
| 2542.00       | 5.3531                          | 13.4156                  | 7.1584                   | 968.46         | 367.65         |
| 2591.50       | 5.3748                          | 13.4741                  | 7.1804                   | 960.36         | 363.64         |
| 2640.00       | 5.3962                          | 13.5311                  | 7.2031                   | 952.00         | 359.71         |
| 2690.00       | 5.4176                          | 13.5899                  | 7.2253                   | 940.88         | 354.61         |
| 2740.00       | 5.4387                          | 13.6498                  | 7.2485                   | 933.21         | 350.88         |
| 2740.00       | 5.6934                          | 13.6498                  | 7.2485                   | 722.73         | 271.74         |
| 2789.67       | 5.7196                          | 13.6533                  | 7.2593                   | 726.87         | 273.97         |
| 2839.33       | 5.7458                          | 13.6570                  | 7.2700                   | 725.11         | 273.97         |
| 2891.50       | 5.7721                          | 13.6601                  | 7.2817                   | 723.12         | 273.97         |
| 2891.50       | 9.9145                          | 8.0000                   | 0.0000                   | 57822.00       | 0.00           |
| 2939.33       | 9.9942                          | 8.0382                   | 0.0000                   | 57822.00       | 0.00           |
| 2989.66       | 10.0722                         | 8.1283                   | 0.0000                   | 57822.00       | 0.00           |
| 3039.99       | 10.1485                         | 8.2213                   | 0.0000                   | 57822.00       | 0.00           |
| 3090.32       | 10.2233                         | 8.3122                   | 0.0000                   | 57822.00       | 0.00           |
| 3140.66       | 10.2964                         | 8.4001                   | 0.0000                   | 57822.00       | 0.00           |
| 3190.99       | 10.3679                         | 8.4861                   | 0.0000                   | 57822.00       | 0.00           |
| 3241.32       | 10.4378                         | 8.5692                   | 0.0000                   | 57822.00       | 0.00           |
| 3291.65       | 10.5062                         | 8.6496                   | 0.0000                   | 57822.00       | 0.00           |
| 3341.98       | 10.5731                         | 8.7283                   | 0.0000                   | 57822.00       | 0.00           |
| 3392.31       | 10.6385                         | 8.8036                   | 0.0000                   | 57822.00       | 0.00           |
| 3442.64       | 10.7023                         | 8.8761                   | 0.0000                   | 57822.00       | 0.00           |
| 3492.97       | 10.7647                         | 8.9461                   | 0.0000                   | 57822.00       | 0.00           |
| 3543.30       | 10.8257                         | 9.0138                   | 0.0000                   | 57822.00       | 0.00           |
| 3593.64       | 10.8852                         | 9.0792                   | 0.0000                   | 57822.00       | 0.00           |
| 3643.97       | 10.9434                         | 9.1426                   | 0.0000                   | 57822.00       | 0.00           |
| 3694.30       | 11.0001                         | 9.2042                   | 0.0000                   | 57822.00       | 0.00           |
| 3744.63       | 11.0555                         | 9.2634                   | 0.0000                   | 57822.00       | 0.00           |
| 3794.96       | 11.1095                         | 9.3205                   | 0.0000                   | 57822.00       | 0.00           |
| 3845.29       | 11.1623                         | 9.3760                   | 0.0000                   | 57822.00       | 0.00           |
| 3895.62       | 11.2137                         | 9.4297                   | 0.0000                   | 57822.00       | 0.00           |

**Table 3.2 (continued)**

| Depth<br>(km) | Density<br>(g/cm <sup>3</sup> ) | v <sub>p</sub><br>(km/s) | v <sub>s</sub><br>(km/s) | Q <sub>α</sub> | Q <sub>μ</sub> |
|---------------|---------------------------------|--------------------------|--------------------------|----------------|----------------|
| 3945.95       | 11.2639                         | 9.4814                   | 0.0000                   | 57822.00       | 0.00           |
| 3996.28       | 11.3127                         | 9.5306                   | 0.0000                   | 57822.00       | 0.00           |
| 4046.62       | 11.3604                         | 9.5777                   | 0.0000                   | 57822.00       | 0.00           |
| 4096.95       | 11.4069                         | 9.6232                   | 0.0000                   | 57822.00       | 0.00           |
| 4147.28       | 11.4521                         | 9.6673                   | 0.0000                   | 57822.00       | 0.00           |
| 4197.61       | 11.4962                         | 9.7100                   | 0.0000                   | 57822.00       | 0.00           |
| 4247.94       | 11.5391                         | 9.7513                   | 0.0000                   | 57822.00       | 0.00           |
| 4298.27       | 11.5809                         | 9.7914                   | 0.0000                   | 57822.00       | 0.00           |
| 4348.60       | 11.6216                         | 9.8304                   | 0.0000                   | 57822.00       | 0.00           |
| 4398.93       | 11.6612                         | 9.8682                   | 0.0000                   | 57822.00       | 0.00           |
| 4449.26       | 11.6998                         | 9.9051                   | 0.0000                   | 57822.00       | 0.00           |
| 4499.60       | 11.7373                         | 9.9410                   | 0.0000                   | 57822.00       | 0.00           |
| 4549.93       | 11.7737                         | 9.9761                   | 0.0000                   | 57822.00       | 0.00           |
| 4600.26       | 11.8092                         | 10.0103                  | 0.0000                   | 57822.00       | 0.00           |
| 4650.59       | 11.8437                         | 10.0439                  | 0.0000                   | 57822.00       | 0.00           |
| 4700.92       | 11.8772                         | 10.0768                  | 0.0000                   | 57822.00       | 0.00           |
| 4751.25       | 11.9098                         | 10.1095                  | 0.0000                   | 57822.00       | 0.00           |
| 4801.58       | 11.9414                         | 10.1415                  | 0.0000                   | 57822.00       | 0.00           |
| 4851.91       | 11.9722                         | 10.1739                  | 0.0000                   | 57822.00       | 0.00           |
| 4902.24       | 12.0001                         | 10.2049                  | 0.0000                   | 57822.00       | 0.00           |
| 4952.58       | 12.0311                         | 10.2329                  | 0.0000                   | 57822.00       | 0.00           |
| 5002.91       | 12.0593                         | 10.2565                  | 0.0000                   | 57822.00       | 0.00           |
| 5053.24       | 12.0867                         | 10.2745                  | 0.0000                   | 57822.00       | 0.00           |
| 5103.57       | 12.1133                         | 10.2854                  | 0.0000                   | 57822.00       | 0.00           |
| 5153.50       | 12.1391                         | 10.2890                  | 0.0000                   | 57822.00       | 0.00           |
| 5153.50       | 12.7037                         | 11.0427                  | 3.5043                   | 633.26         | 85.03          |
| 5204.61       | 12.7289                         | 11.0585                  | 3.5187                   | 629.89         | 85.03          |
| 5255.32       | 12.7530                         | 11.0718                  | 3.5314                   | 626.87         | 85.03          |
| 5306.04       | 12.7760                         | 11.0850                  | 3.5435                   | 624.08         | 85.03          |
| 5356.75       | 12.7980                         | 11.0983                  | 3.5551                   | 621.50         | 85.03          |
| 5407.46       | 12.8188                         | 11.1166                  | 3.5661                   | 619.71         | 85.03          |
| 5458.17       | 12.8387                         | 11.1316                  | 3.5765                   | 617.78         | 85.03          |
| 5508.89       | 12.8574                         | 11.1457                  | 3.5864                   | 615.93         | 85.03          |
| 5559.60       | 12.8751                         | 11.1590                  | 3.5957                   | 614.21         | 85.03          |
| 5610.31       | 12.8917                         | 11.1715                  | 3.6044                   | 612.62         | 85.03          |
| 5661.02       | 12.9072                         | 11.1832                  | 3.6126                   | 611.12         | 85.03          |
| 5711.74       | 12.9217                         | 11.1941                  | 3.6202                   | 609.74         | 85.03          |
| 5762.45       | 12.9351                         | 11.2041                  | 3.6272                   | 608.48         | 85.03          |
| 5813.16       | 12.9474                         | 11.2134                  | 3.6337                   | 607.31         | 85.03          |
| 5863.87       | 12.9586                         | 11.2219                  | 3.6396                   | 606.26         | 85.03          |
| 5914.59       | 12.9688                         | 11.2295                  | 3.6450                   | 605.28         | 85.03          |
| 5965.30       | 12.9779                         | 11.2364                  | 3.6498                   | 604.44         | 85.03          |
| 6016.01       | 12.9859                         | 11.2424                  | 3.6540                   | 603.69         | 85.03          |
| 6066.72       | 12.9929                         | 11.2477                  | 3.6577                   | 603.04         | 85.03          |
| 6117.44       | 12.9988                         | 11.2521                  | 3.6608                   | 602.49         | 85.03          |
| 6168.15       | 13.0036                         | 11.2557                  | 3.6633                   | 602.05         | 85.03          |
| 6218.86       | 13.0074                         | 11.2586                  | 3.6653                   | 601.70         | 85.03          |

**Table 3.2 (continued)**

| Depth<br>(km) | Density<br>(g/cm <sup>3</sup> ) | v <sub>p</sub><br>(km/s) | v <sub>s</sub><br>(km/s) | Q <sub>α</sub> | Q <sub>μ</sub> |
|---------------|---------------------------------|--------------------------|--------------------------|----------------|----------------|
| 6269.57       | 13.0100                         | 11.2606                  | 3.6667                   | 601.46         | 85.03          |
| 6320.29       | 13.0117                         | 11.2618                  | 3.6675                   | 601.32         | 85.03          |
| 6371.00       | 13.0122                         | 11.2622                  | 3.6678                   | 601.27         | 85.03          |

**Note:** The bulk Q<sub>κ</sub> given in Table 1 differs from the Q<sub>α</sub> for P waves given in Table 3.2. The following relationship holds:

$$Q_{\alpha}^{-1} = L Q_{\mu}^{-1} + (1 - L)Q_{\kappa}^{-1}$$

where  $L = (4/3)(\beta/\alpha)^2$ ,  $\alpha$  = P-wave velocity v<sub>p</sub> and  $\beta$  = S-wave velocity v<sub>s</sub>. However, Q<sub>μ</sub> = Q<sub>β</sub> for S waves and Q<sub>α</sub> = Q<sub>κ</sub> in the liquid outer core because of  $\mu$  and Q<sub>μ</sub> = 0.

## References

- Dziewonski, A. M., and Anderson, D. L. (1981). Preliminary reference Earth model. *Phys. Earth Planet. Inter.*, **25**, 297-356.
- Kennett, B. L. N. (Ed.) (1991). IASPEI 1991 Seismological Tables. Research School of Earth Sciences, Australian National University, 167 pp.
- Kennett, B. L. N., and Engdahl, E. R. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, **105**, 429-465.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, **122**, 108-124.
- Montagner, J.-P., and Kennett, B. L. N. (1996). How to reconcile body-wave and normal-mode reference Earth models?. *Geophys. J. Int.*, **125**, 229-248.
- Shearer, P. M. (1999). Introduction to seismology. Cambridge University Press, 260 pp.

| Topic   | The IASPEI standard nomenclature of seismic phases                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|---------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p>Peter Bormann (formerly GeoForschungsZentrum Potsdam, Telegrafenberg, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> <p>Dmitry A. Storchak, International Seismological Centre, Pipers Lane, Thatcham, Berkshire RG19 4NS, UK, England, E-mail: <a href="mailto:dmitry@isc.ac.uk">dmitry@isc.ac.uk</a></p> <p>Johannes Schweitzer, NORSAR, P.O. Box 53, N-2027, Kjeller, Norway, E-mail: <a href="mailto:johannes.schweitzer@norsar.no">johannes.schweitzer@norsar.no</a></p> |
| Version | September 2013; DOI: 10.2312/GFZ.NMSOP-2_IS_2.1                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |

|                                                                                | page         |
|--------------------------------------------------------------------------------|--------------|
| <b>1 Introduction</b>                                                          | <b>1</b>     |
| <b>2 Standard letters, signs and syntax used for describing seismic phases</b> | <b>3</b>     |
| 2.1 Capital letters                                                            | 3            |
| 2.2 Lower case letters and signs                                               | 4            |
| 2.3 Syntax of generating complex phase names                                   | 5            |
| <b>3 Examples for creating complex standard phase names</b>                    | <b>6</b>     |
| 3.1 Refracted and converted refracted waves                                    | 6            |
| 3.2 Pure reflected waves                                                       | 6            |
| 3.3 Reflected waves with conversion at the reflection point                    | 7            |
| <b>4 Ray-paths diagrams for some of the IASPEI standard phases</b>             | <b>7</b>     |
| 4.1 Seismic rays of crustal phases                                             | 8            |
| 4.2 Seismic rays of mantle phases                                              | 9            |
| 4.3 Seismic rays of phases travelling through the Earth core                   | 10           |
| <b>5 IASPEI standard seismic phase list</b>                                    | <b>12</b>    |
| 5.1 Crustal phases                                                             | 12           |
| 5.2 Mantle phases                                                              | 13           |
| 5.3 Core phases                                                                | 15           |
| 5.4 Near surface reflections (depth phases)                                    | 16           |
| 5.5 Surface waves                                                              | 17           |
| 5.6 Acoustic waves                                                             | 17           |
| 5.7 Unidentified arrivals                                                      | 17           |
| 5.8 Amplitude measurement phases                                               | 18           |
| <b>7 References</b>                                                            | <b>19-20</b> |

## 1 Introduction

A Working Group (WG) on Standard Phase Names was set up by the IASPEI Commission on Seismological Observation and Interpretation in 2001. The working group was chaired by D. A. Storchak of the ISC. Members of the group were R. D. Adams, P. Bormann, R. E. Engdahl, J. Havskov, B. L. N. Kennett and J. Schweitzer. The working group has put together a modified standard nomenclature of seismic phases, which was meant to be concise, consistent and self-explanatory on the basis of agreed rules. The result is not a complete list of all phases. The list is not meant to satisfy specific requirements of seismologists to name various phases used in a particular type of research. Instead, the new phase list aims at inviting data analysts and other users to ensure an expanded standardized data reporting and



exchange. This will result in a broader and unambiguous database for research and practical applications. At the same time the attached list and its principles outlined below may be a useful guidance when proposing names to previously unknown seismic phases.

After numerous consultations with the seismological community, the Standard Seismic Phase List was finalized and adopted by the CoSOI/IASPEI at its meeting in Sapporo on July 04, 2003. The original draft list of standard seismic phase names was first published as part of the New Manual of Seismological Observatory Practice (Storchak et al., 2002). Later on, the list was published in the Seismological Research Letters (Storchak et al., 2003). Since then various updates to the list were required due to progress in observational seismology and relevant changes in other observational standards. These changes were accommodated in a contribution by Storchak et al. (2011) to the Encyclopedia of Solid Earth Geophysics. The current amended version of IS 2.1 takes all these updates into account. Moreover, it aims at unifying some wording in the definitions. The List should be used as a guide to identify such phases in real seismic records by taking into account additional resources provided in DS11.1 to DS11.4 of the NMSOP-2.

The new IASPEI nomenclature partially modifies and complements the earlier one published in the last edition of the Willmore (1979) Manual of Seismological Observatory Practice and in the Summaries of the Bulletin of the International Seismological Centre (2010). It is more in tune with phase definitions according to modern Earth and travel-time models (see Chapter 2, section 2.7) and the definition of pronounced travel-time branches, of core phases in particular (see manual sections 11.5.2.4 and 11.5.3). As opposed to former practice, the WG tried to make sure that a phase name generally reflects the type of the wave and the path it has traveled. Accordingly, symbols for characterizing onset quality, polarity etc. are no longer part of the phase name. Also, the WG acknowledges that there exist several types of seismic phases, particularly among crustal phases, that are common in some regions yet never or rarely found in other regions, such as, e.g., Pb (P\*), PnPn, PbPb, etc. The names and definitions of acoustic and hydroacoustic phases and amplitude measurement are likely to be reviewed based on the results of recent developments and new analysis practices being established.

The extended list of phase names, as presented below in section 5, accounts for significantly increased detection capabilities of modern seismic sensors and sensor arrays, especially for weak phases that were rarely found using the classical analog records. It also accounts for improved possibilities of proper phase identification by means of digital multi-channel data processing such as frequency-wavenumber (f-k) analysis and polarization filtering, by modeling the observations with synthetic seismograms or by showing on the records the theoretically predicted onset times of phases. Furthermore, limitation of classical formats for wave parameter reporting to international data centers, such as the Telegraphic Format (TF), which allowed only the use of capital letters and numbers, are no longer relevant in times of data exchange via the Internet. Finally, the newly adopted IASPEI Seismic Format (ISF; see <http://www.isc.ac.uk/standards/isf/download/isf.pdf>, Chapter 10, section 10.2.5 and IS 10.2) is much more flexible than the old formats accepted by the NEIC, ISC and other data centers. It also allows reporting, computer parsing and archiving of all phases, including previously uncommon ones. The ISF also accepts complementary parameters such as onset quality, signal-to-noise ratio, measured backazimuth and slowness, amplitudes and periods of other phases in addition to P and surface waves, for components other than vertical ones, and for non-standard response characteristics.

This increased flexibility of the parameter-reporting format requires improved standardization, which limits an uncontrolled growth of incompatible and ambiguous parameter data. Therefore, the WG agreed on certain rules. They are outlined below prior to the listing of standardized phase names. In order to ease the understanding of verbal definitions of the phase names and ray diagrams are presented. The ray diagrams have been calculated for local seismic sources on the basis of an average one-dimensional two-layer crustal model and for regional and teleseismic sources by using the global, radial symmetrical Earth model *ak135* (Kennett et al., 1995; see also Fig. 2.79).

Before elaborating short-cut seismic phase names one should agree first on the *language* to be used and its rules. As in any other language we need a suitable alphabet (here Latin letters), numbers (here Arabic numbers and + and - signs), an *orthography*, which regulates, e.g., the use of capital and lower case letters, and a *syntax*, i.e., rules of correct order and mutual relationship of the language elements. One should be aware, however, that the seismological nomenclature will inevitably develop exceptions to the rules, as any historically developed language, and also depending on the context in which it is used. Although not fully documented below, some exceptions will be mentioned. Note that our efforts are mainly aimed at standardized names to be used in international data exchange so as to build up unique, unambiguous global databases for research. Many of the exceptions to the rules are related to specialized, mostly local research applications. The identification of related seismic phases often requires specialized procedures of data acquisition and processing, which are not part of seismological routine data analysis. Also, many of these *exceptional* phases are rarely or never used in seismic event location, magnitude determination, source mechanism calculations etc., which are the main tasks of international data centers. Below, we focus therefore on phases, which are particularly important for seismological data centers as well as for the refinement of regional and global Earth models on the basis of widely exchanged and accumulated parameter readings from such phases. In addition, we added for some phase definitions references to which the particular phase names can be traced back.

## 2 Standard letters, signs and syntax used for describing seismic phases

### 2.1 Capital letters

Individual capital letters that stand for primary types of seismic *body waves* such as:

- P: longitudinal wave that has traveled through the Earth's crust and mantle, from *undae primae* (Latin) = first waves (Borne, 1904);
- K: longitudinal wave that has traveled through the Earth's outer core, K, from Kern (German) = core (Sohon, 1932; Bastings, 1934);
- I: longitudinal wave that has traveled through the Earth's inner core (Jeffreys and Bullen, 1940);
- S: transverse wave that has traveled through the Earth's crust and mantle, from *undae secundae* (Latin) = second waves (Borne, 1904);
- T: wave that has partly traveled as sound wave in the sea, from *undae tertiae* = third waves (Linehan, 1940);
- J: transverse wave that has traveled through the Earth's inner core (Bullen, 1946).

**Exceptions:**

- A **capital letter N** used in the nomenclature does not stand for a phase name but rather for the number of legs traveled (or N-1 reflections made) before reaching the station. N should usually follow the phase symbol to which it applies. For examples see syntax below.
- The **lower case letters p and s** may stand, in the case of seismic events below the Earth's surface, for the relatively short **upgoing leg** of P or S waves, which continue, after reflection and possible conversion at the free surface, as downgoing P or S wave. Thus seismic *depth phases* (e.g., pP, sP, sS, pPP, sPP, pPKP, etc.) are uniquely defined. The identification and reporting of such phases is of utmost importance for source depth determination (Scrase, 1931; Stechschulte, 1932; Gutenberg et al., 1933; Macelwane et al., 1933).
- Many researchers working on detailed investigations of crustal and upper-mantle discontinuities denote both the up- and downgoing short legs of converted or multiply reflected P and S phases as lower case letters p and s, respectively.

Individual or double capital letters that stand for *surface waves* such as:

- L: (relatively) long-period surface wave, unspecified, from *undae longae* (Latin) = long waves (Borne, 1904);
- R: Rayleigh waves (short- to very long-period waves in crust and upper mantle) (Angenheister, 1921);
- Q: Love waves, from Querwellen (German) = transverse waves (Angenheister, 1921);
- G: (very long-period) global (mantle) Love waves, firstly observed and reported by Gutenberg and Richter (1934); in honor of Gutenberg, Byerly proposed the usage of G for Gutenberg for these waves Richter (1958);
- LR: long-period Rayleigh waves, usually relating to the Airy-phase maximum in the surface wave train;
- LQ: long-period Love waves.
- PL: fundamental leaking mode following P onsets, firstly observed and reported by Somville (1930, 1931)

## 2.2 Lower case letters and signs

*Single lower case letters* generally specify in which part of Earth's crust or upper mantle in which a phase has its *turning point* or at which discontinuity it has been *reflected and/or eventually converted*:

- g (following a general phase name): characterizes waves "bottoming" (i.e., having their turning point in the case of P- or S-body waves) or just travel (surface waves) within the upper ("granitic") Earth's crust (e.g., Pg, Sg; Rg), (Jeffreys, 1926);
- b (following a general phase name): characterizes body waves "bottoming" in the lower ("basaltic") Earth's crust (Jeffreys, 1926) (e.g., Pb, Sb; alternative names for these phases are P\*, S\*, (Conrad, 1925)); also used for phases radiated directly to the Earth's surface from events in the lower crust;
- n (following a general phase name): characterizes a P or S wave that is "bottoming" in the Earth's uppermost mantle or is traveling as head wave below the

Mohorovičić (Moho) discontinuity (e.g., Pn, Sn), introduced after Andrija Mohorovičić discovered the Earth's crust and separated the crustal travel-time curve from the normal (=n) mantle phase (Mohorovičić, 1910); also used for phases radiated directly to the Earth's surface from events in the uppermost mantle;

- m: (upward) reflections from the *outer side* of the Moho (e.g., PmP, SmS);
- c: reflections from the *outer side* of the core-mantle boundary (CMB), usage proposed by James B. Macelwane (see Gutenberg, 1925);
- i: reflections from the *outer side* of the inner core boundary (ICB);
- z: reflections from a discontinuity (other than free surface, CMB or ICB) at depth z (measured in km). Upward reflections from the *outer side* of the discontinuity *may* additionally be complemented by a + sign (e.g., P410+P; this, however, is not compulsory) while downward reflections from the *inner side* of the discontinuity *must* be complemented by a – sign (e.g., P660–P).

An exception from these rules is the use of lowercase p or s to indicate arrivals of longitudinal or transverse waves that were from a source below the Earth's surface first radiated to go up toward the free surface to be reflected / converted back into the Earth as normal P or S waves (see near source surface reflections and conversions section of the phase list below).

To call phases radiated directly to the Earth's surface in the near source region from events in the lower crust (Conrad discontinuity) Pb and from below the Moho Pn is a quite pragmatic decision, which helps analysts and source location programs to separate the different branches of the travel-time curves.

*Double lower case letters* following a capital letter phase name indicate the travel-time branch to which this phase belongs. Due to the geometry and velocity structure of the Earth the same type of seismic wave may develop a triplication of its travel-time curve with different, in some cases well separated branches (see Chapter 2, Fig. 2.32). Thus it is customary to differentiate between different branches of core phases and their multiple reflections at the free surface or the CMB. Examples are PKPab, PKPbc, PKPdf, SKSac, SKKSac, etc.. The separation of the different PKP branches with letters ab, bc and df was introduced by Jeffreys and Bullen (1940).

*Three lower case letters* may follow a capital letter phase name in order to specify its character, e.g., as a forerunner (pre) to the main phase, caused by scattering (e.g., PKPpre) or as a diffracted wave extending the travel-time branch of the main phase into the outer core shadow (e.g., Pdif in the outer core shadow for P).

## 2.3 Syntax of generating complex phase names

Due to refraction, reflection and conversion in the Earth most of phases have a complex path history before they reach the station. Accordingly, most phases cannot be described by a single capital letter code in a self-explanatory way. By combining the capital and lower case letters as mentioned above, one can describe the character of even rather complex refracted, reflected or converted phases. The order of symbols (syntax) regulates the sequence of phase legs due to refraction, reflection and conversion events in time (from left to right) and in space.

### 3 Examples for creating complex standard phase names

#### 3.1 Refracted and converted refracted waves

- PKP is a pure refracted longitudinal wave. It has traveled the first part of its path as P through crust and mantle, the second through the outer core, and the third again as P through mantle and crust. An alternative name for PKP is P' (Angenheister, 1921), which should be read as “P prime”.
- PKIKP (alternative to PKP<sub>df</sub>) is also a pure refracted longitudinal wave too. It has traveled as P the first part of its path through crust and mantle, the second through the outer core, the third through the inner core, and the fourth and fifth parts back again through outer core and mantle/crust.
- SKS is a converted refracted wave. It has traveled as a shear wave through crust and mantle, being converted into a longitudinal P wave when refracted into the outer core and converted back again into an S wave when entering the mantle.
- SKP or PKS are converted refracted waves with only one conversion from S to P when entering the core or from P to S when leaving the core, respectively.

#### 3.2 Pure reflected waves

- In the case of (downward only) reflections at the free surface or from the inner side of the CMB, the phase symbol is just repeated, e.g., PP, SS (Geiger, 1909), PPP, SSS, KK, KKK etc.
- In the case of (upward) reflections from the outer side of the Moho, the CMB, or the ICB, this is indicated by inserting between the phase symbols m, c or i, respectively, between the phase symbols, e.g., PmP, PcP, ScS; PKiKP.
- Reflections from any other discontinuity in mantle or crust at depth z may be from the inner side (-; i.e., downward back into the mantle) or from the outer side (+; i.e., back towards the surface). To differentiate between these two possibilities, the sign has to follow z (or the respective number in km); for example, P410+P or P660-P.
- To abbreviate names of multi-leg phases due to repeated reflections, one can also write PhasenameN. This kind of abbreviation, is rather customary in the case of multiple phases with long phase names such as PmP2 for PmPPmP (free-surface reflection of PmP), SKS2 for SKSSKS (alternative name for S'2, the free-surface reflection of SKS), PKP3 for PKPPKPPKP (double free-surface reflection of PKP; alternative name to P'3) or P4KP for PKKKKP (triple reflection of P at the inner side of the CMB).

**Note 1:** PKP2 = PKPPKP are now alternative names for P'2 or P'P', respectively. This should not be mistaken for the *old* name PKP2 for PKPab.

**Note 2:** In the case of multiple reflections from the inner side of the CMB the WG followed the established tradition of placing the number N not after but *in front* of the related phase symbol K.

### 3.3 Reflected waves with conversion at the reflection point

In the case that a phase changes its character from P to S, or vice versa, one writes:

- PS (first leg P, second leg S) or SP (first leg P, second leg S) in the case of reflection/conversion from the free surface downward into the mantle (Geiger and Gutenberg, 1912a, 1912b);
- PmS or SmP, respectively, for reflection/conversion from the outer side of the Moho;
- PcS or ScP for reflection/conversion from the outer side of the CMB;
- Pz+S or Sz-P for reflection/conversion from the outer (+) side or inner (-) side, respectively, of a discontinuity at depth  $z$ . Note that the - is compulsory, the + not.

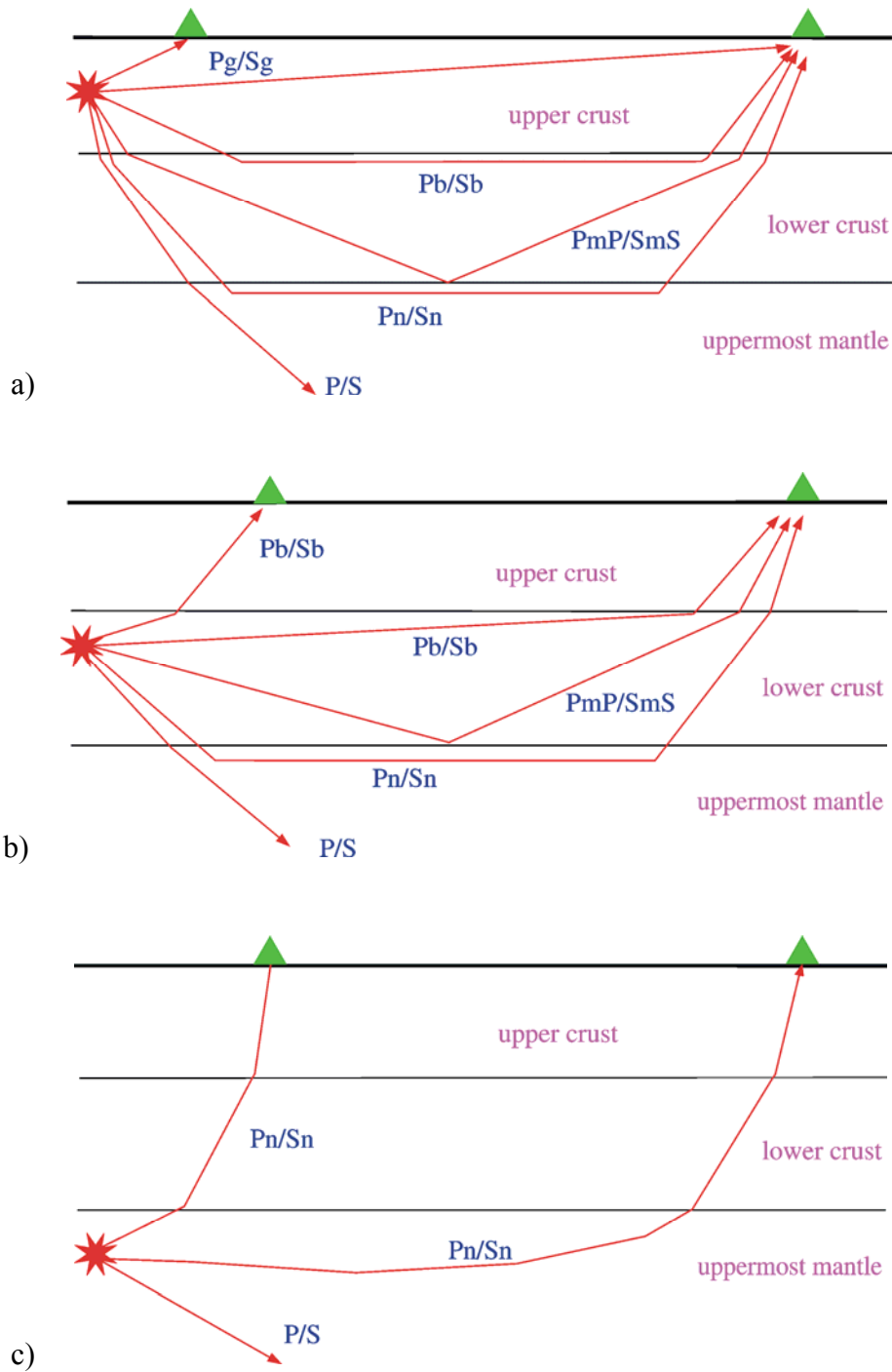
In this context it is worth mentioning that mode conversion is impossible for reflections from the inner side of the CMB back into the outer core because the liquid outer core does not allow the propagation of S waves.

The WG determined the new IASPEI standard phase names along these lines and rules. Where these deviate from other traditionally used names the latter are given as well. Either, the traditional names are still acceptable *alternative names* (alt) or they are now *old names* (old), which should no longer be used.

## 4 Ray-paths diagrams for some of the IASPEI standard phases

In this section we show ray paths through the Earth for most of the mentioned phases. The three figures for crustal phases are just sketches showing the principal ray paths in a two-layer crust. The rays in all other figures were calculated by using the ray picture part of the WKB3 code (Chapman, 1978; Dey-Sarkar and Chapman, 1978); as velocity model we chose the standard Earth model *ak135* (Kennett et al., 1995). For some types of P and S phases the ray paths through the Earth are very similar because the velocity ratio  $v_P/v_S$  does not change enough to give very different ray pictures. In these cases, we calculated only the ray paths for the P-type ray (i.e., P, Pdif, pP, PP, P3, PcP, PcP2, P660P and P660-P) and assume that the corresponding ray paths of the respective S-type phases are very similar. To show the different ray paths for phases with similar phase names, we show on many figures rays leaving the source once to the left and once to the right in different colors. The three most important discontinuities inside the Earth are indicated as black circles (i.e., the border between upper and lower mantle, the CMB, and the ICB).

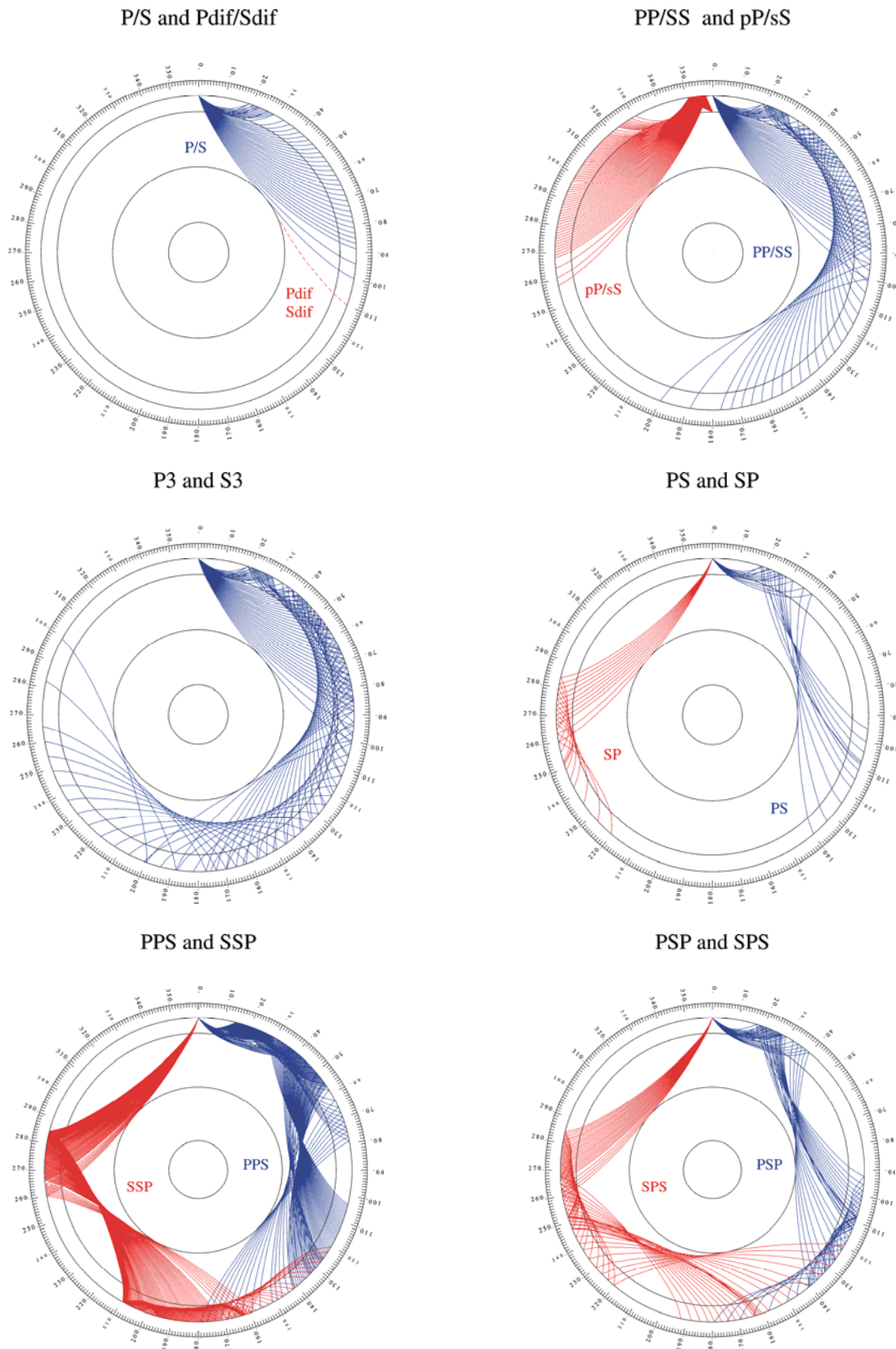
#### 4.1 Seismic rays of crustal phases



**Figure 1** Seismic rays of „crustal phases“ observed in the case of a two-layer crust in local and regional distance ranges ( $0^\circ < D < \text{about } 20^\circ$ ) from the seismic source in the: a) upper crust; b) lower crust; and c) uppermost mantle.

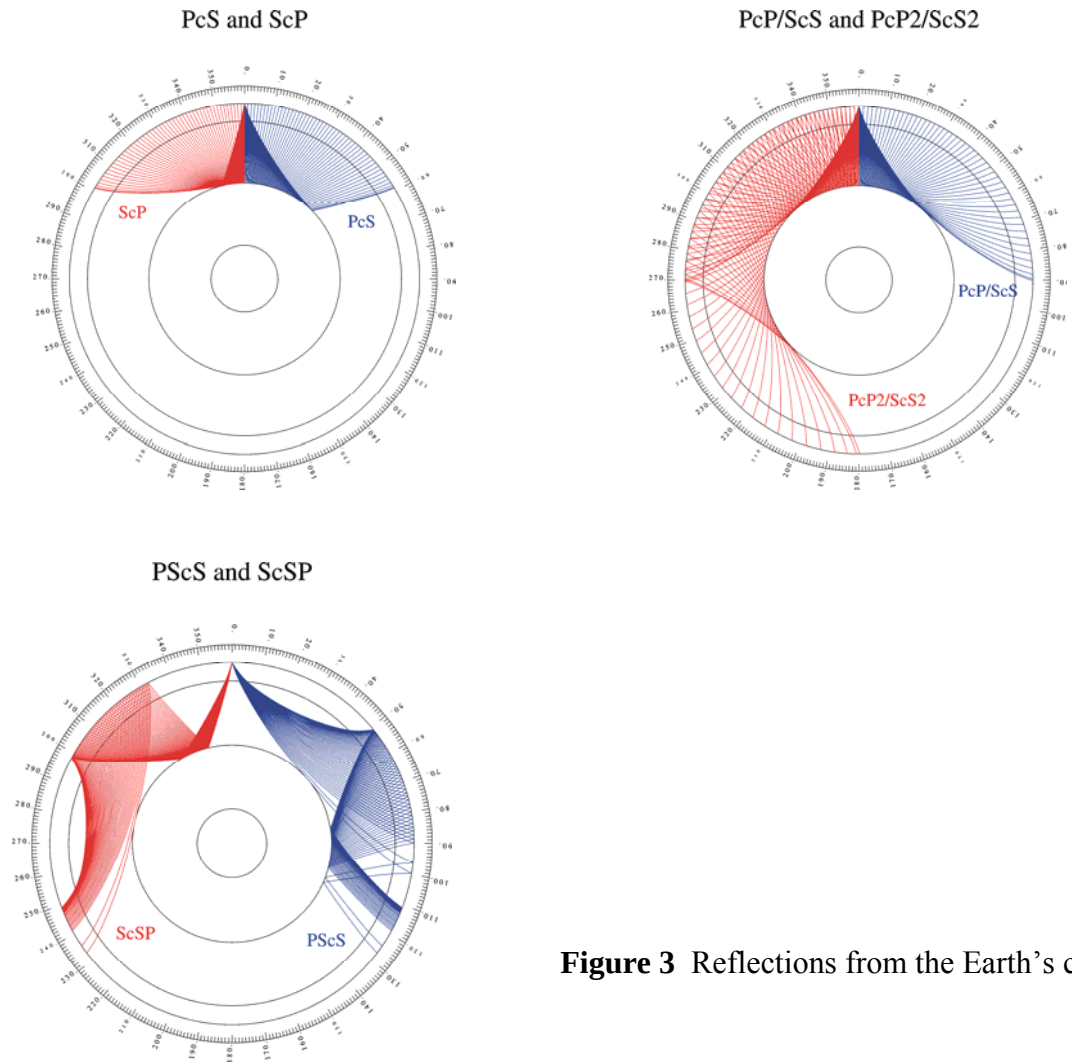
## 4.2 Seismic rays of mantle phases

**Note:** The density of seismic rays arriving at a given location is proportional to the relative amplitude of the respective seismic phase observed at this location.



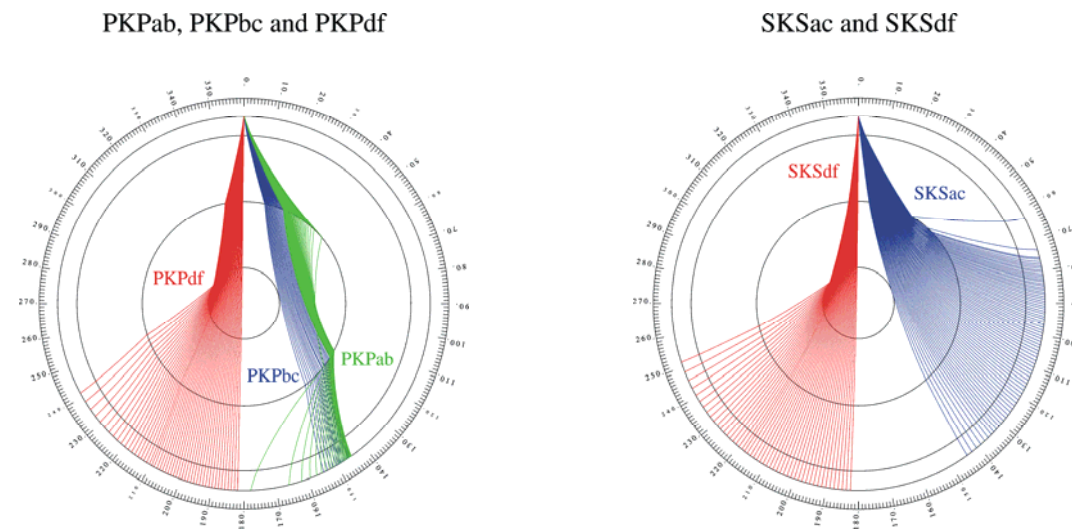
**Figure 2** Mantle phases observed at the teleseismic distance range  $D > \text{about } 20^\circ$ .



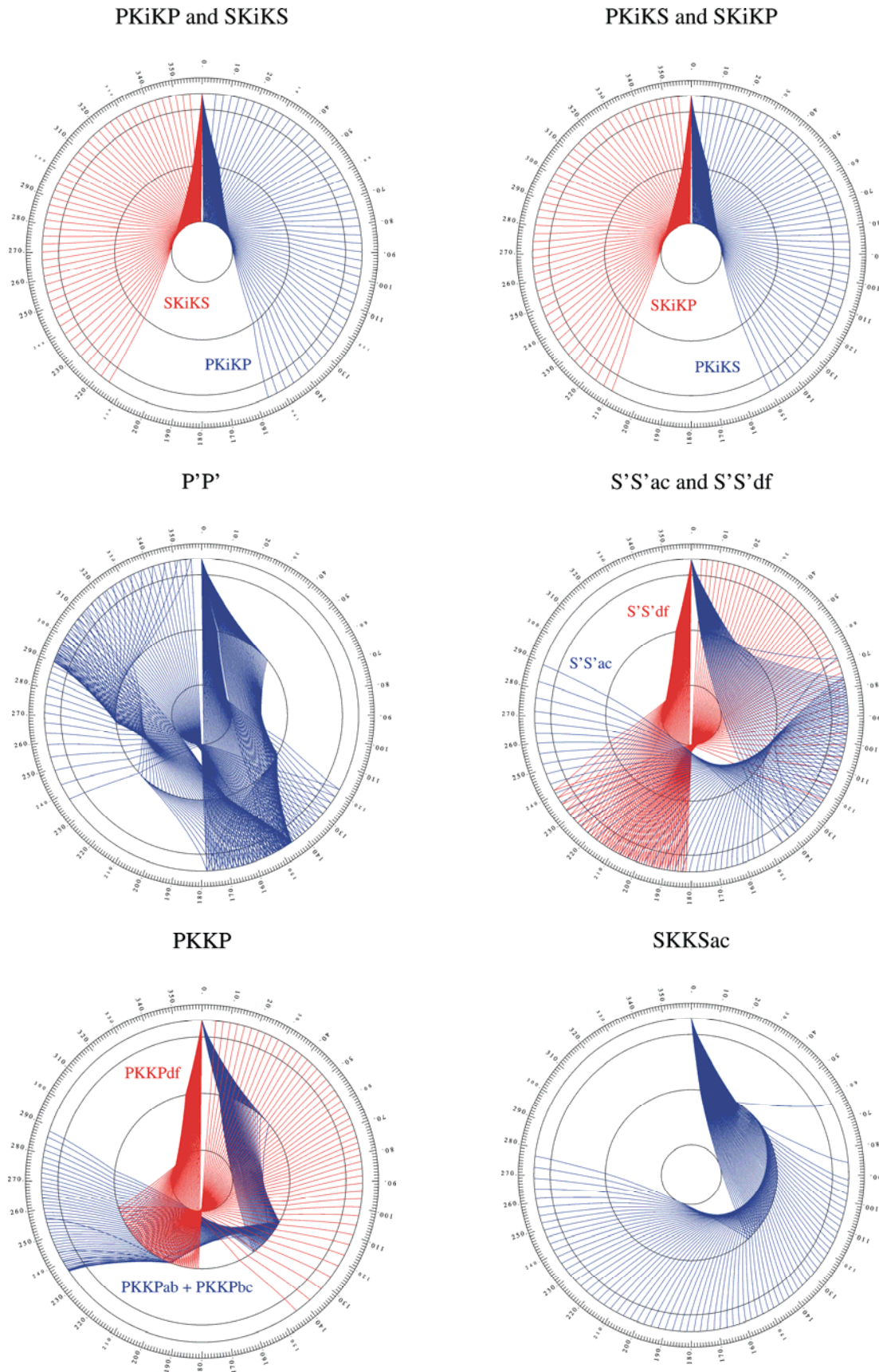


**Figure 3** Reflections from the Earth's core.

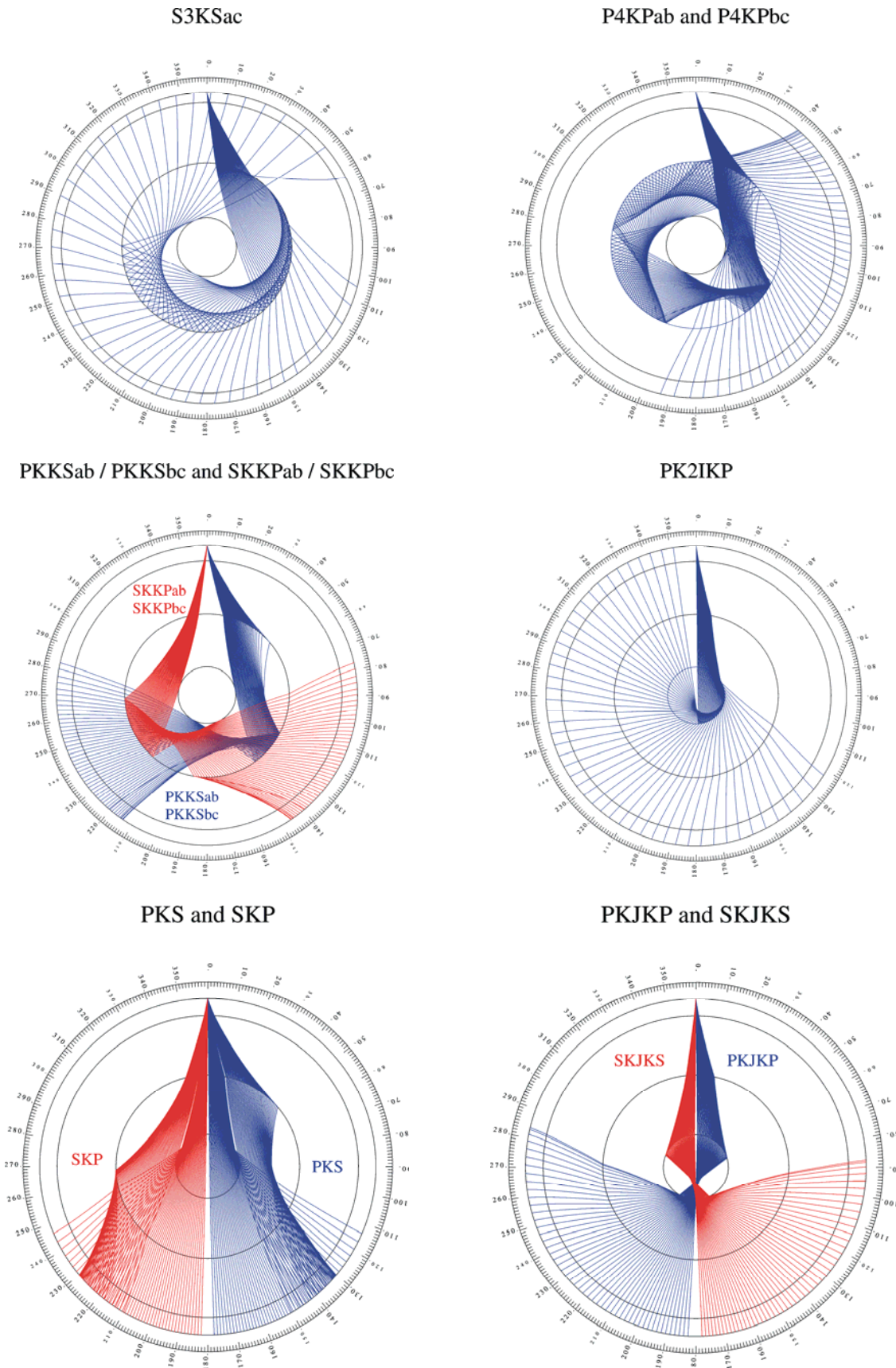
#### 4.3 Seismic rays of phases travelling through the Earth's core



**Figure 4** Seismic rays of direct core phases



**Figure 5** Seismic rays of single-reflected core phases



**Figure 6** Seismic rays of multiple-reflected and converted core phases.

## 5 IASPEI Standard Seismic Phase List

### 5.1 CRUSTAL PHASES (For ray traces see Figure 1; for more examples see Chapter 2, section 2.6.1; Chapter 11, section 11.5.1, DS 11.1 and EX 11.1)

|      |                                                                                                                                                                                                                                                                                     |
|------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Pg   | At short distances, either an upgoing P wave from a source in the upper crust or a P wave bottoming in the upper crust. At larger distances also arrivals caused by multiple P-wave reverberations inside the whole crust with a group velocity around 5.8 km/s                     |
| Pb   | (alt: P*) Either an upgoing P wave from a source in the lower crust or a P wave bottoming in the lower crust                                                                                                                                                                        |
| Pn   | Any P wave bottoming in the uppermost mantle or an upgoing P wave from a source in the uppermost mantle                                                                                                                                                                             |
| PnPn | Pn free-surface reflection                                                                                                                                                                                                                                                          |
| PgPg | Pg free-surface reflection                                                                                                                                                                                                                                                          |
| PmP  | P reflection from the outer side of the Moho                                                                                                                                                                                                                                        |
| PmPN | PmP multiple free-surface reflection; N is a positive integer. For example, PmP2 is PmPPmP                                                                                                                                                                                          |
| PmS  | P to S reflection/conversion from the outer side of the Moho                                                                                                                                                                                                                        |
| Sg   | At short distances, either an upgoing S wave from a source in the upper crust or an S wave bottoming in the upper crust. At larger distances also, arrivals caused by superposition of multiple S-wave reverberations and SV to P and/or P to SV conversions inside the whole crust |
| Sb   | (alt: S*) Either an upgoing S wave from a source in the lower crust or an S wave bottoming in the lower crust                                                                                                                                                                       |
| Sn   | Any S wave bottoming in the uppermost mantle or an upgoing S wave from a source in the uppermost mantle                                                                                                                                                                             |
| SnSn | Sn free-surface reflection                                                                                                                                                                                                                                                          |
| SgSg | Sg free-surface reflection                                                                                                                                                                                                                                                          |
| SmS  | S reflection from the outer side of the Moho                                                                                                                                                                                                                                        |
| SmSN | SmS multiple free-surface reflection; N is a positive integer. For example, SmS2 is SmSSmS                                                                                                                                                                                          |
| SmP  | S to P reflection/conversion from the outer side of the Moho                                                                                                                                                                                                                        |
| Lg   | A wave group observed at larger regional distances and caused by superposition of multiple S-wave reverberations and SV to P and/or P to SV conversions inside the whole crust. The maximum energy travels with a group velocity around 3.5 km/s                                    |
| Rg   | Short-period crustal Rayleigh wave                                                                                                                                                                                                                                                  |

### 5.2 MANTLE PHASES (record and ray tracing examples see Chapter 2, section 2.6.2; Chapter 11, section 11.5.2; DS 11.2; EX 11.2 and animations from 42° up to 152° via IS 11.3)

|    |                                                                                                                                       |
|----|---------------------------------------------------------------------------------------------------------------------------------------|
| P  | A longitudinal wave, bottoming below the uppermost mantle; also an upgoing longitudinal wave from a source below the uppermost mantle |
| PP | Free-surface reflection of a P wave leaving a source downward                                                                         |



|      |                                                                                                                                                                                                   |
|------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| PS   | P, leaving a source downward, reflected as an S at the free surface. At shorter distances the first leg is represented by a crustal P wave                                                        |
| PPP  | Analogous to PP                                                                                                                                                                                   |
| PPS  | PP which is converted to S at the second reflection point on the free surface; travel time matches that of PSP                                                                                    |
| PSS  | PS reflected at the free surface                                                                                                                                                                  |
| PcP  | P reflection from the core-mantle boundary (CMB)                                                                                                                                                  |
| PcS  | P converted to S when reflected from the CMB                                                                                                                                                      |
| PcPN | PcP reflected from the free surface N-1 times; N is a positive integer. For example PcP2 is PcPPcP                                                                                                |
| Pz+P | (alt: PzP) P reflection from outer side of a discontinuity at depth z; z may be a positive numerical value in km. For example, P660+P is a P reflection from the top of the 660 km discontinuity  |
| Pz-P | P reflection from inner side of a discontinuity at depth z. For example, P660-P is a P reflection from below the 660 km discontinuity, which means it is precursory to PP                         |
| Pz+S | (alt: PzS) P converted to S when reflected from outer side of a discontinuity at depth z                                                                                                          |
| Pz-S | P converted to S when reflected from inner side of a discontinuity at depth z                                                                                                                     |
| PScS | P (leaving a source downward) to ScS reflection at the free                                                                                                                                       |
| Pdif | (old: Pdiff) P diffracted along the CMB in the mantle                                                                                                                                             |
| S    | Shear wave, bottoming below the uppermost mantle; also an upgoing shear wave from a source below the uppermost mantle                                                                             |
| SS   | Free surface reflection of an S wave leaving a source downward                                                                                                                                    |
| SP   | S, leaving a source downward, reflected as P at the free surface. At shorter distances, the second leg is represented by a crustal P wave.                                                        |
| SSS  | Analogous to SS                                                                                                                                                                                   |
| SSP  | SS converted to P when reflected from the free surface; travel time matches that of SPS                                                                                                           |
| SPP  | SP reflected at the free surface                                                                                                                                                                  |
| ScS  | S reflection from the CMB                                                                                                                                                                         |
| ScP  | S converted to P when reflected from the CMB                                                                                                                                                      |
| ScSN | ScS multiple free-surface reflection; N is a positive integer. For example ScS2 is ScSScS                                                                                                         |
| Sz+S | (alt: SzS) S reflection from outer side of a discontinuity at depth z; z may be a positive numerical value in km. For example S660+S is an S reflection from the top of the 660 km discontinuity. |
| Sz-S | S reflection from inner side of a discontinuity at depth z. For example, S660-S is an S reflection from below the 660 km discontinuity, which it is precursory to SS                              |
| Sz+P | (alt: SzP) S converted to P when reflected from outer side of a discontinuity at depth z                                                                                                          |
| Sz-P | S converted to P when reflected from inner side of a discontinuity at depth z                                                                                                                     |
| ScSP | ScS to P reflection at the free surface                                                                                                                                                           |
| Sdif | (old: Sdiff) S diffracted along the CMB in the mantle                                                                                                                                             |

---

### 5.3 CORE PHASES (record examples see Chapter 2, section 2.6.2; Chapter 11, section 11.5.2, DS 11.3, EX 11.3 and animation files 5 and 6-10 via IS 11.3)

---

|         |                                                                                                                                                               |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------|
| PKP     | (alt: P') unspecified P wave bottoming in the core                                                                                                            |
| PKPab   | (old: PKP2) P wave bottoming in the upper outer core; ab indicates the retrograde branch of the PKP caustic                                                   |
| PKPbc   | (old: PKP1) P wave bottoming in the lower outer core; bc indicates the prograde branch of the PKP caustic                                                     |
| PKPdf   | (alt: PKIKP) P wave bottoming in the inner core                                                                                                               |
| PKPpre  | (old: PKhKP) a precursor to PKPdf due to scattering near or at the CMB                                                                                        |
| PKPdif  | P wave diffracted at the inner core boundary (ICB) in the outer core                                                                                          |
| PKS     | Unspecified P wave bottoming in the core and converting to S at the CMB                                                                                       |
| PKSab   | PKS bottoming in the upper outer core                                                                                                                         |
| PKSbc   | PKS bottoming in the lower outer core                                                                                                                         |
| PKSdf   | PKS bottoming in the inner core                                                                                                                               |
| P'P'    | (alt: PKPPKP) Free-surface reflection of PKP                                                                                                                  |
| P'N     | (alt: PKPN) PKP reflected at the free surface N-1 times; N is a positive integer. For example P'3 is P'P'P'                                                   |
| P'z-P'  | PKP reflected from inner side of a discontinuity at depth z outside the core, which means it is precursory to P'P'; z may be a positive numerical value in km |
| P'S'    | (alt: PKPSKS) PKP converted to SKS when reflected from the free surface; other examples are P'PKS, P'SKP                                                      |
| PS'     | (alt: PSKS) P (leaving a source downward) to SKS reflection at the free surface                                                                               |
| PKKP    | Unspecified P wave reflected once from the inner side of the CMB                                                                                              |
| PKKPab  | PKKP bottoming in the upper outer core                                                                                                                        |
| PKKPbc  | PKKP bottoming in the lower outer core                                                                                                                        |
| PKKPdf  | PKKP bottoming in the inner core                                                                                                                              |
| PNKP    | P wave reflected N-1 times from inner side of the CMB; N is a positive integer                                                                                |
| PKKPpre | a precursor to PKKPdf due to scattering near the CMB                                                                                                          |
| PKiKP   | P wave reflected from the inner core boundary (ICB)                                                                                                           |
| PKNIKP  | P wave reflected N-1 times from the inner side of the ICB                                                                                                     |
| PKJKP   | P wave traversing the outer core as P and the inner core as S                                                                                                 |
| PKKS    | P wave reflected once from inner side of the CMB and converted to S at the CMB                                                                                |
| PKKSab  | PKKS bottoming in the upper outer core                                                                                                                        |
| PKKSbc  | PKKS bottoming in the lower outer core                                                                                                                        |
| PKKSdf  | PKKS bottoming in the inner core                                                                                                                              |
| PcPP'   | (alt: PcPPKP) PcP to PKP reflection at the free surface; other examples are PcPS', PcSP', PcSS', PcPSKP, PcSSKP                                               |
| SKS     | (alt: S') unspecified S wave traversing the core as P                                                                                                         |
| SKSac   | SKS bottoming in the outer core                                                                                                                               |
| SKSdf   | (alt: SKIKS) SKS bottoming in the inner core                                                                                                                  |
| SPdifKS | (alt: SKPdifS) SKS wave with a segment of mantle-side Pdif at the source and/or the receiver side of the ray path                                             |
| SKP     | Unspecified S wave traversing the core and then the mantle as P                                                                                               |
| SKPab   | SKP bottoming in the upper outer core                                                                                                                         |
| SKPbc   | SKP bottoming in the lower outer core                                                                                                                         |
| SKPdf   | SKP bottoming in the inner core                                                                                                                               |

|        |                                                                                                                                                               |
|--------|---------------------------------------------------------------------------------------------------------------------------------------------------------------|
| S'S'   | (alt: SKSSKS) Free-surface reflection of SKS                                                                                                                  |
| S'N    | SKS reflected at the free surface N-1 times; N is a positive integer                                                                                          |
| S'z-S' | SKS reflected from inner side of a discontinuity at depth z outside the core, which means it is precursory to S'S'; z may be a positive numerical value in km |
| S'P'   | (alt: SKSPKP) SKS converted to PKP when reflected from the free surface; other examples are S'SKP, S'PKS                                                      |
| S'P    | (alt: SKSP) SKS to P reflection at the free surface                                                                                                           |
| SKKS   | Unspecified S wave traversing the core as P with one reflection from inner side of the CMB                                                                    |
| SKKSac | SKKS bottoming in the outer core                                                                                                                              |
| SKKSdf | SKKS bottoming in the inner core                                                                                                                              |
| SNKS   | S wave traversing the core as P, reflected N-1 times from inner side of the CMB; N is a positive integer                                                      |
| SKiKS  | S wave traversing the outer core as P and reflected from the ICB                                                                                              |
| SKJKS  | S wave traversing the outer core as P and the inner core as S                                                                                                 |
| SKKP   | S wave traversing the core as P with one reflection from the inner side of the CMB and then continuing as P in the mantle                                     |
| SKKPab | SKKP bottoming in the upper outer core                                                                                                                        |
| SKKPbc | SKKP bottoming in the lower outer core                                                                                                                        |
| SKKPdf | SKKP bottoming in the inner core                                                                                                                              |
| ScSS'  | (alt: ScSSKS) ScS to SKS reflection at the free surface; other examples are ScPS', ScSP', ScPP', ScSSKP, ScPSKP                                               |

---

#### 5.4 NEAR SOURCE SURFACE REFLECTIONS (Depth phases); also see Figure 10 in Chapter 2, section 2.6.3, several examples in DS 11.2 and 11.3 and the animation files 4 and 6 in IS 11.3

---

|      |                                                                                                                                                                                                                                                                                                |
|------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| pPy  | All P-type onsets (Py) as defined above, which resulted from reflection of an upgoing P wave at the free surface or an ocean bottom; WARNING: The character "y" is only a wild card for any seismic phase, which could be generated at the free surface. Examples are: pP, pPKP, pPP, pPcP etc |
| sPy  | All Py resulting from reflection of an upgoing S wave at the free surface or an ocean bottom. For example: sP, sPKP, sPP, sPcP etc.                                                                                                                                                            |
| pSy  | All S-type onsets (Sy) as defined above, which resulted from reflection of an upgoing P wave at the free surface or an ocean bottom. For example: pS, pSKS, pSS, pScP, etc.                                                                                                                    |
| sSy  | All Sy resulting from reflection of an upgoing S wave at the free surface or an ocean bottom. For example: sSn, sSS, sScS, sSdif, etc.                                                                                                                                                         |
| pwPy | All Py resulting from reflection of an upgoing P wave at the ocean's free surface                                                                                                                                                                                                              |
| pmPy | All Py resulting from reflection of an upgoing P wave from the inner side of the Moho                                                                                                                                                                                                          |

## 5.5 SURFACE WAVES

|     |                                                                                                                                                                    |
|-----|--------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| L   | Unspecified long-period surface wave                                                                                                                               |
| LQ  | Love wave                                                                                                                                                          |
| LR  | Rayleigh wave                                                                                                                                                      |
| G   | Mantle wave of Love type                                                                                                                                           |
| GN  | Mantle wave of Love type; N is integer and indicates wave packets traveling along the minor arcs (odd numbers) or major arc (even numbers) of the great circle     |
| R   | Mantle wave of Rayleigh type                                                                                                                                       |
| RN  | Mantle wave of Rayleigh type; N is integer and indicates wave packets traveling along the minor arcs (odd numbers) or major arc (even numbers) of the great circle |
| PL  | Fundamental leaking mode following P onsets generated by coupling of P energy into the waveguide formed by the crust and upper mantle                              |
| SPL | S wave coupling into the PL waveguide; other examples are SSPL, SSSPL                                                                                              |

## 5.6 ACOUSTIC PHASES

|     |                                                                                                                                                                                       |
|-----|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| H   | A hydroacoustic wave from a source in the water, which couples in the ground                                                                                                          |
| HPg | H phase converted to Pg at the receiver side                                                                                                                                          |
| HSg | H phase converted to Sg at the receiver side                                                                                                                                          |
| HRg | H phase converted to Rg at the receiver side                                                                                                                                          |
| I   | An atmospheric sound arrival, which couples in the ground                                                                                                                             |
| IPg | I phase converted to Pg at the receiver side                                                                                                                                          |
| ISg | I phase converted to Sg at the receiver side                                                                                                                                          |
| IRg | I phase converted to Rg at the receiver side                                                                                                                                          |
| T   | A tertiary wave. This is an acoustic wave from a source in the solid Earth, usually trapped in a low velocity oceanic water layer called the SOFAR channel (SOund Fixing And Ranging) |
| TPg | T phase converted to Pg at the receiver side                                                                                                                                          |
| TSg | T phase converted to Sg at the receiver side                                                                                                                                          |
| TRg | T phase converted to Rg at the receiver side                                                                                                                                          |

## 5.7 UNIDENTIFIED ARRIVALS

|    |                                                           |
|----|-----------------------------------------------------------|
| x  | (old: i, e, NULL) unidentified arrival                    |
| rx | (old: i, e, NULL) unidentified regional arrival           |
| tx | (old: i, e, NULL) unidentified teleseismic arrival        |
| Px | (old: i, e, NULL, (P), P?) unidentified arrival of P-type |
| Sx | (old: i, e, NULL, (S), S?) unidentified arrival of S-type |



## 5.8 AMPLITUDE MEASUREMENT PHASES

The following set of amplitude measurement names refers to the IASPEI (2013) Magnitude Standard, compliance to which is indicated by the presence of leading letter I. The absence of leading letter I indicates that a measurement is non-standard. Letter A indicates a measurement in nm made on a displacement seismogram, whereas letter V indicates a measurement in nm/s made on a velocity seismogram. All amplitude measurements for standard magnitudes, with the exception for ML, are made on vertical-component seismic records. For details on the measurement standards, e.g., on the required response simulation before measuring band-limited magnitudes like mb, mb(Lg), ML or Ms(20), the related measurements of periods, of the measurement time windows as well as on the relationship between standard and classical magnitudes see IS 3.3.

|         |                                                                                                                                                                             |
|---------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| IAML    | Displacement amplitude measured according to the IASPEI standard for local magnitude ML                                                                                     |
| IAMs_20 | Displacement amplitude measured according to IASPEI standard for surface-wave magnitude Ms(20)                                                                              |
| IVMs_BB | Velocity amplitude measured according to IASPEI standard for broadband surface-wave magnitude Ms(BB)                                                                        |
| IAmb    | Displacement amplitude measured according to IASPEI standard for short-period teleseismic body-wave magnitude mb                                                            |
| IVmB_BB | Velocity amplitude measured according to IASPEI standard for broadband teleseismic body-wave magnitude mB(BB)                                                               |
| IAmb_Lg | Displacement amplitude measured according to IASPEI standard for short-period Lg-wave magnitude mb(Lg)                                                                      |
| AX_IN   | Displacement amplitude of phase of type X (e.g., P, PP, S, etc.), measured on an instrument of type IN (wild card) (e.g., SP, short-period; LP, long-period; BB, broadband) |
| VX_IN   | Velocity amplitude of phase of type X and instrument of type IN (wild card as above)                                                                                        |
| A       | Unspecified displacement amplitude measurement                                                                                                                              |
| V       | Unspecified velocity amplitude measurement                                                                                                                                  |
| AML     | Displacement amplitude measurement for nonstandard local magnitude                                                                                                          |
| AMs     | Displacement amplitude measurement for nonstandard surface-wave magnitude                                                                                                   |
| Amb     | Displacement amplitude measurement for nonstandard short-period body-wave magnitude                                                                                         |
| AmB     | Displacement amplitude measurement for nonstandard medium to long-period body-wave magnitude                                                                                |
| END     | Time of visible end of record for duration magnitude                                                                                                                        |

## References

- Aki, K., and Richards, P. G. (2002). *Quantitative seismology*. Second Edition, ISBN 0-935702-96-2, University Science Books, Sausalito, CA., xvii + 700 pp.
- Aki, K., and Richards, P. G. (2009). *Quantitative seismology*. 2. ed., corr. Print., Sausalito, Calif., Univ. Science Books, 2009, XVIII + 700 pp., ISBN 978-1-891389-63-4 (see [http://www.ldeo.columbia.edu/~richards/Aki\\_Richards.html](http://www.ldeo.columbia.edu/~richards/Aki_Richards.html)).
- Angenheister, G. H. (1921). Beobachtungen an pazifischen Beben. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*, Mathematisch-physikalische Klasse, 113-146.
- Bastings, L. (1934). Shear waves through the Earth's core. *Nature*, **134**, 216-217.
- Borne, G. von dem (1904). Seismische Registrierungen in Göttingen, Juli bis Dezember 1903. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*, Mathematisch-physikalische Klasse, 440-464.
- Bullen, K. E. (1946). A hypothesis on compressibility at pressures of the order of a million atmospheres. *Nature*, **157**, 405.
- Chapman, C. H. (1978). A new method for computing synthetic seismograms. *Geophys. J. Royal Astron. Soc.*, **54**, 481-518.
- Conrad, V. (1925). Laufzeitkurven des Tauernbebens vom 28. November, 1923. *Mitteilungen der Erdbeben-Kommission der Akademie der Wissenschaften in Wien*, Neue Folge **59**.
- Dey-Sarkar, S. K., and Chapman, C. H. (1978). A simple method for the computation of body wave seismograms. *Bull. Seism. Soc. Am.*, **68**, 1577-1593.
- Geiger, L. (1909). Seismische Registrierungen in Göttingen im Jahre 1907 mit einem Vorwort über die Bearbeitung der Erdbebendiagramme, *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*, Mathematisch-physikalische Klasse, 107-151.
- Geiger, L., and Gutenberg, B. (1912a). Konstitution des Erdinnern, erschlossen aus der Intensität longitudinaler und transversaler Erdbebenwellen, und einige Beobachtungen an den Vorläufern. *Physikalische Zeitschrift*, **13**, 115-118.
- Geiger, L., and Gutenberg, B. (1912b). Ueber Erdbebenwellen. VI. Konstitution des Erdinnern, erschlossen aus der Intensität longitudinaler und transversaler Erdbebenwellen, und einige Beobachtungen an den Vorläufern. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*. Mathematisch-physikalische Klasse, 623-675.
- Gutenberg, B. (1925). Bearbeitung von Aufzeichnungen einiger Weltbeben. *Abhandlungen der Senckenbergischen Naturforschenden Gesellschaft*, **40**, 57-88.
- Gutenberg, B., and Richter, C. F. (1934). On seismic waves (first paper). *Gerl. Beitr. Geophysik*, **43**, 56 - 133.
- Gutenberg, B., Wood, H. O., and Richter, C. F. (1933). Re suggestion by Dr. Harold Jeffreys regarding  $P$  and  $Pg$ . *Gerl. Beitr. Geophysik*, **40**, 97-98.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf).
- IASPEI (2013) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)

- Jeffreys, H. (1926). On near earthquakes. *Month. Not. Roy. Astr. Soc., Geophys. Suppl.*, **1**, 385-402.
- Jeffreys, H., and Bullen, K. E. (1940, 1948, 1958, 1967, and 1970). *Seismological Tables*. British Association for the Advancement of Science, Gray Milne Trust, London, 50 pp.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, **122**, 108-124.
- Linehan, D. (1940). Earthquakes in the West Indian region. *Trans. Am. Geophys. Union, EOS* **30**, 229-232.
- Macelwane J. B., Brunner, G. J., and Joliat, J. S. (1933). Re suggestion by Doctor Harold Jeffreys and others regarding P and Pg. *Gerlands Beiträge zur Geophysik*, **40**, 98.
- Mohorovičić, A. (1910). Potres od 8. X 1909. God. *Izvešće Zag. met. Ops. Zag.* 1909, Zagreb (Das Beben vom 8. X 1909. Jahrbuch des meteorologischen Observatoriums in Zagreb für das Jahr 1909), **9**, Part 4, 1-63.
- Scraser, F. J. (1931). The reflected waves from deep focus earthquakes. *Proc. Roy. Soc. of London A*-**132**, 213-235.
- Sohon, F. W. (1932). Seismometry. Part II, of Macelwane, J.B., and F.W. Sohon: *Introduction to theoretical seismology*, New York 1932, 9+149 pp.
- Somville, O. (1930). A propos d'une onde longue dans la première phase de quelques séismogrammes. *Gerl. Beitr. Geophysik*, **27**, 437-442.
- Somville, O. (1931). A propos d'une onde longue dans la première phase de quelques séismogrammes (II). *Gerl. Beitr. Geophysik*, **29**, 247-251.
- Stechschulte, V. C. (1932). The Japanese earthquake of March 29, 1928. *Bull. Seism. Soc. Am.*, **22**, 81-137.
- Storchak, D. A., Bormann, P., and Schweitzer, J. (2002). Standard nomenclature of seismic phases. In: Bormann, P. (eds.), *New Manual of Seismological Observatory Practice*, GeoForschungsZentrum, Potsdam, Vol. 2, IS2.1, 1-18.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2003). The IASPEI standard seismic phase list. *Seismological Research Letters*, **74**, 761-772.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2011). Seismic phase names: IASPEI standards. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, Vol. 2, 1162-1173; doi: 10.1007/978-90-481-8702-7.
- Summary of the Bulletin of the International Seismological Centre, 2010, January-June (2013), 47, 1-6, 51-57.
- Willmore, P. L. (1979). *Manual of Seismological Observatory Practice*. World Data Center A for Solid Earth Geophysics, Report SE-20, September 1979, Boulder, Colorado, 165 pp.

# Chapter 3

## Seismic Sources and Source Parameters

(Version August 2013; DOI: 10.2312/GFZ.NMSOP-2\_ch3)

Peter Bormann<sup>1)</sup>, Siegfried Wendt<sup>2)</sup>, and Domenico Di Giacomo<sup>3)</sup>

- <sup>1)</sup> Formerly GFZ German Research Centre for Geosciences, Department 2: Physics of the Earth, Telegrafenberg, 14473 Potsdam, Germany; E-mail: [pb65@gmx.net](mailto:pb65@gmx.net);  
<sup>2)</sup> Geophysical Observatory Collm, University of Leipzig, 04779 Wermsdorf, Germany; E-mail: [wendt@rz.uni-leipzig.de](mailto:wendt@rz.uni-leipzig.de)  
<sup>3)</sup> International Seismological Centre (ISC), Pipers Lane, Thatcham, Berkshire, RG19 4NS, United Kingdom; E-mail: [domenico@isc.ac.uk](mailto:domenico@isc.ac.uk)

**Note:** Figure and section numbers followed by “??” relate to Chapter 11 of NMSOP-1 (2002). They are accessible via this web site too. The numbers will be modified when the revised Chapter 11 becomes available.

|                                                                                                                         | Page |
|-------------------------------------------------------------------------------------------------------------------------|------|
| <b>3.1 Introduction to seismic sources and source parameters (P. Bormann)</b>                                           | 3    |
| 3.1.1 Types and peculiarities of seismic source processes                                                               | 4    |
| 3.1.1.1 Tectonic earthquakes                                                                                            | 4    |
| 3.1.1.2 Volcanic earthquakes                                                                                            | 7    |
| 3.1.1.3 Explosions, implosions, induced and triggered seismic events                                                    | 8    |
| 3.1.1.4 Microseisms                                                                                                     | 12   |
| 3.1.2 Parameters which characterize size, strength and mechanism of seismic sources                                     | 13   |
| 3.1.2.1 Macroseismic intensity                                                                                          | 13   |
| 3.1.2.2 Magnitude and early relationships to seismic energy                                                             | 13   |
| 3.1.2.3 Scalar seismic and geometric moment, source spectrum, source dimension and stress drop in relation to magnitude | 14   |
| 3.1.2.4 The influence of rupture velocity and duration on magnitude estimates                                           | 19   |
| 3.1.2.5 Factors which influence energy radiation and seismic efficiency                                                 | 22   |
| 3.1.2.6 Parameters which describe and control the source mechanism                                                      | 25   |
| 3.1.3 Mathematical, physical and geological representation of earthquakes                                               | 26   |
| 3.1.4 Empirical analysis of rupture geometry, kinematics and dynamics in space and time                                 | 27   |
| 3.1.5 Summary and conclusions                                                                                           | 35   |
| <b>3.2 Magnitude of seismic events (P. Bormann)</b>                                                                     | 36   |
| 3.2.1 History, scope and limitations of the magnitude concept                                                           | 36   |
| 3.2.2 General assumptions and definition of magnitude                                                                   | 41   |
| 3.2.3 General rules and procedures for magnitude measurement                                                            | 45   |
| 3.2.3.1 General rules and procedures when working with analog data or on digital screen plots                           | 45   |
| 3.2.3.2 Aim and specified procedures for IASPEI standard magnitudes                                                     |      |

|         |                                                                                                                                    |     |
|---------|------------------------------------------------------------------------------------------------------------------------------------|-----|
|         | and possible modifications in automated procedures                                                                                 |     |
|         | (P. Bormann, J. Saul and S. Wendt)                                                                                                 | 50  |
| 3.2.4   | Magnitude scales for local and regional events                                                                                     | 59  |
| 3.2.4.1 | The original Richter magnitude scale ML                                                                                            | 60  |
| 3.2.4.2 | The new IASPEI standard ML                                                                                                         | 64  |
| 3.2.4.3 | Other Ml scales based on amplitude measurements                                                                                    | 65  |
| 3.2.4.4 | mb_Lg                                                                                                                              | 67  |
| 3.2.4.5 | Duration and coda magnitudes (Md and Mc)                                                                                           | 70  |
| 3.2.4.6 | The Russian K-class system for classifying earthquakes                                                                             | 75  |
| 3.2.4.7 | The Japanese $M_{JMA}$ scale                                                                                                       | 78  |
| 3.2.5   | Common teleseismic magnitude scales (P. Bormann and S. Wendt)                                                                      | 78  |
| 3.2.5.1 | Surface wave magnitude scales $M_s$ and the new IASPEI standards                                                                   | 80  |
|         | Question 1: Are the new IASPEI $M_s$ standards compatible with Gutenberg $M_s$ ?                                                   | 84  |
|         | Question 2: Is $M_s$ measured on horizontal and vertical components compatible?                                                    | 87  |
|         | Question 3: How reliably do the calibration formulas for $M_s$ determination compensate for the distance dependence of amplitudes? | 87  |
|         | Question 4: Why is there no depth term in the $M_s$ calibration functions?                                                         | 88  |
|         | Question 5: Are depth correction terms for $M_s$ available and feasible?                                                           | 89  |
|         | Question 6: How do travel paths affect the period at which $M_s_{BB}$ is measured?                                                 | 90  |
|         | Question 7: Why does the IASPEI $M_s$ formula not include a period dependence?                                                     | 92  |
|         | Question 8: Exist $M_s$ scales with a frequency-dependent calibration term?                                                        | 93  |
|         | Question 9: How do lateral velocity inhomogeneities affect the $M_s$ estimates?                                                    | 94  |
|         | Question 10: How do deviations from the standard instrument responses affect $M_s_{BB}$ and $M_s_{20}$ ?                           | 97  |
| 3.2.5.2 | Body-wave magnitude scales and the new IASPEI standards                                                                            | 97  |
|         | Question 1: Is modern $mB_B$ compatible with the classical Gutenberg $mB$ ?                                                        | 102 |
|         | Question 2: How relates standard $mb$ to $mb(NEIC)$ ?                                                                              | 103 |
|         | Question 3: How relate standard $mb$ and $mB_{BB}$ to each other?                                                                  | 105 |
|         | Question 4: How does the large scatter of period readings in $BB$ records affect the estimate of $mB_{BB}$ ?                       | 106 |
|         | Question 5: How do deviations from the standard responses affect $mb$ and $mB_{BB}$ ?                                              | 107 |
|         | Question 6: To what extent are differences between $mb$ and $mB_{BB}$ due to frequency-dependent attenuation?                      | 107 |
|         | Question 7: How reliable are the Gutenberg-Richter $Q(\Delta, h)_{PV}$ curves/tables?                                              | 108 |
|         | • Still open problems                                                                                                              | 109 |
| 3.2.6   | Other amplitude, period, intensity or tsunami based magnitude scales                                                               | 109 |
|         | (P. Bormann)                                                                                                                       |     |
| 3.2.6.1 | Broadband and spectral P-wave magnitude scales                                                                                     | 110 |
|         | • Broadband P-wave magnitude scale for intermediate and deep earthquakes                                                           | 110 |
|         | • Spectral P-wave magnitudes                                                                                                       | 111 |
| 3.2.6.2 | 1-Hz P-wave magnitude $P(\Delta, h)$                                                                                               | 112 |
| 3.2.6.3 | Short-period PKP-wave magnitude                                                                                                    | 115 |
| 3.2.6.4 | High-frequency moments and magnitudes                                                                                              | 115 |
| 3.2.6.5 | Earthquake early-warning magnitudes                                                                                                | 116 |
| 3.2.6.6 | Macroseismic magnitude                                                                                                             | 119 |
| 3.2.6.7 | Tsunami magnitude                                                                                                                  | 124 |
| 3.2.7   | Non-saturating magnitude scales $M_w$ and $M_e$                                                                                    |     |
|         | (P. Bormann and D. Di Giacomo)                                                                                                     | 124 |
| 3.2.7.1 | Moment magnitude $M_w$                                                                                                             | 124 |
| 3.2.7.2 | Energy magnitude $M_e$                                                                                                             | 126 |
| 3.2.8   | Complementary non-saturating magnitude scales (P. Bormann)                                                                         | 133 |
| 3.2.8.1 | Cumulative body-wave magnitude $mB_c$                                                                                              | 133 |

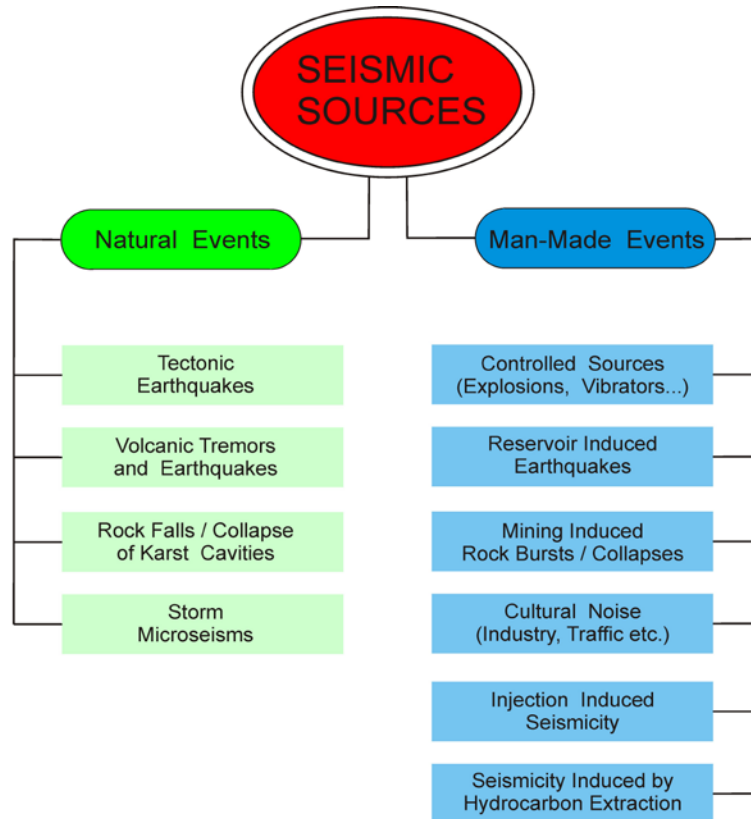
|            |                                                                                                                                                                  |         |
|------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------|---------|
| 3.2.8.2    | Mwp                                                                                                                                                              | 135     |
| 3.2.8.3    | M <sub>ED</sub> and Mwpd                                                                                                                                         | 137     |
| 3.2.8.4    | Hara's maximum amplitude-duration magnitude M                                                                                                                    | 140     |
| 3.2.8.5    | Mantle magnitude Mm                                                                                                                                              | 140     |
| 3.2.8.6    | W phase rapid Mw estimates and moment tensor solutions                                                                                                           | 141     |
| 3.2.9      | Relationships among different magnitude scales                                                                                                                   | 144     |
| 3.2.9.1    | Remarks about regression and correlation relationships<br>(P. Bormann)                                                                                           | 144     |
| 3.2.9.2    | Summary of classical magnitude relationships (P. Bormann)                                                                                                        | 152     |
| 3.2.9.3    | Linear orthogonal and standard regressions between IASPEI<br>standard magnitudes (P. Bormann, D. Di Giacomo and S. Wendt)                                        | 156     |
| 3.2.9.4    | Linear orthogonal and standard regressions between IASPEI<br>standard magnitudes, classical mb and Ms with Mw and Me<br>(P. Bormann, D. Di Giacomo and S. Wendt) | 162     |
| 3.2.9.5    | Non-linear ISC-GEM relationships between Ms and mb with<br>with Mw (D. Di Giacomo and P. Bormann)                                                                | 173     |
| 3.2.9.6    | Relationships for converting local and regional magnitudes<br>into Mw (P. Bormann)                                                                               | 175     |
| 3.2.10     | Summary remarks about magnitudes and their perspective (P. Bormann)                                                                                              | 177     |
| <b>3.3</b> | <b>Similarity conditions and seismic scaling relations</b>                                                                                                       | 182     |
| 3.3.1      | Similarity of seismic sources and the definition of seismic scaling<br>relations (P. Bormann)                                                                    | 182     |
| 3.3.2      | Relationships between seismic energy and magnitudes<br>(D. Di Giacomo and P. Bormann)                                                                            | 185     |
| 3.3.3      | Relationships between M <sub>0</sub> , E <sub>S</sub> and magnitudes<br>(P. Bormann and D. Di Giacomo)                                                           | 192     |
| 3.3.4      | Scaling relations of magnitudes, M <sub>0</sub> and E <sub>S</sub> with fault parameters<br>(P. Bormann)                                                         | 202     |
| <b>3.4</b> | <b>Determination of fault-plane solutions</b> (P. Bormann and S. Wendt)                                                                                          | 209     |
| 3.4.1      | Introduction                                                                                                                                                     | 209     |
| 3.4.2      | Determination of fault-plane solutions from P-wave polarities                                                                                                    | 215     |
| 3.4.3      | Accuracy of fault-plane solutions                                                                                                                                | 225     |
| 3.4.4      | Computer-assisted fault-plane solutions (P. Bormann)                                                                                                             | 226     |
| 3.4.5      | Estimating seismic moment, the size of rupture area, average slip<br>and stress drop from measured seismic spectra (P. Bormann)                                  | 228     |
|            | <b>Comments and acknowledgments</b>                                                                                                                              | 230     |
|            | <b>Recommended overview readings</b>                                                                                                                             | 231     |
|            | <b>References</b>                                                                                                                                                | 231-259 |

### 3.1 Introduction to seismic sources and source parameters (P. Bormann)

*Seismic waves* (Chapter 2), generated by *seismic sources* (Chapter 3), are oscillations due to elastic deformations which propagate through the Earth and can be recorded by *seismic sensors* (Chapter 5) and *data acquisition systems* (Chapter 6). The *seismic moment* (section 3.1.2.3; IS 3.8, 3.10 and 3.11) and *seismic energy* (section 3.1.2.5; IS 3.6 and 3.9) released by these sources may cover a tremendous range of associated *magnitudes* (see 3.2) and thus can stir up a wide range of shaking *intensities* (Chapter 12) with potentially related damages. These and other source characterizing parameters, their mutual relationships as well as the

determination of seismic source mechanisms based on observation of polarity patterns and amplitude ratios between P and S waves as a function of azimuth are dealt with in detail in this chapter, related Information Sheets (IS) and Exercises (EX).

### 3.1.1 Types and peculiarities of seismic source processes



**Fig. 3.1** Schematic classification of various kinds of events which generate seismic waves. To the “classical” man-made events one should nowadays also add induced seismicity due to the injection of liquids into the underground and the extraction of hydrocarbons.

#### 3.1.1.1 Tectonic earthquakes

Tectonic earthquakes are caused when the brittle part of the Earth’s crust is subjected to stress that exceeds its breaking strength. Sudden rupture will occur, mostly along pre-existing faults but sometimes along newly formed faults. Rocks on each side of the rupture “snap” into a new position. For very large earthquakes, the length of the ruptured zone may be as much as 1000 km and the slip along the fault can reach several meters, sometimes even over a decameter.

Laboratory experiments show that homogeneous consolidated rocks under pressure and temperature conditions at the Earth’s surface will fracture at a volume strain on the order of  $10^{-2}$  -  $10^{-3}$  (i.e., about 0.1 % to 1% volume change) depending upon their porosity. Rock strength is generally smaller under tension or shear than under compression. Shear strains on the order of about  $10^{-4}$  or less may already cause fracturing of solid brittle rock. Rock strength is further reduced if the rock is pre-fractured, which is usually the case in continental Earth crust subjected to millions of years of still ongoing or previous phases of increased tectonic activity.

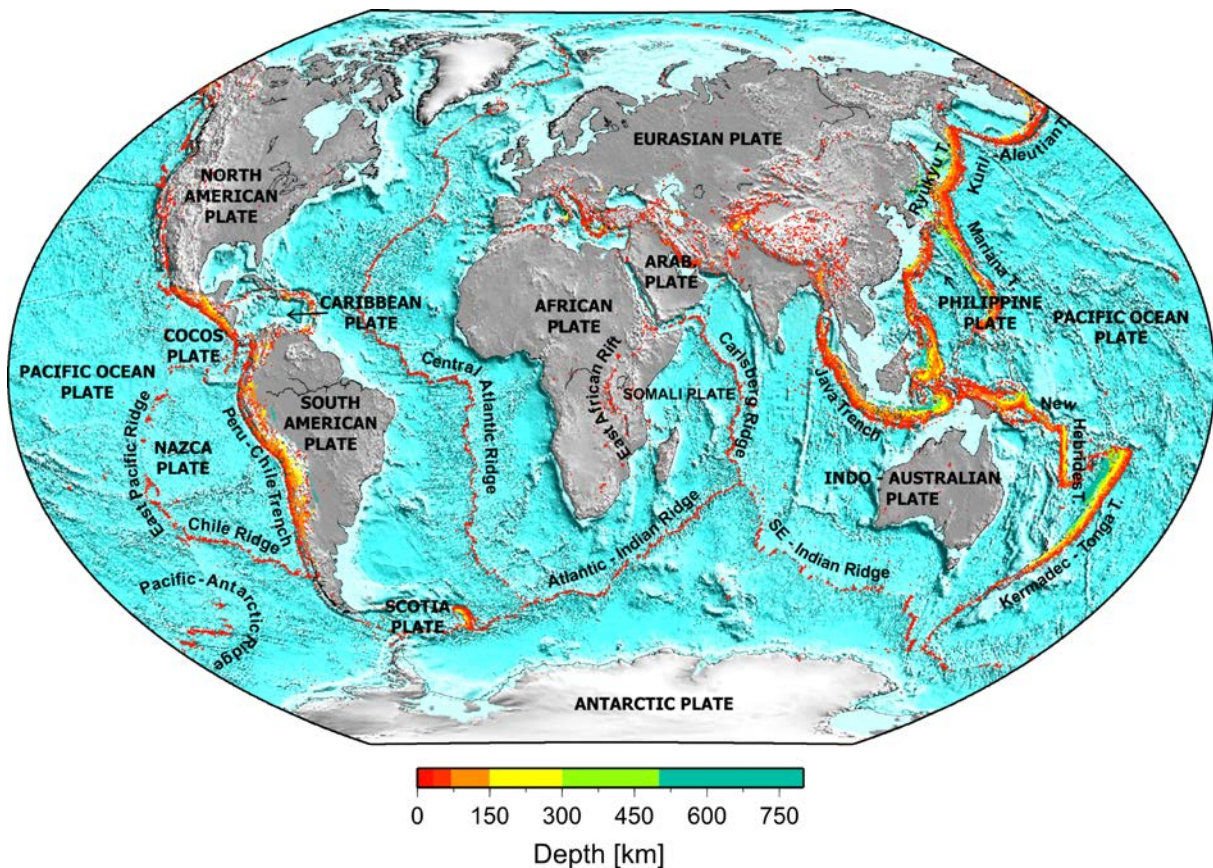
The strength of pre-fractured rock is much less than that of unbroken competent rock and is mainly controlled by the frictional resistance to motion of the two sides of the fault. Frictional resistance, which depends on the orientation of the faults with respect to the stress field and other conditions (see Scholz, 1990 and 2002), can vary over a wide range. Accordingly, deformations on the order of only  $10^{-5}$  to  $10^{-7}$ , which correspond to bending of a lithospheric plate by about 0.1 mm to 1 cm over a distance of 1 km, may cause shear faulting along pre-existing zones of weakness. But the shear strength depends also on the composition and fabric (anisotropy) of rock, its temperature, the confining pressure, the rate of deformation, etc. as well as the total cumulative strain. More details on the physics of earthquake faulting and related geological and seismotectonic conditions in the real Earth can be found in Scholz (1990 and 2002) and in section 3.1.4. Additional recommended overview articles on the rheology of the stratified lithosphere and its relation to crustal composition, age and heat flow were published by Meissner and Wever (1988), Ranalli and Murphy (1987) and Wever et al. (1987). They also explain the influence of these parameters on the thickness and maximum depth of the seismogenic zone in the Earth crust and lithosphere, i.e., the zone within which brittle fracturing of the rocks is possible when the strains exceed the breaking strength or elastic limit of the rock (see Chapter 2, Fig. 2.1).

Stress generated by the relative motions of lithospheric plates is the main cause of tectonic earthquakes. The plates are driven, pushed and pulled by the slow motion of convection currents in the more plastic hot material of the mantle beneath the lithosphere. These relative motions are in the order of several cm per year. Fig. 3.2 shows the global pattern of earthquake belts and the major tectonic plates. There are also numerous small plates called sub- or micro-plates. Shallow earthquakes, within the upper part of the crust, take place mainly at plate boundaries but may also occur inside plates (interplate and intraplate earthquakes, respectively). Intermediate (down to about 300 km) and deep earthquakes (down to a maximum of 700 km depth) occur under ocean trenches and related subduction zones where the lithosphere plates are thrust or pulled down into the upper mantle. The major trenches are found around the Circum-Pacific earthquake and volcanic belt (see Fig. 3.2). However, intermediate and deep earthquakes may occur also in some other marine or continental collision zones (e.g., the Tyrrhenian and Aegean Sea or the Carpathians and Hindu Kush, respectively).

Most earthquakes occur along the main plate boundaries. These boundaries constitute either zones of extension (e.g., in the up-welling zones of the mid-oceanic ridges or intra-plate rifts), transcurrent shear zones (e.g., the San Andreas fault in the west coast of North America or the North Anatolian fault in Turkey), or zones of plate collision (e.g., the Himalayan thrust front) or zones of subduction (mostly along deep sea trenches). Accordingly, tectonic earthquakes may be associated with many different faulting types (strike-slip, normal, reverse, thrust faulting or mixed; see text and Figures in 3.4.2).

The largest strain rates are observed near active plate boundaries (about  $10^{-8}$  to  $3 \times 10^{-10}$  per year). Strain rates are significantly less in active plate interiors (about  $5 \times 10^{-10}$  to  $3 \times 10^{-11}$  per year) or within stable continental platforms (about  $5 \times 10^{-11}$  to  $10^{-12}$  per year) (personal communication by Giardini, 1994). Consequently, the critical cumulative strain for the pre-fractured/faulted seismogenic zone of lithosphere, which is on the order of about  $10^{-6}$  to  $10^{-7}$ , is reached roughly after some 100, 1000 to 10,000 or 10,000 to 100,000 years of loading, respectively. This agrees well with estimates of the mean return period of the largest possible events (seismic cycles) in different plate environments (Muir-Wood, 1993; Scholz, 1990).





**Fig. 3.2** Global distribution of earthquake epicenters according to the data catalog of the United States National Earthquake Information Center (NEIC), January 1960 to November 2011, and the related major lithosphere plates. (Figure by courtesy of Regina Milkereit, GFZ German Research Centre for Geosciences).

Although there are hundreds of thousands of weak tectonic earthquakes globally every year, most of them can only be recorded by sensitive nearby instruments. But in the long-term global statistical average about 100,000 earthquakes are strong enough (magnitude  $M \geq 3$ ) to be potentially perceptible by humans in the near-source area. A few thousand are strong enough ( $M \geq 5$ ) to cause slight damage and some 100 with magnitude  $M > 6$  can cause heavy damage, if there are nearby settlements and built-up areas; while several events every year (with  $M \geq 8$ ) may result in wide-spread devastation and disaster. During the 20<sup>th</sup> century, the 1995 Great Hanshin/Kobe earthquake caused the greatest economic loss (about 100 billion US\$), the 1976 Tangshan earthquake inflicted the most terrible human loss (about 243,000 people killed) while the Chile earthquake of 1960 released the largest amount of *seismic energy*  $E_s$  of about  $5 \cdot 10^{18}$  to  $10^{19}$  Joule. The latter corresponds to about 25 to 100 years of the long-term annual average of global seismic energy release which is about  $1 - 2 \times 10^{17}$  J (Lay and Wallace, 1995) and to about half a year of the total kinetic energy contained in the global lithosphere plate motion. The total *seismic moment* (see 3.1.2.3. below) of the Chile earthquake was about  $3 \cdot 10^{23}$  Nm. It ruptured about 800 - 1000 km of the subduction zone interface at the Peru-Chile trench in a width of about 200 km (Boore, 1977; Scholz, 1990).

The commonly accepted *moment magnitude* value of  $M_w = 9.5$  has been published first by Kanamori and Cipar (1974). Considering also later papers, the ISC-GEM Global Instrumental Catalogue 1900-2009 (<http://www.isc.ac.uk/iscgem/>) adopted  $M_w = 9.6$  with a large uncertainty of  $\pm 0.3$  due to deficiency of the then seismic instrumentation and the complexity of this earthquake. According to Cifuentes and Silver (1989) this great Valdivia May 22<sup>nd</sup> 1960 Chile earthquake was in fact a multiple rupture, consisting of three events with a combined seismic moment of at least  $M_0 = 5.5 \times 10^{23}$  Nm, corresponding to  $M_w = 9.76$ , and an overall duration of about 1500 s. The faster rupturing main shock was preceded by a large scale slip with a rise time of 300 s and observable only at low frequencies, starting some 1150 s before the more high-frequency main shock initiated and another event that followed the main shock about 350 s later. When fully accounting for the probable very long-period energy release, associated with the long lasting slow rupture, which was not yet properly measurable with the 1960 available instrumentation, even  $M_w \approx 9.9$  might be conceivable. If this can be confirmed, then about 4 times more seismic energy than discussed above was released in this huge seismotectonic event.

In summary: about 85 % of the total world-wide seismic moment release by earthquakes occurs in subduction zones and more than 95 % by shallow earthquakes along plate boundaries. The other 5 % are distributed between intraplate events and deep and intermediate focus earthquakes. The single 1960 Chile earthquake accounts for about 25 % of the total seismic moment release between 1904 and 1986. And some of the devastating great earthquakes of the early 21<sup>st</sup> century were almost comparably large, such as the  $M_w 9.3$  Sumatra-Andaman earthquake of December 26, 2004 with a rupture duration of about 9 min, a rupture length of about 1300 km and with some 15 to 20 m fault slip, causing a tsunami with maximum run-up heights on land of almost 30 m locally, killing in total more than 230,000 people ([http://wikipedia.org/wiki/2004 Indian Ocean earthquake and tsunami](http://wikipedia.org/wiki/2004_Indian_Ocean_earthquake_and_tsunami); last accessed 03.07.2013).

It should be noted that most of the total energy released by seismic sources,  $E_T$ , e.g., the total available elastic strain energy, chemical, nuclear or other stored energy, is required to power the growth of the earthquake fracture, to produce frictional heat, sound or other. Only a small fraction of  $E_T$  goes into producing seismic waves. The seismic efficiency, i.e., the ratio of  $E_S/E_T$ , is usually only about 0.01 to 0.1. It depends both on the *stress drop* during the rupture as well as on the total stress in the source region (Spence, 1977; Scholz, 1990), the ambient medium and other factors (see sections 3.1.1.3 and 3.1.2.5).

### 3.1.1.2 Volcanic earthquakes

Although the total energy released by the strongest historically known volcanic eruptions was even larger than  $E_T$  of the Chile earthquake, the seismic efficiency of volcanic eruptions is generally much smaller, due to their long duration. Nevertheless, volcanic earthquakes of the tectonic type triggered by changes in the failure strength of the rock-fault system of the volcanic edifice or in its neighborhood due to volcanic processes, may in some cases locally reach the shaking strength of destructive earthquakes (e.g., with *magnitudes* of about 6). However, most of the seismic oscillations measured at volcanoes are of the tremor type and produced in conjunction with sub-surface magma flows. Tremors are long-lasting and more or less monochromatic oscillations which come from a two- or three-phase (liquid- and/or gas-solid) source process which is not narrowly localized in space and time. They cannot be analyzed in the traditional way of seismic recordings from tectonic earthquakes or explosions

nor be described with traditional source parameters (see Chapter 13). Such volcanic earthquakes contribute only an insignificant amount to the global seismic moment release (see Scholz, 1990). However, perturbations of the regional stress field by magmatic processes associated with an eruption or its preparation may also trigger tectonic earthquakes in the wider surroundings as part of the overall volcanic episode.

Explosions are mostly anthropogenic, i.e., “man-made”, and controlled, i.e., with known location and source time. However, strong natural explosions in conjunction with volcanic eruptions or meteorite impacts, such as the Tunguska meteorite of 30 June 1908 in Siberia or the Chelyabinsk/Ural meteorite impact on 15 February 2013 (see seismic records in Fig. 3.3a), may also occur.

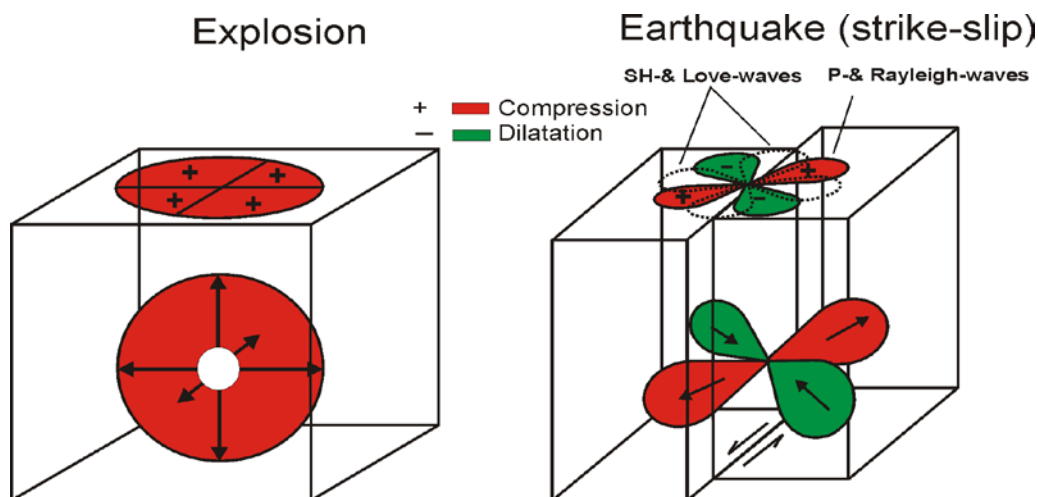
**Fig. 3.3a** Seismic records of the Chelyabinsk/Ural meteorite impact of February 15, 2013, compiled at the GFZ German Research Centre for Geosciences (GFZ).



$10^{-3}$ , i.e., only 1 % to 0.1 % of the total released explosion energy is transformed into seismic energy). By means of decoupling techniques  $\varepsilon$  may be reduced by 2 to 5 orders (Adams and Allen, 1961; Adams and Swift, 1961; Latter et al., 1961a and b; Willies and Wilson, 1962). The coupling factor of explosions on the surface or in the atmosphere is much less ( $\varepsilon \approx 10^{-3}$  to  $10^{-6}$  depending on the altitude) and in the hydrosphere much larger ( $\varepsilon \approx 10^{-1}$ ) (see compilation of  $\varepsilon$  values in Bormann, 1966; or Båth, 1962; Griggs and Press, 1961; Pasečnik et al., 1960; Pomeroy and Oliver, 1960; Willies, 1963).

Explosions are expected to produce an outward directed compressional first motion in all directions while tectonic earthquakes produce first motions of different amplitude and polarity in different directions (Fig. 3.3b). These characteristics can be used to identify the type of source process (see 3.4) and to discriminate between explosions and tectonic earthquakes.

Compared to tectonic earthquakes, the *duration* of the source process of explosions and the *rise time* to the maximum level of displacement is much shorter (milliseconds as compared to seconds up to a few minutes) and more impulsive (Fig. 3.4). Accordingly, explosions of comparable body-wave magnitude excite more high-frequency oscillations and source spectra with about one order higher corner frequencies than average earthquakes having the same level of low-frequency energy (see Fig. 3.5).



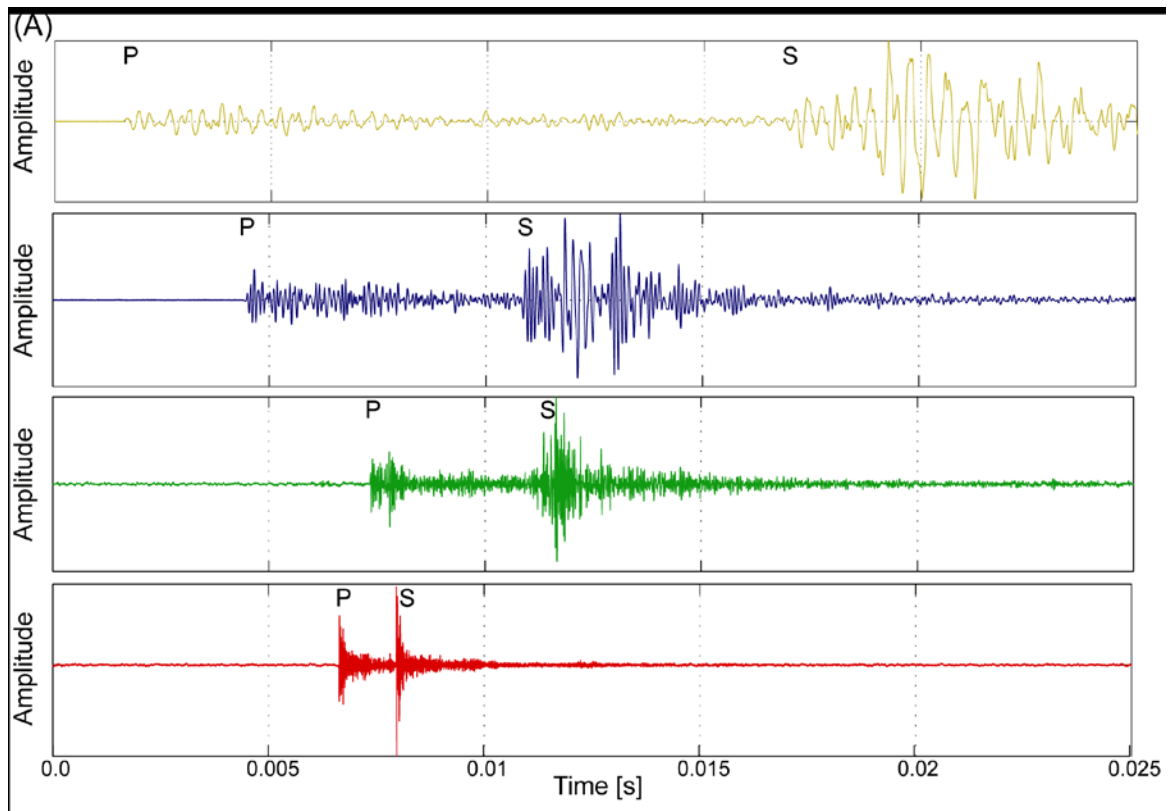
**Fig. 3.3b** Schematic sketches of an idealized underground explosion and of a strike-slip earthquake along a vertically dipping fault. The fault motion is "left-lateral", i.e., counter-clockwise. The arrows show the directions of compressional (outward, polarity +, red shaded) and dilatational (inward, polarity -, green shaded) motions. The patterns shown on the surface, termed amplitude or polarity patterns, indicate the azimuthal variation of observed amplitudes or of the direction of first motions in seismic records (UP = + or DOWN = -), respectively. While point-like explosions in an isotropic medium should show no azimuth-dependent amplitudes and compressional first motions only, amplitudes and polarities vary for a tectonic earthquake. The dotted amplitude lobes in Fig. 3.3, right side, indicate qualitatively the different azimuth dependence of shear (S) waves as compared to longitudinal (P) waves (rotated by  $45^\circ$ ) but their absolute displacement amplitudes are – on average – about 5 times larger than those of P waves.

Rock falls may last for several minutes and cause seismic waves but generally with less distinct onsets and less separation of wave groups than tectonic earthquakes. The collapse of karst caves, mining-induced rock bursts or collapses of mining galleries are generally of an *implosion* type. Accordingly, their first motion patterns should show dilatations in all azimuths if a secondary tectonic event has not been triggered by the collapse. The strongest events may reach magnitudes up to about  $M = 5.5$  and be recorded world-wide (e.g., Bormann et al., 1992).

Besides this there are several other classes of induced and triggered seismicity. McGarr et al., (2002) summarize the case histories of several distinct stimulated seismic events and their causes, mainly due to changes in the ambient stress and/or pore pressure conditions in response to either other earthquakes in the surroundings or due to human activities. We shortly refer to 4 main types of such events which are:

- *reservoir induced earthquakes* which have frequently been observed in conjunction with the impoundment of water or rapid water level changes behind large dams. These events are triggered along pre-existing and pre-stressed tectonic faults, and they show the typical polarity patterns of tectonic earthquakes. The strongest event reported so far, the Koyna, India, earthquake of 1967 (Gupta and Rastogi, 1976) reached a magnitude of 6.5.
- *earthquakes induced by the injection of fluids* into a tectonically stressed crust. An interesting case is described by Horton (2012). Due to the disposal of hydrofracturing waste fluid, injected into aquifers at several hundred meters down to more than 3 km depth in Central Arkansas, an earthquake swarm was triggered in the surroundings of a major fault which has the potential to release a damaging earthquake.
- *earthquakes induced by hydrocarbon withdrawal* from the underground. Grasso and Wittlinger (1990) describe the result of 10 years monitoring of a gas field in France which has been seismically completely inactive prior to the gas extraction. Within the first 10 production years some 800 earthquakes with magnitudes ranging between 1.0 and 4.2 have been recorded. Pennington et al. (1986) relate to the pronounced depressuring of fault planes in oil and gas fields in Texas, where fluid pressures had dropped to less than 20% of their original values, producing earthquakes up to magnitude 3.9. For three more  $M \approx 6$  earthquakes in California, which occurred all directly over major fields of oil extraction (Coalinga, 1983; Kettleman North Dome, 1985; Whittier Narrows, 1987) it is suspected that they all have been induced by oil extraction over a still growing anticline (McGarr, 1991), although the focal depth of these events was some 8 km deeper. However, the ratio of net liquid production to total seismic moment released by all three events was nearly the same, as was the epicentral extension of the aftershock sequences of all three earthquakes which approximately coincided with the extension of the overlying oil fields. Similarly, the sequence of three  $M = 7$  earthquakes in April and Mai 1976 and in March 1984 in the Gazli region of Uzbekistan, with no known history of so strong earthquakes before the exploitation of the enormous gas resources there, had also been interpreted by Simpson and Leith (1985) as being likely related to the gas extraction. This supposition has also been supported by McGarr (1991) but questioned by Bossu (1996), who estimated that the total extracted mass of gas was only about  $1/4^{\text{th}}$  of the mass of the water which subsequently infiltrated the void space in the gas field.
- *earthquakes induced by stress changes* in the surrounding rocks caused by creating the mined-out void, often connected with blasting activity. Several such cases have been described in the literature. An early one relates to the East Rand Proprietary gold

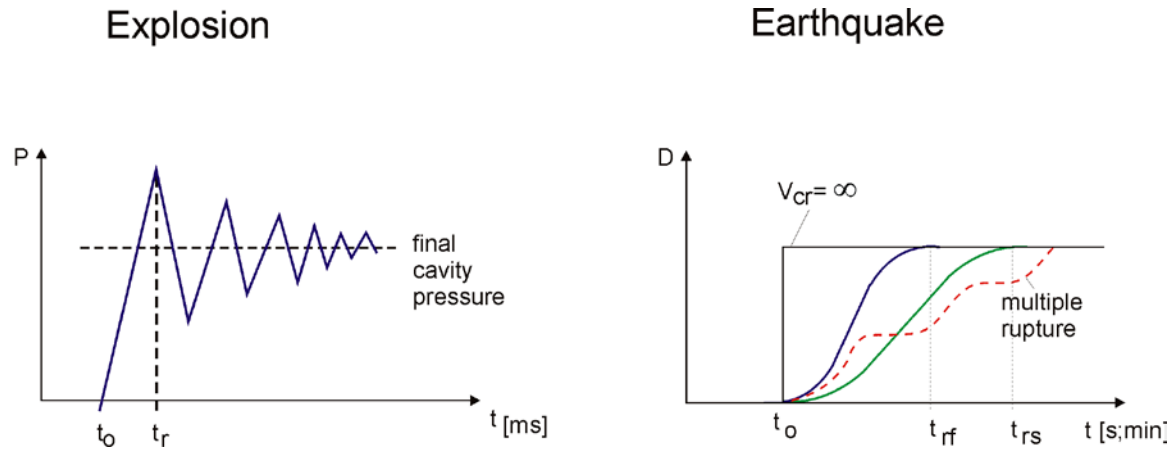
Mines (ERPM, McGarr et al., 1975), a very recent one to the Mponeng gold mine, (Kwiatek et al., 2010; Plenkers et al., 2010), both in South Africa. These events are usually rather small. In the first case, magnitudes between -1.5 and 2.5 had been recorded, with the much more sensitive local underground network in 3.5 km depth in the Mponeng mine even events with  $M_w$  between -4.4 and -1.3. Monitoring of such production induced seismicity is also part of a mining safety alarm system. Since the events are so small and so close to the sensors (within meters to several ten meters), the frequencies recorded with acoustic emission sensors (AE) are very high, ranging between about 0.7 kHz and 200 kHz, yet their records show distinct P and S phases (Fig. 3c).



**Fig. 3.3c** Uncorrected waveforms in 25-ms windows for induced seismic events of different frequency content recorded at 3.5 km depth in the Mponeng gold mine, South Africa.

**Yellow:** low-frequency event ( $< 1$  kHz); **Blue:** event with dominating frequencies between 1 kHz and 25 kHz; **Green:** High-frequency (HF1) event with  $f > 25$  kHz; **Red:** Highest-frequency (HF2) events with  $f > 40$  kHz. Copy of Figure 2A of Plenkers et al. (2010), *Seism. Res. Lett.*, vol. 81, no. 3, p. 470; © Seismological Society of America.

However, earthquakes induced by stress redistribution are not a phenomena restricted to the very local range of mines and as a consequence of human action. Also stress redistribution after a strong earthquake may trigger in the prolongation of its own fault system or at adjacent ones another rather strong earthquake, which might have taken, without the preceding one, much longer to reach the critical state of rupture initiation. A good example is the  $M_w 7.2$  Düzce, Turkey, earthquake of 12 November 1999. It followed the  $M_w 7.4$  Izmit earthquake of 16 August 1999 along a nearby branching extension of the North Anatolian fault segment.



**Fig. 3.4** Schematic diagrams of the different source functions of explosions (left) and earthquakes (right).  $P$  - pressure in the explosion cavity,  $D$  - fault displacement,  $t$  - time,  $t_0$  - origin time of the event,  $t_r$  - rise time of  $P$  or  $D$  to its maximum values,  $t_{rf}$  - rise time of fast rupture,  $t_{rs}$  - rise time of slow rupture; the step function in the right diagram would correspond to an earthquake with infinite velocity of crack propagation  $v_{cr}$ . Common rupture models assume  $v_{cr}$  to be about 0.6 to 0.9 times of the velocity of shear-wave propagation,  $v_s$  (e.g., Madariaga, 1976), yet for very slow *tsunami earthquakes* it may be as slow as about 0.2  $v_s$  (Kanamori and Brodsky, 2004; Venkataraman and Kanamori, 2004b). Note the different time-scales for explosions and earthquakes.

#### 3.1.1.4 Microseisms and seismic noise

Seismic signals produced by storms over oceans or large water basins (seas, lakes, reservoirs) as well as by wind action on topography, vegetation or built-up surface cover are called *microseisms*. In contrast, seismic signals due to human activities such as rotating or hammering machinery, traffic etc., are *cultural seismic noise*. Rushing waters or gas/steam (in rivers, water falls, dams, pipelines, geysers) may be additional sources of natural or anthropogenic *seismic noise*. They are neither well localized in space nor fixed to a defined *origin time*. Accordingly, they produce more or less permanent on-going non-coherent interfering signals of more or less random amplitude fluctuations in a very wide frequency range of about 16 octaves (from about 50 Hz to 1 mHz) which are often controlled in their intensity by the season (natural noise) or time of day (anthropogenic noise). Despite the large range of ambient noise displacement amplitudes (about 6 to 10 orders of magnitude; see Chapter 4, Fig. 4.3) they are generally much smaller than those of earthquakes and not felt by people.

Differences between signals from coherent seismic sources on the one hand and microseisms/seismic noise on the other hand are dealt with in more detail in Chapter 4, filter, processing and installation procedures for the improvement of the signal-to-noise ratio (SNR) in seismic records are discussed in Chapters 4 and 7 and the use of ambient noise for microzonation is discussed in Chapter 14.

### 3.1.2 Parameters which characterize size, strength and mechanism of seismic sources

#### 3.1.2.1 Macroseismic intensity

The effect of a seismic source may be characterized by its *macroseismic intensity*, **I**. Intensity describes the strength of shaking in terms of human perception, damage to buildings and other structures, as well as changes in the surrounding environment. **I** depends on the distance from the source and the soil conditions and is mostly classified according to macroseismic scales of 12 degrees (e.g., Wood and Newman, 1931; Medvedev, 1962; Stover and Coffman, 1993; Grünthal, 1998). Exceptions are, e.g., the 7 degrees scales of Japan (see Chapter 12) and of Taiwan (Wu et al., 2003). From an analysis of the areal distribution of felt reports and damage one can estimate the epicentral intensity **I**<sub>0</sub> in the source area as well as the source depth, *h* (e.g., Sponheuer, 1960). There exist empirical relationships between **I**<sub>0</sub> and other instrumentally determined measures of the earthquake size such as the *magnitude*, peak ground velocity (PGV) and peak ground acceleration (PGA). Although these relationships are rather noisy it has been found that the earthquake damage statistics is much closer correlated with the PGV than with PGA (Wu et al., 2003). And magnitude-wise **I** is best correlated with local magnitude *M*<sub>l</sub> and energy magnitude *M*<sub>e</sub>, provided that the effect of distance between the source and the site where **I** is observed, is accounted for. Because, for two earthquakes with a given *M*<sub>w</sub>, the earthquake with the lower *M*<sub>e</sub> and *M*<sub>l</sub> tends to have a larger source, so that there may be more locations situated very close to a causative fault. Also, very shallow earthquakes tend to have relatively low *M*<sub>e</sub>, yet they are often intensively destructive in the immediate epicentral area or along the outcropping fault trace. For more details on macroseismic effects see Chapter 12 and for a comprehensive summary Grünthal (2011).

#### 3.1.2.2 Magnitude and early relationships to seismic energy

*Magnitude* is a logarithmic measure of the size of an earthquake or explosion based on instrumental measurements. The magnitude concept was first proposed by Richter (1935). Magnitudes are commonly derived from ground motion amplitudes and periods or from *signal duration* measured on instrumental records. There is no *a priori* scale limitation to magnitudes as it exists for macroseismic intensity scales. The largest moment magnitude, *M*<sub>w</sub>, observed so far, was that of the Chile earthquake in 1960 (*M*<sub>w</sub> ≈ 9.5 according to Kanamori, 1977). Nowadays, highly sensitive instrumentation close to the sources may record local events with magnitude even smaller than zero (e.g., *M*<sub>w</sub> down to -4.4 for induced seismicity recorded close-up in gold mines, Kwiatek et al., 2010). According to Richter's original definition these magnitude values become negative. With empirical *energy-magnitude-relationships* the *seismic energy*, *E*<sub>s</sub>, radiated by the seismic source as seismic waves can be estimated. Common relationships are those given by Gutenberg and Richter (1954, 1956a, b; Richter, 1958) between *E*<sub>s</sub> and the surface-wave magnitude *M*<sub>s</sub> and the body-wave magnitude *m*<sub>B</sub>:  $\log E_s = 11.8 + 1.5 M_s$  and  $\log E_s = 5.8 + 2.4 m_B$ , respectively (when *E*<sub>s</sub> is given in erg; 1 erg = 10<sup>-7</sup> Joule). According to these relationship, a change of *M*<sub>s</sub> or *m*<sub>B</sub> by one unit corresponds to a change in *E*<sub>s</sub> by a factor of about 32 times or 250 times, respectively.

Based on the analysis of digital recordings, there exist also direct procedures to estimate *E*<sub>s</sub> (e.g., Purcaru and Berckhemer, 1978; Seidl and Berckhemer, 1982; Boatwright and Choy, 1986; Kanamori et al., 1993; Choy and Boatwright, 1995) and to define an "energy



magnitude"  $M_e$  (see section 3.2.7.2). Because most of the seismic energy is concentrated in the higher frequencies around the *corner frequency*  $f_c$  of the radiated spectrum (see next section),  $M_e$  is a very suitable measure of the earthquakes' potential for damage. In contrast, the seismic moment, which is estimated from the displacement amplitude plateau at frequencies  $f \gg f_c$ ,  $M_0$  is related to the final static displacement after an earthquake and consequently, the moment magnitude,  $M_w$ , is more closely related to the tectonic effects of an earthquake (see Choy and Kirby, 2004; Di Giacomo et al., 2010; Bormann and Di Giacomo, 2011; IS 3.5).

### 3.1.2.3 Scalar seismic and geometric moment, source spectrum, source dimension and stress drop in relation to magnitude

Above we have already repeatedly referred to the terms *seismic moment* ( $M_0$ ) and *moment magnitude* ( $M_w$ ). The scalar *seismic moment*  $M_0$  (for its derivation see IS 3.1) is a measure of the "size" or "work" accomplished by a seismic shear source:

$$M_0 = \mu \bar{D} A \quad (3.1)$$

with  $\mu$  - rigidity or shear modulus of the source medium,  $\bar{D}$  - average final displacement after the rupture,  $A$  - the surface area of the rupture.  $M_0$  is a measure of the irreversible inelastic deformation in the rupture area. This inelastic strain is described in (3.1) by the product  $\bar{D} A$ . On the basis of reasonable average assumptions about  $\mu$  and the stress drop  $\Delta\sigma$  its determination is now standard in the routine analysis of strong earthquakes by means of waveform inversion of long-period digital records (see Dziewonski et al., 1981 and IS 3.8).

Although the scalar seismic moment provides a better physical quantification of the earthquake source than magnitude it is not directly measurable, since according to Ben-Zion (1989 and 2001) linear elastic waves generated by a slip source have no information on material properties at the source region itself. Only the product  $P = \bar{D} A$ , termed "geometric moment" by King (1978) or "potency  $P$ " by Ben-Menahem and Singh (1981), can be determined directly from the zero spectral asymptote of the source in a seismogram, respectively by integrating over the area underneath the restituted broadband displacement  $P$  and/or  $S$  waveforms recorded in the far-field of the source (e.g., Seidl and Berckhemer, 1982; Seidl and Hellweg (1988; see, e.g., Figs. 3.9, 3.63, 3.65a). The values for the rigidity  $\mu$ , however, has to be *assumed*. And different authors may assume in their moment calculations different values, e.g., when estimating  $M_0$  for earthquakes in very different seismotectonic environments and at very different source depths in the crust or upper mantle, in continental or oceanic areas. Then earthquakes of volumetrically equal size or potency have different seismic moment. Therefore, Ben-Zion (2001) recommends to convert reported moments to potencies by removing the assumed rigidities in order to make earthquakes sizes geometrically, respectively volumetrically comparable. In practice, however this often meets the difficulty that the rigidities assumed in the calculation of  $M_0$  are often not formally recorded. However, for southern California earthquakes with  $1.0 < M_l < 7.0$ , Ben-Zion and Zhu (2002) demonstrated a proper potency-magnitude scaling.

In a homogeneous half-space  $M_0$  can be determined from the spectra of seismic waves observed at the Earth's surface by using the relationship:

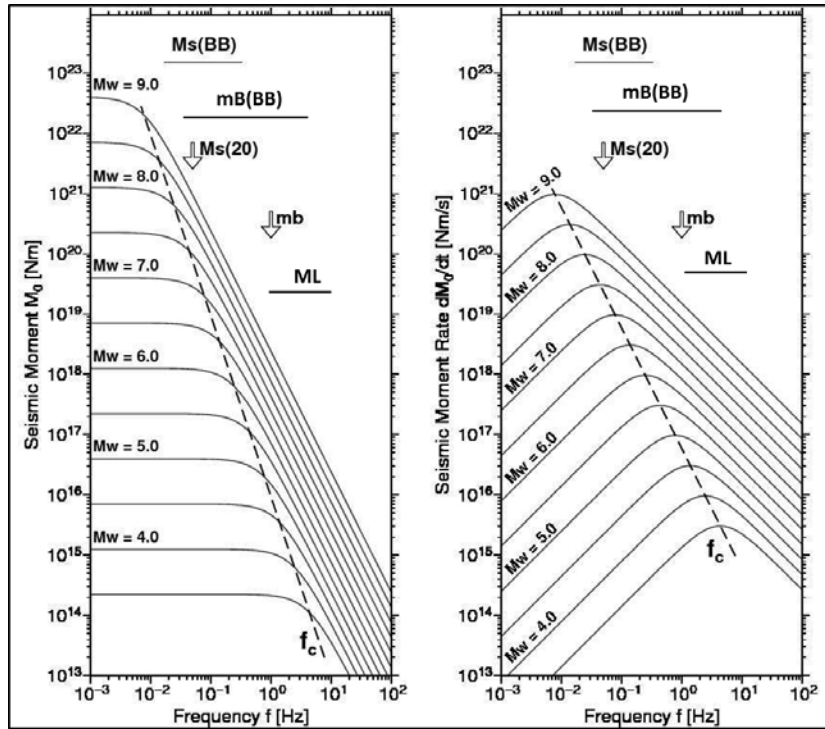
$$M_0 = 4\pi d \rho v_{p,s}^3 u_0 / R_{\theta,\phi}^{p,s} \quad (3.2)$$

with:  $d$  - hypocentral distance between the event and the seismic station;  $\rho$  - average density of the rock and  $v_{p,s}$  - velocity of the P or S waves around the source;  $R_{\theta,\phi}^{p,s}$  - a factor correcting the observed seismic amplitudes for the influence of the radiation pattern of the seismic source, which is different for P and S waves (see Figs. 3.3 and 3.100-3.103),  $u_0$  - the low-frequency level of displacement amplitudes as derived from the seismic spectrum of P or S waves, corrected for the instrument response, wave attenuation and surface amplification. For details see EX 3.4.

According to Aki (1967) a simple seismic shear source with linear rupture propagation shows in the far-field smooth displacement and velocity spectra. When corrected for the effects of geometrical spreading and attenuation we get "source spectra" similar to the generalized ones shown in Fig. 3.5. There the displacement spectral amplitudes have been scaled to the scalar seismic moment  $M_0$  (left) and the related velocity amplitudes to moment rate  $dM_0/dt$  (right), respectively. Additionally shown are the bandwidth ranges for measuring the IASPEI standard body wave magnitudes  $mb$  and  $mb(BB)$  as well as the standard surface-wave magnitudes  $Ms(20)$  and  $Ms(BB)$  (see IS 3.3). The spectral curves (solid lines) have been calculated on the basis of an  $\omega^{-2}$  source model (Aki, 1967) in which the moment rate function can be expressed as  $\hat{M}(f) = (M_0 f_c^2) / (f^2 + f_c^2)$  (Houston and Kanamori, 1986; Polet and Kanamori, 2000) with the corner frequency  $f_c = c\beta(\Delta\sigma/M_0)^{1/3}$  (Brune, 1970, 1971),  $\beta$  the shear-wave velocity near the source (assumed to be 3.75 km/s),  $c = 0.49$  and  $\Delta\sigma$  the stress drop in the source.  $\Delta\sigma$  has been assumed to be 3 MPa in agreement with Kanamori's (1977) condition  $E_s/M_0 = 5 \times 10^{-5}$ , on which the derivation of the  $M_w$ - $\log M_0$  relationship (Hanks and Kanamori, 1979) is based. According to this model the amplitudes decay is  $\sim f^{-2}$  for  $f \gg f_c$ . Although this "omega-squared" model is widely accepted as the best simple model of source spectra it is recognized that it does not adequately describe some source spectra. In real data, many kinds of departure from the  $f^{-2}$  decay law have been observed, with slopes ranging between about -1 and -5. Further, for earthquakes of a given seismic moment,  $\Delta\sigma$  may differ by about 3 orders of magnitude (e.g., Choy and Boatwright, 1995) and thus  $f_c$  according to the Brune relation by about a factor of 10. Moreover,  $f_c$  in P-wave and S-wave source spectra may differ. How much depends on the type of rupture model. If the corner frequency were determined entirely by source duration, then one would expect it to be independent on the wave type. According to Madariaga (1976), however, the corner-frequency is determined by the stopping phase from the sides of the fault nearest to and farthest from the observer. If this is true then the corner frequencies depend on the wave-type, but also vary with the position of the observer with respect to the source rupture and its radiation pattern. According to Table 2 and Eq. (3) in EX 3.4 it holds on average that  $f_c$  for P-waves is about 1.4 to 1.5 times larger than  $f_c$  for S-waves in the case of circular ruptures assumed by the Brune, respectively Madariaga models.

Before discussing the inferences one can draw from *seismic scaling laws*, respectively *source spectra* corrected for propagation effects (geometric spreading and attenuation), such as the ones in Fig. 3.5, we should make clear that details of theoretical "source spectra" depend on the model assumptions of the rupture process. E.g., when the rupture is bilateral, as in the Aki (1967)  $\omega^{-2}$  source model, the displacement spectrum of the source-time function drops for  $f \gg f_c$  proportional to  $f^{-2}$ , as in Fig. 3.5. In contrast, for an unilateral rupture (Haskell, 1964, 1966), the high-frequency decay is proportional to  $f^{-3}$  ( $\omega^{-3}$  model). The high-frequency drop-

off of real spectra typically ranges between about -1 and -3 (e.g., Hartzel and Heaton, 1985; Boatwright and Choy, 1989; Polet and Kanamori, 2000; IS 3.4), depending also on the specific source-time function. But even steeper decays have been observed. On the other hand, when the linear dimensions of the fault rupture differ in length and width then two corner frequencies will occur. Whether the two or three corner frequencies are resolvable will depend on their separation along the spectral axis and the spectral signal-to-noise ratio. In contrast to the smooth theoretical spectral curves in Fig. 3.5 real spectra calculated from noisy records of limited duration and bandwidth from earthquakes with different source-time functions and fault geometries will show fluctuating spectral amplitudes, sometime more than one corner frequency or – more likely - a rather broad range of transition with different slopes from the displacement plateau to the final drop-off. Therefore, real source spectra may not be well matched by such a smooth “average”  $\omega^{-2}$  source model and not allow to discriminate between different types of rupture propagation and source geometry. Nevertheless, Fig. 3.5 allows us to discuss essential differences and required bandwidth ranges for reasonably reliable measurements of  $M_0$  and  $E_S$ , of the related magnitudes  $M_w$  and  $M_e$  or complementary ones, as well as of the derivation of other source parameters of interest.



**Fig. 3.5 Left:** "Source spectra" of ground displacement amplitudes  $A$  as a function of frequency  $f$  for "average" seismic shear sources assuming the  $\omega^{-2}$  source model and constant stress drop of  $\Delta\sigma = 3$  MPa, scaled to seismic moment  $M_0$  and the equivalent moment magnitude  $M_w$ . **Right:** The same as left, but for ground motion velocity amplitudes  $V$ , scaled to seismic moment rate and  $M_w$ . Note that the maximum of seismic energy  $E_S \sim V^2$  is radiated around  $f_c$ . The open arrows point to the center frequencies on the abscissa at which the 1 Hz body-wave magnitude  $m_b$  and the 20 sec surface-wave magnitude  $M_s(20)$ , respectively, are determined. The horizontal interval bars mark the range of frequencies within which the maximum Sg/Lg-, P-wave and Rayleigh-wave amplitudes for ML, mB(BB) and Ms(BB) should be measured. For procedure of calculation see text (courtesy of Domenico

Di Giacomo, 2008). Modified version of Figure 1 in Bormann et al. (2009, BSSA, Vol. 99, No. 3, p. 1870); © granted by the Seismological Society of America.

With reference to Fig. 3.5, we ask and answer now the following questions:

- What does the shape of the spectra in Fig. 3.5 tell us about the source process?
- Which parameters should accordingly be measured?
- How do they relate to the size, shape, slip and stress drop in the seismic source?
- How do these parameter influence the measurement of magnitudes in different spectral ranges?

The following general features are obvious from Fig. 3.5:

- "Source spectra" are characterized by a "plateau" of constant displacement for frequencies smaller than the "corner frequency"  $f_c$  which is inversely proportional to the source dimension, i.e.,  $f_c \sim 1/L$ .
- The decay of spectral displacement amplitude beyond  $f > f_c$  is on average approximately proportional to  $f^{-2}$ .
- The plateau amplitude increases with seismic moment  $M_0$  and magnitude, while at the same time  $f_c$  decreases in the case of the validity of the  $\omega^{-2}$  model proportional to  $M_0^{-3}$  (see Aki, 1967).
- Magnitude estimates of earthquake size are closely related to seismic moment  $M_0$ , if the displacement amplitudes are sampled at frequencies smaller than the corner frequency  $f_c$  of the source spectrum, i.e., on or near to the low-frequency asymptote plateau of the source spectrum. This can be expected for  $m_b$  only at moment magnitudes  $M_w < 5.5$  and for  $M_s(20)$  at  $M_w < 8$ . For larger earthquakes, however, these magnitudes are likely to have numerical values less than  $M_w$ ,  $m_b$  much more than  $M_s(20)$ . In contrast,  $mB(BB)$  and  $M_s(BB)$  are measured in a wider range of periods resulting in reduced differences to  $M_w$  in certain magnitude ranges. These general rough estimates are supported by empirical data of the average differences between  $m_b$ ,  $mB$ ,  $M_s$  and  $M_w$  (see section 3.2.5).
- Since wave energy is proportional to the square of ground motion particle velocity, i.e.,  $E_s \sim (2\pi f u)^2 = (\omega u(\omega))^2$ , its maximum occurs at  $f_c$ . Therefore, for magnitudes to be a good measure of the amount of released seismic energy and thus of the earthquake strength in terms of ground-shaking and damage potential,  $V_{max}$  and the related period should be measured at or near to  $f_c$ .
- This is in agreement with the  $\log(A/T)_{max}$  input into the classical teleseismic magnitude formulas (see DS 3.1 and sections 3.2.5.11 and 3.2.5.2), provided that no a priori limits for the period range are imposed. The latter, however, is the case for  $m_b$  and  $M_s(20)$ , which sample  $V_{max}$  only in very limited ranges of frequency, not, however,  $mB(BB)$  and  $M_s(BB)$ , which cover the average maxima of the released velocity spectra reasonably well in the magnitude ranges 4 to 8 and 5.5 to 8.5, respectively.
- Good estimates of energy magnitude  $M_e$  require to integrate squared velocity source spectra sufficiently wide towards both sides of the spectral velocity peak (see Fig. 3.5, section 3.2.7.2 and Bormann and Di Giacomo, 2011).

- measuring displacement amplitudes at  $f > f_c$  or velocity amplitudes at  $f < \text{or} > f_c$  results in systematic underestimation of either moment or energy related magnitudes, the more the larger the frequencies differ from  $f_c$ .

**Note:** There have been objections against a preference for broadband magnitudes from those in favour of continuing to measure as classical teleseismic magnitudes only mb and Ms(20). They claim that one should be able, by **assuming** the same omega-squared model and no bias in the magnitude measurement, to estimate energy comparably well also from mb or Ms(20). But, in contrast to estimates based on direct broadband  $V_{\max}$  measurements, such estimates are even more model dependent. Measuring  $V_{\max}$  directly eliminates the influence of measurement errors in period, which is required for mb and Ms(20) calculations. Such errors may become rather large if displacement amplitudes are measured on the spectral slope which may for individual earthquakes significantly differ from the assumed omega-squared model. Moreover, also the ground displacement amplitudes calculated via measurements on narrowband filtered records are particularly sensitive to measurement errors if the period of the recorded  $A_{\max}$  falls into the period range of the steep slopes of the record transfer functions (see Fig. 3.16). Furthermore, nobody can tell for sure, without additional measurements at other record positions whether  $(A_{\max}/T)$  really corresponds to  $(A/T)_{\max}$  on the given record and how the latter relates to the true  $V_{\max}$  at a rather different period of the broadband ground motion spectrum at the given site. Therefore, only by hypothesis free direct measurement of  $V_{\max}$  one can expect to get the least biased (although also not error free) estimate of the actually released seismic energy.

The effect of underestimating the magnitude and thus the (dimensional) *size* or (potential shaking) *strength* of the earthquake is often misleadingly termed *magnitude saturation*. More correct would be to speak of a systematic underestimation of a magnitude with respect to a true non-saturating moment or energy related reference magnitude which grows linearly and everywhere with the same slope proportional with  $\log M_0$  or  $\log E_s$ , respectively. The underestimation of a non-reference magnitude is then due to the slower increase of its related measurement amplitudes with growing Mw or Me when they are measured at frequencies outside the displacement plateau or the velocity maximum, respectively. In fact, any conventional magnitude that is measured at a constant period or in a narrow period range, such as Ml, mb, mb(Lg) or Ms\_20, features “saturation” as soon as the measured periods fall below the corner period of the radiated source spectrum or are much smaller than the rupture duration. Yet, for short, we continue to term this effect in the following as the *spectral component of magnitude “saturation”*.

The general shape of the displacement source spectrum in Fig. 3.5 can be understood as follows:

We know from optics that under a microscope no objects can be resolved which are smaller than the wavelength  $\lambda$  of the light with which it is observed. In this case the objects appear as a blurred point or dot. In order to resolve more details, electron microscopes are used which operate with much smaller wavelengths. The same holds in seismology. When observing a seismic source of radius  $r$  with wavelengths  $\lambda \gg r$  at a great distance, one can not derive any information about the details of the source process but only of its overall (integral) features, i.e., one “sees” a *point source*. Accordingly, spectral amplitudes with wavelengths  $\lambda \gg r$  are constant and form a spectral plateau (if source duration is neglected). On the other hand, wavelengths that have  $\lambda \ll r$  can resolve internal details of the rupture process. In the case of

an earthquake they correspond to smaller and smaller elements of the rupture processes or of the fault roughness (*asperities* and *barriers*). Therefore, their spectral amplitude contributions decay rapidly with size and thus related higher frequencies.

Accordingly, the corner frequency,  $f_c$ , marks a critical position in the spectrum which is obviously related to the size of the source. According to Brune (1970; 1971) and Madariaga (1976), both of whom modeled a circular fault, the corner frequency in the P- or S-wave spectrum, respectively, is  $f_{c\ p/s} = c_m v_{p,s} / \pi r$ . In contrast, assuming a rectangular fault, Haskell (1964) gives the relationship  $f_{c\ p/s} = c_m v_{p,s} / (L \times W)^{1/2}$  with  $L$  the length and  $W$  the width of the fault. The values  $c_m$  are model-dependent constants and thus different for circular and rectangular faults but even for different models of circular faults [see Table 2 and Eq. (3) in EX 3.4]. Accordingly, one may roughly define also a critical wavelength  $\lambda_c \approx v/f_c = c_m \pi r$  or  $\lambda_c \approx c_m (L \times W)^{1/2}$  beyond which the source can be “seen” as a *point source* only.

But real fault ruptures may have any other shape which can only be approximated roughly by either a circular fault (more likely for small earthquakes within the brittle fracturing seismogenic zone of the Earth crust) or by a rectangular fault of different *aspect ratio*  $L/W$ , which is growing with the magnitude of great earthquakes that rupture the whole width (depth range) of the seismogenic zone and may, therefore, grow further only in  $L$  direction. This has consequences for seismic scaling laws for small and large earthquakes [see related discussions and controversies, e.g., Scholz (1982, 1994, 1997) and Wang and Ou (1998) with further related references].

Nevertheless, being aware of the model dependence and inherent uncertainties of seismic scaling laws such as the ones in Fig. 3.5, related rupture models and formulas as given, e.g. by Brune (1970, 1971) and Madariaga (1976), allow to roughly estimate the following source parameters by reading the following parameters from seismic source spectra that have been corrected for propagation effects (geometric spreading and attenuation):

- $M_0$  via Eq. (3.2) when measuring the displacement plateau amplitude  $u^0$  and accounting for the other parameters in that formula;
- Source radius  $R \sim 1 / 2\pi f_{cp/s}$ ;
- Area of the assumed circular rupture plane:  $A = \pi R^2$ ;
- Average dislocation  $\bar{D} = M_0 / (\mu A)$  by assuming a reasonable value for rigidity  $\mu = v_s^2 \rho$  in the source volume;
- Stress drop  $\Delta\sigma = (7/16)M_0 / R^3$  according to Eshelby (1957) and Keilis-Borok (1959).

Stress drop means the difference in acting stress at the source region before and after the earthquake. For more details see Figure 10 in IS 3.1. Note, however, that  $\Delta\sigma \sim R^{-3}$  and estimates of  $R$  not very reliable. Reasons are the relatively large reading errors in  $f_c$  plus the model-dependence of  $R$  (or other geometric parameter) estimates, which also depend on the assumed *rupture velocity*  $v_R$ . Therefore, stress-drop estimates maybe rather uncertain within about an order of magnitude. For more details see EX 3.4 with discussions.

#### 3.1.2.4 The influence of rupture velocity and duration on magnitude estimates

For reducing saturation, *rupture duration*, depending on rupture length and speed, has additionally to be taken into account, especially, when the largest amplitudes of non-dispersive of P-wave trains have to be measured. The average rupture duration  $T_{\text{Rav}}$  increases according to Bormann and Saul (2009a) with magnitude  $M$  according to

$$\log T_{\text{Rav}} \text{ (in s)} \approx 0.6 M - 2.8. \quad (3.3)$$

$M$  means in this relationship the largest measured magnitude, which is for  $M_w < 6$  usually the local magnitude  $M_l$  or  $m_b$  and for larger events  $M_s$  or  $M_w$ . Relationship (3.3) has been derived by using data published in Olson and Allen (2005) that are based on rupture velocity estimates in Somerville et al. (1999) and scaling relations between rupture length and magnitude published by Wells and Coppersmith (1994), complementing them by some direct rupture duration estimates of recent great earthquakes. This simple average relationship is easy to recall and yields with 0.5 magnitude units increase a doubling of  $T_{\text{Rav}}$ , e.g., for  $M = 6$  about 6 s, for  $M = 7$  about 25 s, for  $M = 8$  about 100 s, for  $M = 9$  about 400 s, and for  $M = 10$  about 1600 s. The figure for  $M = 9$  is close to respective estimates for the somewhat greater  $M_w 9.3$  Sumatra-Andaman earthquake of December 26, 2004 by Bormann and Wylegalla (2005), Krüger and Ohrnberger (2005), Gusev et al. (2007), Hara (2007), Bormann and Saul (2009b), and Lomax and Michelini (2009a and b, 2011). All these authors, with the exception of Gusev et al., describe procedures of non-saturating magnitude estimates in near real-time based on broadband P-wave data and take explicitly rupture duration into account. The  $M = 10$  figure of 1600 s agrees well with rupture durations given in Cifuentes and Silver (1989) for the “revised”  $M_w \approx 9.9$  of the great 1960 Chile earthquake (see discussion in 3.1.1.1).

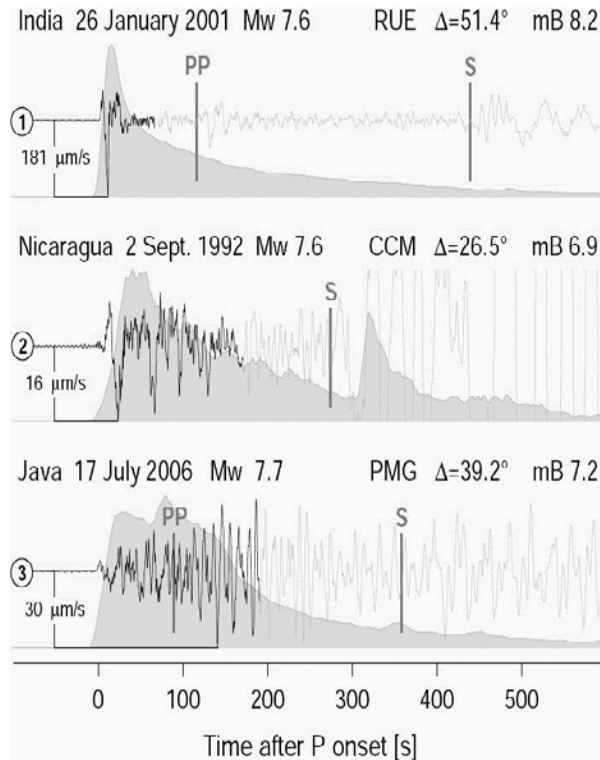
Average estimates of  $T_{\text{Rav}}$  according to relationship (3.3) agree within a factor less than two with those that could be estimated from the corner frequencies  $f_c$  of the source spectra in Fig. 3.5 as a function of  $M_w$  when one assumes an average rupture velocity of about 2.5 km/s and that  $f_c$  is inversely proportional to  $T_{\text{Rav}}$ . For a given event magnitude, however, depending mainly on variations in stress drop (e.g., Choy and Boatwright, 1995) and rigidity in the source region (for *tsunami earthquakes* see., e.g., Houston, 1999; Polet and Kanamori, 2000), but other factors as well (see 3.1.2.5), true  $T_R$  may be a factor of 2 to 3 smaller or larger than  $T_{\text{Rav}}$  because of related differences in rupture velocity,  $v_R$ .  $v_R$  may range between about 0.2 and 0.9 times of the shear-wave velocity  $v_s$  (Venkataraman and Kanamori, 2004b), or even more than  $v_s$  if the hypothesis and still debated observations of supershear slip pulses with rupture velocities up to 5 km/s holds (e.g., Bouchon and Vallée, 2003). Fig. 3.6 shows an example of very large differences in rupture duration and thus duration of the P-wave trains radiated from events with equal  $M_w$ .

If the largest amplitude for magnitude determination is measured only within an a priori fixed limited measurement-time windows after the first body-wave onset that is shorter than the rupture duration, then the required **maximum** amplitude in the whole P-wave train may have been missed and thus  $m_b$  or  $m_B$  been underestimated. Since  $T_{\text{Rav}} > 6$  s for earthquakes with  $M > 6$ , it is likely that this is indeed the case for such earthquakes when  $A_{\text{max}}$  is measured within a time window  $< 6$  s, as it is common practice at the International Data Center (IDC) of the CTBTO (Comprehensive Test Ban Treaty Organization; see Chapter 15). We term this the *measurement time-window component* of saturation, which adds to the pronounced spectral saturation component of  $m_b$ . In combination, both component may indeed result in true saturation, as shown in Fig. 3.7 for  $m_b(\text{test})$ . In contrast, the new IASPEI

standard mb, which measures always the maximum amplitude within the whole P-wave train, including the range of depth phases, and ending (preferably) before the onset of PP is only reduced by the spectral component of saturation. Yet, standard mb keeps growing up to values of about 7.6 for the largest earthquakes with Mw around 9.

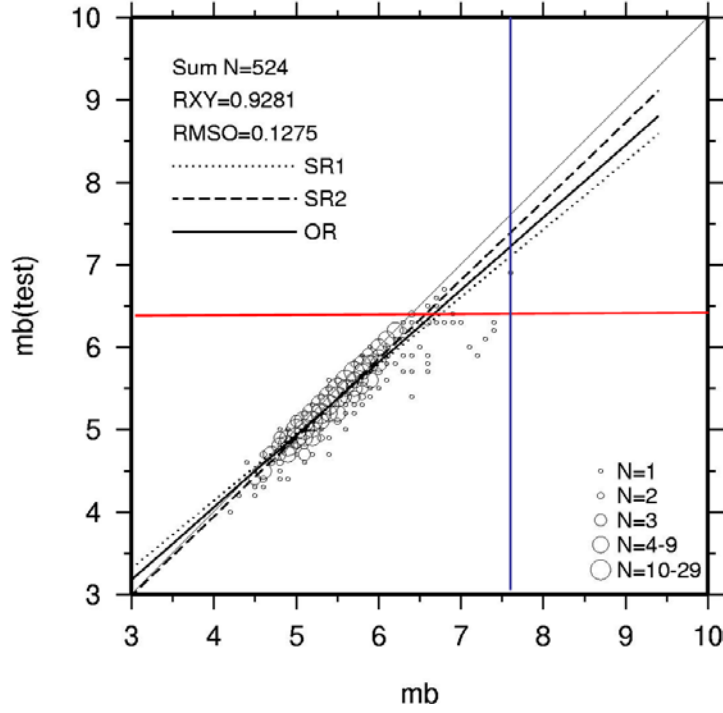
Compared with an earthquake of the same seismic moment or magnitude, the corner frequency  $f_c$  of a well contained underground nuclear explosion (UNE) in hard rock is about ten times larger. Accordingly, an UNE produces more high-frequent energy and thus has a larger  $E_s$  as compared with an earthquake of comparable mb. Accordingly, mb(IDC) does not underestimate mb even for the largest UNE but only for earthquakes with comparable seismic moment but much smaller corner frequency. Main cause of this difference in  $E_s$  and high-frequency content between UNE and earthquakes is that the source *duration* and thus the *rise time*,  $t_r$ , to the final level of static displacement is much shorter for explosions (see Fig. 3.4).

Moreover, the shock-wave front of an explosion, which causes the deformation and fracturing of the surrounding rocks and thus the generation of seismic waves, propagates with approximately the P-wave velocity  $v_p$  while the velocity of crack propagation along a shear fracture/fault is only about 0.5 to 0.9 of the S-wave velocity, i.e., about 0.3 to 0.5 times that of  $v_p$ . Additionally, the equivalent wave radiating surface area in the case of an explosion is a sphere with  $A = 4\pi r^2$  and not a plane with  $A = \pi r^2$ . Accordingly, for equal  $A$  the source radius  $R$  in the case of an explosion is smaller and thus the related corner frequency larger.



**Fig. 3.6** Velocity broadband records of earthquakes with practically identical Mw but very different rupture durations and therefore length of the P-wave trains. Record 2 and 3 are slow tsunami earthquakes. Grey shaded areas: Envelopes of high-frequency (1-3 Hz) filtered BB records. Estimated rupture durations range from about 25 s (record 1) to 190 s (record 3). The average TRav according to Eq. (3.3) for Mw = 7.7 would be 66 s. Figure copied from Figure 2 of Bormann and Saul Bormann et al. (2009, p. 700) with © granted by the Seismological Society of America.





**Fig. 3.7** Comparison of mb measured on WWSSN-SP filtered records a) within the first 6 s after the P-wave onset (mb test) and b) within the whole P-wave train according to the new IASPEI standard rules for mb measurement. For instrument response function and measurement rules see IS 3.3. While mb(test) saturates at about 6.4 (**red line**), with early underestimations beginning already at mb > 6, mb itself may grow up to about 7.6 for the greatest earthquakes (**blue line**). RXY = correlation coefficient, RMSO = orthogonal rms error. Figure compiled according to data in Bormann et al. (2009).

### 3.1.2.5 Factors which influence energy radiation and seismic efficiency

The ratio between radiated seismic energy and seismic moment,  $E_S/M_0$ , i.e., the seismic energy radiated per unit of moment release and termed “scaled energy” by Kanamori and Brodsky (2004), would remain constant if a seismic scaling law such as the one in Fig. 3.5, calculated for a specific rupture model with fixed stress drop and rupture velocity, would hold for all earthquakes. In fact, this was assumed by Kanamori (1977) and Hanks and Kanamori (1979) when they developed the moment magnitude scale  $M_w$  under the condition of

$$E_S/M_0 = 5 \times 10^{-5} = \text{constant}. \quad (3.4)$$

This, however, is not the case. According to Wyss and Brune (1968) holds

$$E_S/M_0 = \Delta\sigma/2\mu = \tau_a/\mu \quad (3.5)$$

with  $\tau_a$  = *apparent stress*, which can – with Eq. (3.1) - also be written as  $\tau_a = E_S/(D \cdot A)$ . Thus,  $\tau_a$  is equivalent to the seismic energy release per unit area of the ruptured surface  $A$  and per unit of slip. Since both  $\Delta\sigma$  and  $\mu$  may vary  $E_S/M_0$  is a useful parameter to characterize the dynamic properties of an earthquake (Aki, 1966; Wyss and Brune, 1968).

$\Delta\sigma$  and thus  $\tau_a$  may vary by about 3 to 4 orders of magnitude (e.g., Abercrombie, 1995; Choy and Boatwright, 1995). Accordingly, the difference between energy magnitude  $M_e$  and moment magnitude  $M_w$  may reach 1 m.u. and even more (e.g., Choy and Kirby, 2004; Di Giacomo et al., 2010a and b; Bormann and Di Giacomo, 2011; Figure 5 in IS 3.5).

In the case of tsunami earthquakes unusually low rigidity values  $\mu$  in the rupturing source volume, i.e., along near surface faults in shallow dipping subduction zones with water saturated ocean sediments (Houston, 1999; Polet and Kanamori, 2000 and 2009) may be the reason for their exceptionally slow ( $v_R \approx 0.2 v_s$ ) and unusually long lasting ruptures. The ratio  $E_S/M_0$  of *tsunami earthquakes* ranges between about  $2 \times 10^{-6}$  and almost  $10^{-7}$  and the related values of  $\Theta = \log E_S/M_0$  - termed “slowness parameter” by Okal and Talandier (1989) - between -5.7 and -7 (Lomax and Michelini, 2009a). Note that observed global average values of  $\Theta$  for earthquakes with magnitudes larger than 5.5 range between -4.7 and -4.9 (Choy and Boatwright, 1995; Choy et al., 2006; Weinstein and Okal, 2005), whereas the Kanamori (1977) condition in Eq. (3.4) corresponds to  $\Theta_K = -4.3$ .

According to Kanamori (1972) tsunami earthquakes are characterized by generating a much larger tsunami than it would be expected from their seismic moment. The reason is that according to Eq. (3.1) for a measured  $M_0$  the product  $\bar{D} \times A$  is strongly underestimated when instead of the much smaller real rigidity in the rupture area the much larger Kanamori (1977) value for average crust-upper mantle condition is assumed (Hirshorn and Weinstein, 2009). The latter, however, is routine in  $M_0$  calculations.

In this context we have to clarify that Kanamori (1977), when introducing the relationship

$$W = W_0 (= E_S) = (\Delta\sigma/2\mu) M_0 \quad (3.5)$$

based it on a simplified assumptions of the rupture process which did not account for the total energy balance. According to Kanamori and Brodsky (2004) the total change in potential energy  $W$  due to the rupture process has to be written as

$$\Delta W = E_S + E_G + E_H \quad (3.6)$$

with  $E_G$  = fracture energy and  $E_H$  = energy dissipated as heat due to kinematic friction on the fault and generated sound. Eq. (3.5), however, holds only when both  $E_G = E_H = 0$ , i.e., when neglecting both the fracture energy and the dissipated energy in the total energy budget. Yet, the radiated energy is less, at least (in the case with no dissipation)

$$E_S = (\Delta\sigma/2\mu) M_0 - E_G. \quad (3.7)$$

Husseini (1977) introduced the term of *radiation efficiency*, which for radiated seismic energy would read:

$$\eta_R = E_S/(E_S + E_G) \quad (3.8)$$

According to Kanamori (2006) this can also be written as

$$\eta_R = (2\mu/\Delta\sigma) (E_S/M_0). \quad (3.9)$$

Yet, then it becomes clear that  $E_S = (\Delta\sigma/2\mu) M_0$  holds only for  $\eta_R = 1$ , i.e., when all energy available for strain release is converted into seismic wave energy, and no energy  $E_G$  is dissipated during fracturing. This, however, is not the case.  $\eta_R$  ranges between about 0.02 and (by definition)  $<1$ . For some earthquakes, however,  $\eta_{R\eta} > 1$  were computed, either due to incorrect data correction, model inaccuracies or even real supershear slip-pulses (Bouchon and Vallée, 2003; Venkataraman and Kanamori, 2004b).

Moreover, it has to be made clear that radiation efficiency according to Eq. (3.8) should not be mistaken as the better known seismic efficiency, which additionally accounts for energy dissipation losses due to frictional heat and sound,  $E_H$ , i.e.,

$$\eta = E_S / (E_S + E_G + E_H). \quad (3.10a)$$

Thus, seismic efficiency, defined as the proportion of the total available strain energy which is radiated as seismic waves, may be significantly smaller than  $\eta_R$  the more so the larger  $E_H$ , and  $\eta_R$  could also be called the maximum seismic efficiency of ruptures with negligible dissipation. According to Kanamori (2001) slow ruptures are suggestive of a dissipative process. He gives for longitudinal shear (mode III) cracks the following relationship between  $\eta$ , rupture velocity  $v_R$  and shear-wave velocity  $v_s$ :

$$\eta = 1 - [(v_s - v_R)/(v_s + v_R)]^{1/2}. \quad (3.10b)$$

According to (3.10b) a rupture with  $v_R = 0.2 v_s$  would thus have a value of  $\eta = 0.18$  only. Other crack geometries, however, may yield other values.

In the above discussion we have focussed on the influence of variations in stress drop, respectively of  $\Delta\sigma/\mu$  on the ratio  $E_S/M_0$ , although we mentioned already related velocity variations. According to Newman and Okal (1998), even in the case of a kinematically simple rupture model this ratio is controlled by five dimensionless factors, namely the ratios (with different exponent) between (1) S- versus P-wave velocity to the fifth power, (2) rupture length versus rupture duration and Rayleigh-wave speed to the third power, (3) rupture velocity  $v_R$  versus shear-wave velocity  $v_s$  to the third power, (4) the aspect ratio rupture width versus rupture length to the second power, and (5) fault slip versus rupture width. The first factor is, for most rise and fall times of the source time function and for a Poissonian solid a constant. The second factor, however, depends on the *directivity* of the rupture which will be 1 for unilateral rupture but may increase up to 8 for a symmetric bilateral rupture. However, Venkataraman and Kanamori (2004a) showed for a suite of earthquakes that the directivity corrections were less than a factor three and for dip-slip earthquakes with rupture propagation along strike even less than two at teleseismic distances. The third factor expresses the slowness of rupture, which may vary between about 0.5 for a “regular” ratio  $v_R/v_s \approx 0.8$  and about 0.01 and 0.03 for very slow earthquake ruptures with  $v_R/v_s \approx 0.2-0.3$  and  $v_R \approx 1$  km/s, slowed down by subducted sedimentary material along the fault zone (Kikuchi and Kanamori, 1995). The fourth factor, the squared aspect ratio, might be around  $1/4$  for most rupture geometries, but could be significantly smaller for ribbon-like ruptures on strike-slip faults and in the case of great earthquakes which brake the whole brittle-fracturing seismogenic zone of the lithosphere and can then grow further only in strike direction. And the last factor depends on the geometry model of the source. Thus we see that a great many of factors, most of them not directly measurable, may influence the “scaled energy” or slowness parameter  $\Theta$  of individual earthquakes.

In an effort to reduce this complexity down to the most important factors one finds that according to Kostrov (1966) and Eshelby (1969) the radiation efficiency of transverse shear cracks is

$$\eta_R \approx (v_R/v_s)^2. \quad (3.11)$$

When inserting (3.11) into (3.9) and resolves it for the ratio  $E_S/M_0$  one gets according to Bormann and Di Giacomo (2011):

$$E_S/M_0 = \eta_R \cdot (\Delta\sigma/2\mu) \approx (v_R/v_s)^2 \cdot (\Delta\sigma/2\mu). \quad (3.12)$$

Although being model-dependent (and other models are likely to produce other relationships), Eq. (3.12) illustrates a trade-off between variations in stress drop, rigidity, rupture velocity and shear-wave velocity in controlling the ratio  $E_S/M_0$  and thus, the relationship between  $M_w$  and  $M_e$ . Since, however, stress-drop estimates, despite their principle uncertainty, show the largest variability with 3 to 4 orders of magnitude, most of the observed differences in scaled seismic energy released by individual earthquakes may be attributable to differences in stress drop, the more so, since  $\Delta\sigma/2\mu$  in the source volume seem to set the stage for  $v_R/v_s$ , i.e., the two terms in (3.12) are not fully independent.

Finally, one should note in this context that a thorough investigation by Oth et al. (2010) of 1,826 earthquakes with  $M(JMA)$  magnitudes (see Eq. 3.34) between 2.7 and 8 throughout Japan led to the conclusions that:

- the calculated source spectra can be well characterized by the omega-square model (i.e., as in Fig. 3.5);
- on average self-similar scaling over the entire magnitude range holds, with median stress drops of 1.1 and 9.2 MPa for crustal and subcrustal events, respectively;
- the **ratio**  $E_S/M_0$ , as theoretically expected if the omega-squared model is valid, does not depend on magnitude and that the observed large scatter is mainly related to the scatter in stress drop.

### 3.1.2.6 Parameters which describe and control the source mechanism

Assuming that the earthquake rupture occurs along a planar fault surface the orientation of this plane in space can be described by three angles: *strike*  $\phi$  ( $0^\circ$  to  $360^\circ$  clockwise from north), *dip*  $\delta$  ( $0^\circ$  to  $90^\circ$  against the horizontal) and the direction of slip on the fault by the *rake angle*  $\lambda$  ( $-180^\circ$  to  $+180^\circ$  against the horizontal). Figures 5 and 6 in exercise EX 3.2 define these angles and Figures 3 as well as the solution figures show how to determine them from a stereographic (Wulff) net or equal area (Lambert-Schmidt) projection using observations of first motion polarities. It can be shown that a rupture along a plane perpendicular to the fault plane and with a slip vector perpendicular to the slip on the first plane causes an identical angular distribution of first motions. Therefore, on the basis of first motion analysis alone one can not decide which of the two planes is the truly acting one. For more detailed discussion on this see section 2.4 in IS 1.1.

Note that in the case of a shear model the fault-plane solution (i.e., the information about the orientation of the fault plane and of the fault slip in space) forms, together with the information about the static seismic moment  $M_0$  (see 3.1.2.3), the seismic moment tensor  $M_{ij}$

(see Equation (25) in IS 3.1). The principal axes of the moment-tensor coincide with the direction of the pressure axis, P, and the tension axis, T, associated with fault-plane solutions. However, they should not be mistaken for the principal axes  $\sigma_1$ ,  $\sigma_2$  and  $\sigma_3$  (with  $\sigma_1 > \sigma_2 > \sigma_3$ ) of the acting stress field in the Earth which is described by the stress tensor. Only in the case of a fresh crack in a homogeneous isotropic medium in a whole space with no pre-existing faults or zones of weakness and vanishing internal friction is P in the direction of  $\sigma_1$  while T has the opposite sense of  $\sigma_3$ . P and T are perpendicular to each other and each one forms, under the above conditions, an angle of  $45^\circ$  with the two possible conjugate fault planes (*45°-hypothesis*) which are in this case perpendicular to each other (see Figs. 3.99-3.101, 3.106 and 3.111 in sections 3.4.1 and 3.4.2). The orientation of P and T is also described by two angles each: the *azimuth* and the *plunge*. They can be determined by knowing the respective angles of the fault plane (see EX 3.2). If the above model assumptions hold true, one can, knowing the orientation of P and T in space, roughly estimate the orientations of  $\sigma_1$  and  $\sigma_3$ . Most of the data used for compiling the global stress map (Zoback, 1992; Zoback and Zoback, 2002) come from earthquake fault-plane solutions interpreted under these assumptions.

In reality, the internal friction of rocks is not zero. For most rocks this results, according to Anderson's (1951) theory of faulting in the formation of conjugate pairs of faults which are oriented at about  $\pm 30^\circ$  to  $\sigma_1$ . In this case, the directions of P and T, as derived from fault-plane solutions, will not coincide with the principal stress directions. Moreover, near the surface of the Earth one of the principal stresses is almost always vertical. In the case of a horizontal compressive regime, the minimum stress  $\sigma_3$  is vertical while  $\sigma_1$  is horizontal. This results, when fresh faults are formed in unbroken rock, in thrust faults dipping about  $30^\circ$  and striking parallel or anti-parallel to  $\sigma_2$ . In an extensional environment,  $\sigma_1$  is vertical and the resulting dip of fresh normal faults is about  $60^\circ$ . When both  $\sigma_1$  and  $\sigma_3$  are horizontal, vertical strike-slip faults will develop, striking with  $\pm 30^\circ$  to  $\sigma_1$ . But most earthquakes are associated with the reactivation of pre-existing faults rather than occurring on fresh faults.

Since the frictional strength of faults is generally less than that of unbroken rock, faults may be reactivated at angles between  $\sigma_1$  and fault strike that are different from  $30^\circ$ . In a pre-faulted medium this tends to prevent breaking of a new fault. Accordingly, there is no straightforward way to infer from the P and T directions determined for an individual earthquake the directions of the acting principal stress. On the other hand, it is possible to infer the regional stress based on the analysis of many earthquakes in that region since the possible suite of rupture mechanisms activated by a given stress regime is constrained. This method aims at finding an orientation for  $\sigma_1$  and  $\sigma_3$  which is consistent with as many as possible of the actually observed fault-plane solutions (e.g., Gephart and Forsyth, 1984; Reches, 1987; Rivera, 1989).

### 3.1.3 Mathematical, physical and geological representation of earthquakes

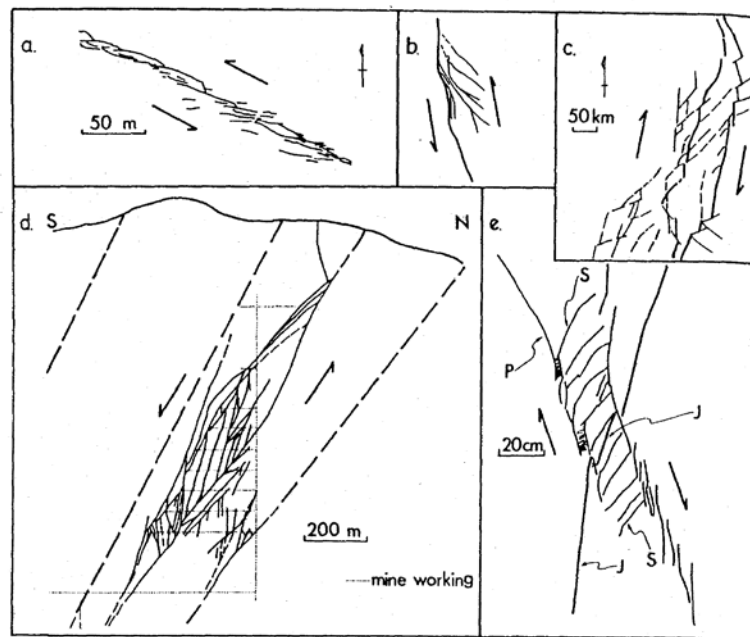
It is beyond the scope of the NMSOP to dwell on the physical and geological models of seismic sources and their mathematical representation. There exists quite a number of good text books on these issues (e.g., Aki and Richards, 1980 and 2002; Ben-Menahem and Singh, 1981 and 2000; Das and Kostrov, 1988; Scholz, 1990 and 2002; Lay and Wallace, 1995; Udías, 1999) as well as review papers on the physics and dynamics of earthquakes (e.g., Madariaga and Olsen, 2002; Teisseyre and Majewski, 2002; Kanamori and Brodsky, 2004;

Madariaga, 2011), on earthquakes as a complex system (Turcotte and Malamud, 2002) and quite many on earthquake geology and mechanics in Lee et al. (2002).

However, many of these texts are rather elaborate and more research oriented. Therefore, we have added to this Manual a more concise introduction into the theory of source representation in IS 3.1. It outlines how the basic relationships used in practical applications of source parameter determinations have been derived, on what assumptions they are based and what their limitations are. We have also added some exercises (EX 3.2-3.4) based on very simplified model assumptions and try to give with the following short section at least some idea of earthquakes as complex activated fault systems in real Earth.

### **3.1.4 Empirical analysis of rupture geometry, kinematics and dynamics in space and time**

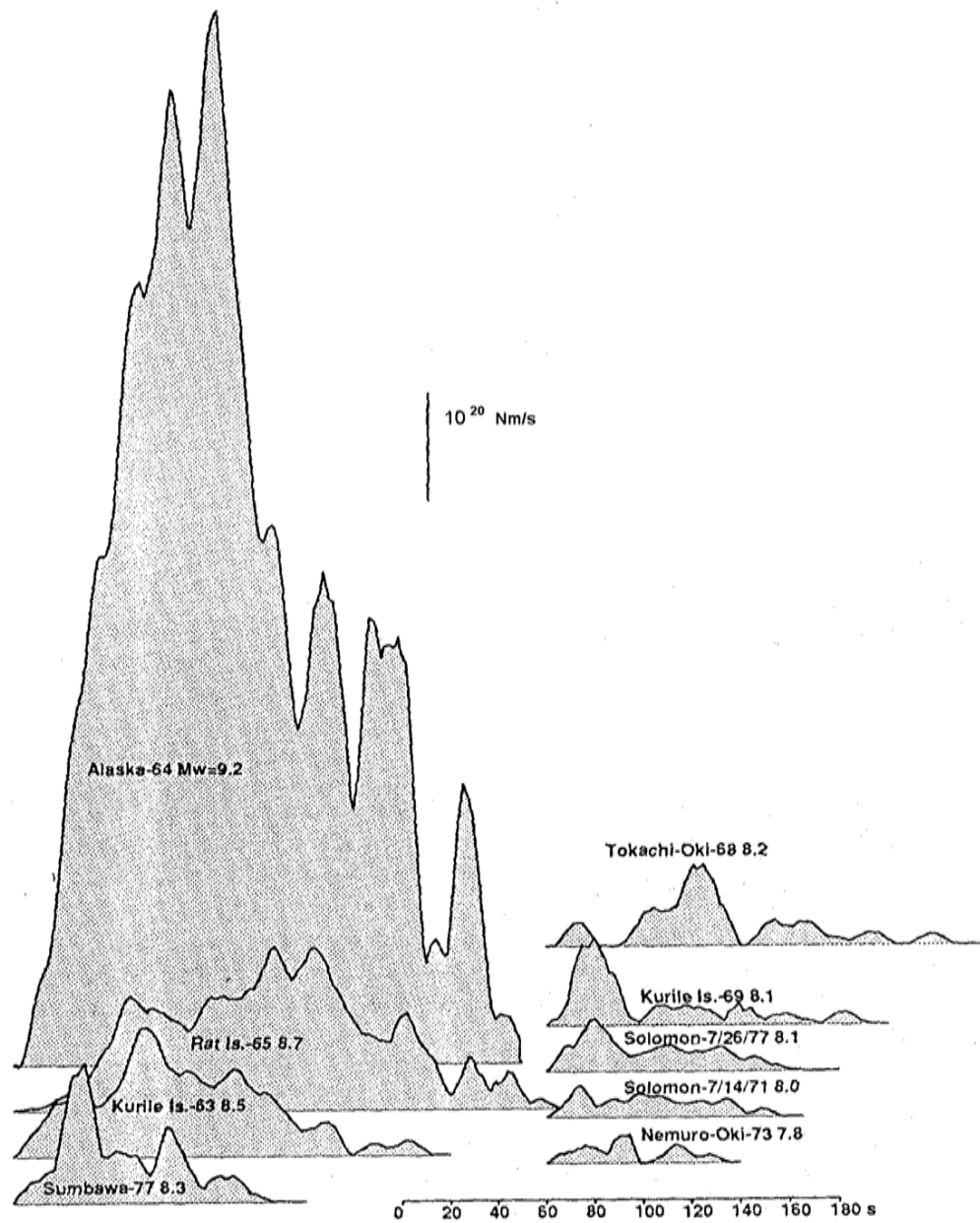
So far we have mainly considered simple earthquake models applied routinely in seismological practice to derive suitable parameters for describing the size and behavior of faulting of earthquakes and to some extent also of explosions. In reality, earthquakes do not rupture along perfect planes, nor are their rupture areas truly circular or rectangular. Rather they occur in inhomogeneous rock, rupture along mechanically heterogeneous planes, do not slip just unilaterally or bilaterally but in patterns that may be very complex both in time and space. Therefore, seismologically derived fault plane and rupture parameters are usually at best first order approximations or simplifications to the truth in order to make the problem mathematically and with limited data tractable. Real faults show jogs, steps, branching, splays, etc., both in their horizontal and vertical extent (Fig. 3.8). Such jogs and steps, depending on their severity, are impediments to smooth or ideal rupture, as are bumps or rough features along the contacting fault surfaces. More examples can be found in Scholz (1990 and 2002). Since these features exist at all scales, which implies the self-similarity of fracture and faulting processes and their fractal nature (Turcotte, 1997; Turcotte and Malamud, 2002), this will necessarily result in heterogeneous dynamic rupturing and finally also in rupture termination.



**Fig. 3.8** Several fault zones mapped at different scales and viewed approximately normal to slip (from Scholz, *The mechanics of earthquakes and faulting*, 1990, Fig. 3.6, p. 106; with permission of Cambridge University Press).

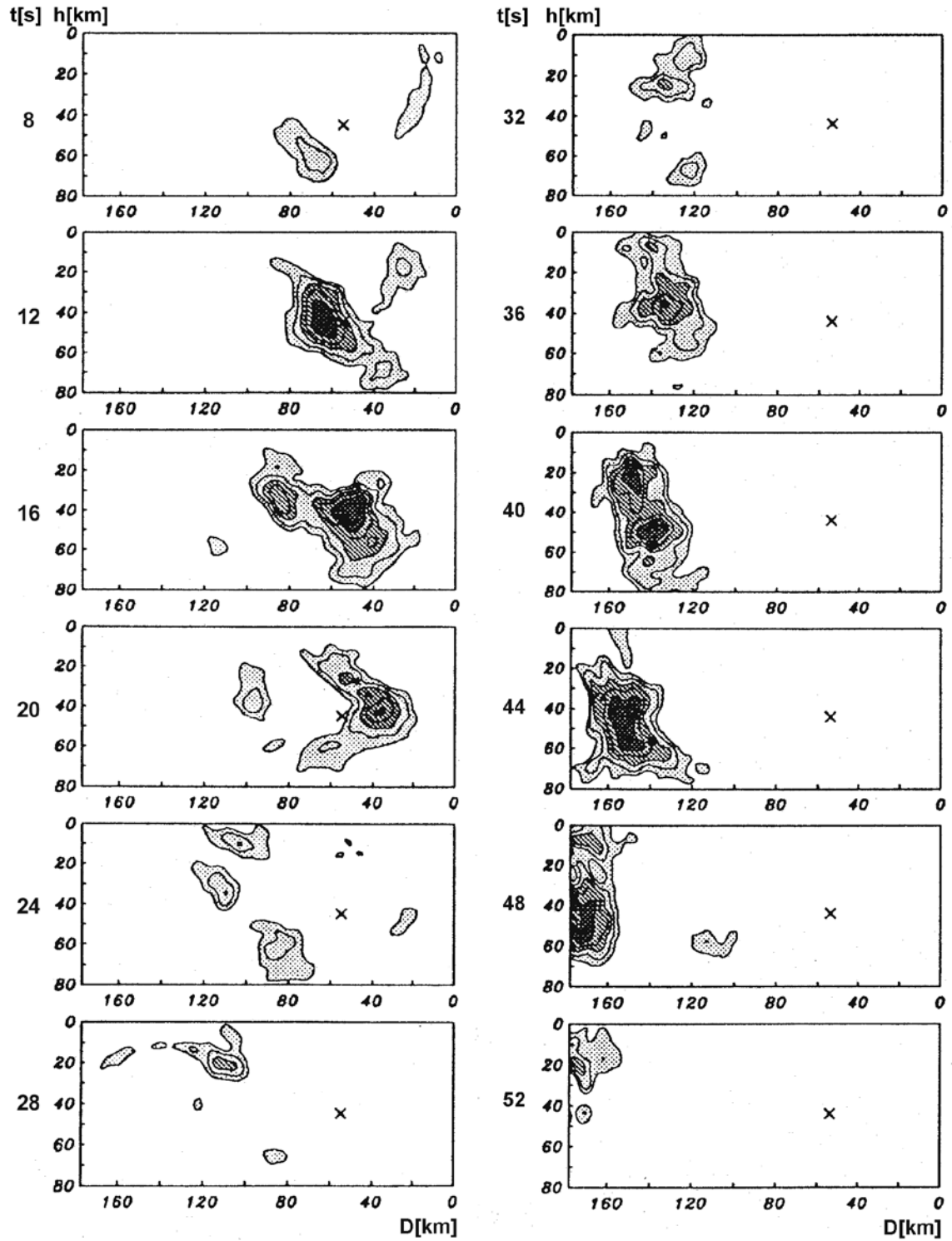
Fig. 3.9 illustrates the complexity of earthquake ruptures over time as a common feature. Each event, often occurring as a multiple rupture process, seems to have its own "moment-rate fingerprint". Only in recent decades dense strong-motion networks been fortuitously deployed in the very source region of a strong earthquake. Strong-motion data recorded on a common time-base enable a detailed analysis of the rupture history in space and time using the moment-rate density.

As an early example, Fig. 3.10 depicts an inversion of data by Mendez and Anderson (1991) for the rupture process of the 1985 Michoacán, Mexico earthquake. Shown are snapshots, 4 s apart from each other, of the dip-slip velocity field. The contours represent dip-slip velocity at 5 cm/s interval, the cross denotes the NEIC hypocenter. Three consecutively darker shadings are used to depict areas with dip-slip velocities in the range: 12 to 22, 22 to 32, and greater than 32 cm/s, respectively. Abbreviations used: t - snapshot time after the origin time of the event, h - depth, D - distance in strike direction of the fault. One recognizes two main clusters of maximum slip velocity being about 120 km and 30 s apart from each other. The related maximum cumulative displacement was more than 3 m in the first cluster and more than 4 m in the second cluster at about 55 km and 40 km distance down-dip from the top of the rupture, respectively. About 90 % of the total seismic moment was released within these two main clusters which had a rupture duration each of only 8 s while the total rupture lasted for about 56 s (Mendez and Anderson, 1991). Although the overall rupture of the 1985 Michoacán, Mexico, earthquake was unilateral, both segments of main moment release expanded also bilaterally both in strike and vertical direction.



**Fig. 3.9** Moment-rate (source time) functions for some large earthquakes in the 1960s and 1970s as obtained by Kikuchi and Fukao (1987) (modified from Fig. 9 in Kikuchi and Ishida, Source retrieval for deep local earthquakes with broadband records, *Bull. Seism. Soc. Am.*, Vol. 83, No. 6, p. 1868, 1993, © Seismological Society of America).





**Fig. 3.10** Snapshots of the development in space and time of the inferred rupture process of the 1985 Michoacán, Mexico, earthquake. See text for discussion (redrawn and modified from Mendez and Anderson, The temporal and spatial evolution of the 19 September 1985 Michoacán earthquake as inferred from near-source ground-motion records, *Bull. Seism. Soc. Am.*, Vol. 81, No. 3, Fig. 6, p. 857-858, 1991; © Seismological Society of America).

Realizing in Fig. 3.10 the very large differences in moment release over the total rupture area one really has to question the relevance for seismic hazard assessment of the average point source displacement in relationship (3.1) for the scalar seismic moment, but also of the concept of the low resolution overall “centroid” time (usually calculated at half-time of the source-time function) and of the related centroid epi- and hypocenter. Calculating the rupture velocity from the ratio of total rupture length  $L$  / total rupture duration  $T_R$ , average slip and/or stress drop would yield a misleading picture which by no means reflects the kinematic and dynamic complexity of such complex earthquakes. Gross (1996) referred to the fact that inversions of waveforms in large earthquakes have consistently shown that the spatial distribution of moment release in real earthquakes is very irregular and not at all like the smooth distribution that one would expect if the classical theory of homogeneous elastic solids according to Kanamori and Anderson (1975) would hold for all earthquakes. This led Koyama (1985) to develop a theory of incoherent fracture of random fault heterogeneities and to discuss from this perspective the physical basis of earthquake magnitudes and the relationships of  $M_s$  on the one hand and of the maximum short-period P-wave magnitude  $m_b^*$  on the other hand (Koyama and Zheng, 1985) with ruptured fault area.

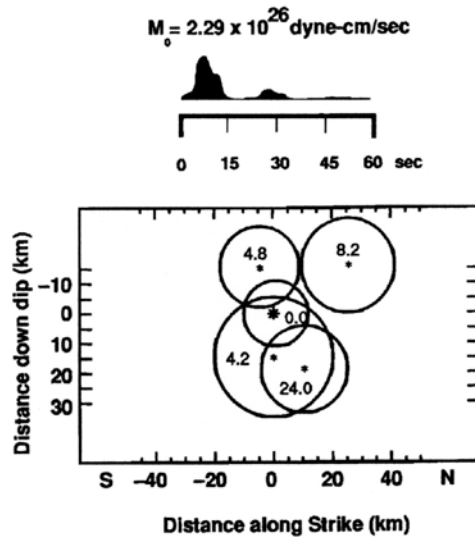
Gross (1996) proposed other heterogeneous slip models. They have the advantage of not only accounting for heterogeneous slip of large earthquakes due to the breaking of major asperities but also of predicting complex sources for small events, i.e., earthquake ruptures with dimensions less than the thickness of the seismogenic zone of brittle fracturing. On the other hand, for small earthquakes, the Gross models resemble classical crack theory by yielding also the well established area scaling relationship between average slip  $S$  and rupture area  $A$ , which is  $S \propto \sqrt{A}$ .

The rupturing of local asperities produces most of the high-frequency content of earthquakes. Accordingly, they contribute more to the cumulative seismic energy release than to the moment release. This is particularly important for engineering seismological assessments of expected earthquake effects. Damage to (predominately low-rise) structures is mainly due to frequencies  $> 2$  Hz. They are grossly underestimated when analyzing strong earthquakes only on the basis of medium and long-period teleseismic records or when calculating model spectra assuming smooth (average point source) rupturing along big faults of large earthquakes.

First results of a detailed investigation of rupture propagation by means of a nearby linear accelerometer line to the Oct. 15, 1979  $M_s = 6.9$  ( $M_w = 6.4$ ) Imperial Valley, California, earthquake were published by Spudich and Cranswick (1984). They showed that the observed high-frequency ground motions (accelerations up to 1.9 g were measured during this earthquake!) originate at irregularly distributed regions on the fault surface or are due to variable rupture velocity or both.

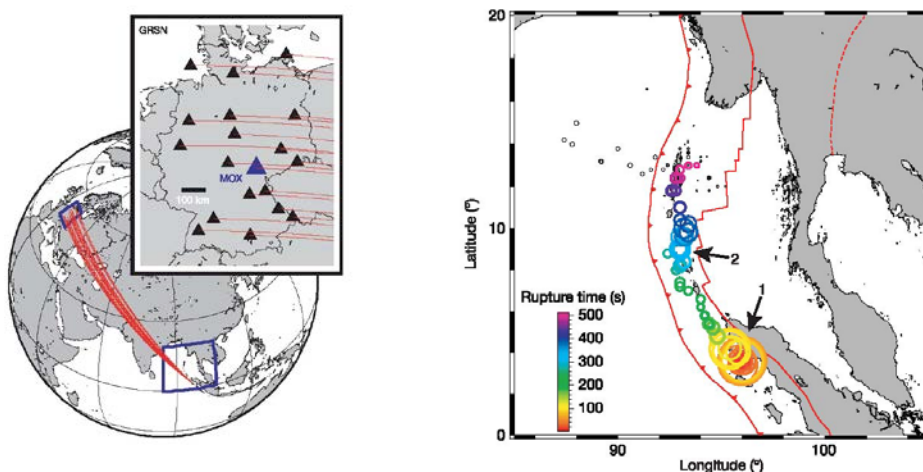
Goldstein and Archuleta (published 1991) made first direct measurements of 2-D earthquake rupture propagation using the SMART 1 array in Taiwan. They showed that the Jan. 29, 1981  $M_l = 6.3$  Taiwan earthquake ruptured unilaterally up-dip and towards the west on a  $60^\circ$  dipping and  $109^\circ$  striking reverse fault. They calculated a fault length of  $25 \pm 18$  km and a rupture duration of  $T_R = 9.4 \pm 3.6$  s. The latter is in perfect agreement with formula (3.3) for average magnitude-dependent rupture duration. Similarly, the overall rupture duration of 35 s of the Nov. 14, 1986, the  $M_w = 7.3$  Hualien, Taiwan, earthquake, also agreed within 10 % with the 38 s expected according to formula (3.3), although its source-time function showed two distinct sub-events with the by far dominating first one breaking within some 12 s only

(see Fig. 3.11). Thus, these strong-motion data highlighted that realistic strong-motion modeling requires much more detailed information about the source process than long-period moment tensor point source solutions can provide. Moreover, a detailed analysis of teleseismic data allowed to prove for the Hualien earthquake a clear migration in both horizontal and vertical directions of the patches of main seismic moment release (Hwang and Kanamori, 1989; Fig. 3.11 below), thus confirming the complexity of the source process both in space and time.



**Fig. 3.11** Source-time function (top) and spatial and time migration of areas of main moment release (bottom) of the 1986  $M_w = 7.3$  Hualien earthquake, Taiwan according to Hwang and Kanamori (1989), derived from teleseismic observations. Circle radii are proportional to the released seismic moment, numbers in the middle of the circles give the time after origin time in seconds. (Copy of Fig. 12 in Goldstein and Archuleta, J. (1991), *Geophys. Res.*, 96, p. 6197; © with kind permission of American Geophysical Society).

After the devastating great 2004  $M_w 9.3$  Sumatra-Andaman earthquake Krüger and Ohrnberger (2005) used the large aperture (500 km by 700 km) German Regional Seismic Network (GRSN) as a broadband array to track the propagation of this some 1200 km long rupture from a distance of about 9000 km (Fig. 3.12). An animation or the rupture tracking can be viewed and downloaded from section 2.5 in IS 1.1.



**Fig. 3.12 Left:** Source area of the great 2004 Sumatra-Andaman earthquake and great circle paths of from the epicenter location to the German Regional Seismic Network. **Right:** Map of maxima of seismic energy released during the earthquake in space and time from below upward. Circle radii are proportional to released seismic energy, circle colors are coded to rupture time. (Compiled from Figures 1 and 3 of Krüger and Ohrnberger (2005), *Nature*, 435, p. 937 and 938; © granted with kind permission by Nature Publishing Group).

The following conclusions could be drawn from this teleseismic array rupture tracking:

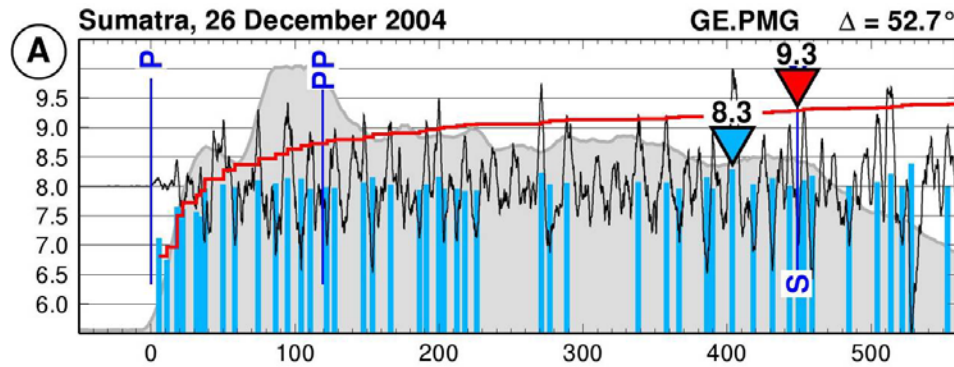
- The propagating rupture front could be tracked over a total length of 1,150 km;
- The average rupture speed was estimated to be 2.3 – 2.7 km/s but with slightly higher velocity on the southern segment (2.4-2.8 km/s) as compared to the northern branch of the rupture (2.1-2.4 km/s);
- The total rupture duration was at least 430, probably between 480 and 500 s;
- During the first 60 s, the position of the energy maximum did not move, which hints to a phase of bilateral growth of the rupture surface, whereas the remaining rupture was unilateral towards the north, with a second center of major energy release about 600 km NNW of the epicenter, which developed some 300 s after rupture initiation;
- The rupture did not progress farther to the south-southeast at the beginning of the rupture. This may indicate that the rupture front hit a barrier in this direction, which broke, in fact three months later during the March 28, 2005,  $M_w = 8.5$  earthquake;
- The complete estimate of source extension by means of rupture tracking of such a great event could be made available within about 30 min when using teleseismic station networks or large aperture arrays.

It should be noted, however, that first rough estimates of several of these aspects of the rupture process as a function of time, such as rupture duration, rise time and identifying time segments of increased energy radiation could already be derived within about 10 to 15 min. after origin time by processing and plotting single station records according to the procedure described by Bormann and Saul (2009b) for a fully automatic determination of the cumulative body-wave magnitude  $mB_c$  (for details see section 3.2.8.1). Applied to a record of the great Sumatra-Andaman earthquake at epicentral distance of  $52.7^\circ$  (Fig. 3.13) one realizes that

- initial rupture episodes around 30 s and 100 s after rupture initiation correspond already to sub-event broadband body-wave magnitudes  $mB$  around 8;
- the second episode released by far most of the energy of the whole rupture process;
- other, although less pronounced, episodes around 300s and 400 s after rupture initiation resulted in similarly large sub-event  $mB$  values (up to 8.3);
- thus the source-time function of the earthquake was rather asymmetric and therefore could not be approximated well by a symmetric triangular moment-rate function as prescribed by the routine Harvard/GCMT moment tensor procedure;
- the latter point, together with the fact that the longest periods analyzed by the GCMT standard procedure were shorter than the overall rupture duration, explains the initial underestimation of  $M_w(\text{GCMT}) = 8.9$  and  $9.0$ , respectively, for this earthquake. This could only be corrected by a later offline multiple CMT source analysis, which approximated the source-time function by trapezoid source-time functions of 5 individual sources, yielding then an overall  $M_w = 9.3$  (Tsai et al., 2005);
- cumulative  $mB_c$  may yield the same value straight away as the asymptotic “saturation” value of the real-time development of  $mB_c$  as a function of time;
- the rupture duration can be estimated reasonably well from the average amplitude envelopes of several high-frequency (1-3Hz) filtered records, when they fall below 40% of their maximum amplitudes (corresponding to 16% of the maximum energy

release), even after the arrival of stronger but in high-frequency content depleted secondary arrivals after P (see also Kanamori and Helmberger, 2005);

- thus estimated rupture duration of 540 s for the Sumatra-Andaman earthquake estimate agrees well (within about 10%) with similar independent estimates (e.g., Lomax, 2005; Kanamori and Helmberger, 2005).
- 



**Fig. 3.13** Velocity-proportional recording of the great Sumatra-Andaman earthquake at station PMG, processed with the mBc procedure originally proposed by Bormann and Khalturin (1975) and developed as a fully automated procedure by Bormann and Saul (2009b). Blue bars correspond to mB values of major amplitude onsets according to the condition  $V_i \geq 0.6V_{\max,t}$  with  $V_{\max,t}$  being the maximum velocity amplitude up to the time  $t$ , with the largest single amplitude mB marked by a blue triangle. The red curve shows the development of mBc with time and the red triangle the mBc values reached prior to the onset of S. The grey shaded area marks the high-frequency P-wave envelop (courtesy of J. Saul, 2009)

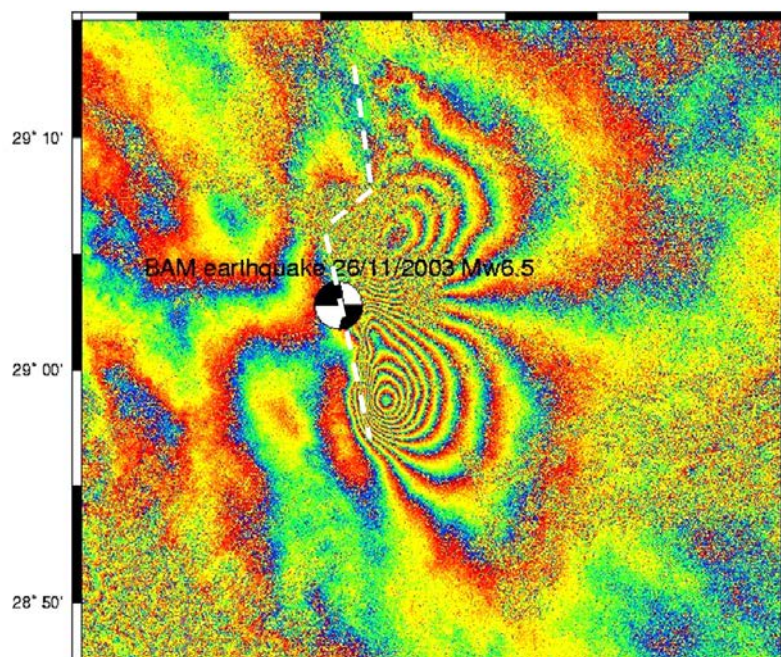
With respect to source complexity and related energy radiation of this gigantic earthquake Kanamori (2006) added that the ratio  $E_s/M_0 = 4.6 \times 10^{-6}$  was slightly smaller than that for other large subduction-zone earthquakes and that the average radiation efficiency was only 0.16, which ranges between regular earthquakes and tsunami earthquakes. Yet this values changed with time during the rupture process, being initially 0.21 for the southern Sumatra segment and dropping down to 0.053 for the Nicobar segment in the North. The very low value hints to large amounts of energy dissipation associated with water-filled thick sediments.

For another very interesting example of rupture tracking (with animation) of the rather complex Mw8.8 Chile earthquake of 2010 see section 2.6 or IS 1.1. The tracking was made by using records of US Array and permanent California network stations. The rupture was neither unilateral nor bilatetelal but “jumped” several times forth and back in several sub-events covering the whole subsequent aftershock area, which extended over some 600 km.

Nowadays, modern geodetic data may provide very useful independent information on seismic moment tensor, rupture mechanism and magnitude, even in the early warning context, e.g., for real-time predictions of the amplitudes of local tsunami, which strongly depend on the rupture model and source-time function. Sobolev et al. (2006 and 2007) demonstrated how reliable input data for the simulation process, which are for local tsunami mainly the mean and maximum fault slip rather than seismic moment, could be acquired by GPS-Shield

arrays that are strategically deployed on off-shore islands in front of the mainland coasts at risk. But a GPS-Shield array or even a single GPS station can also resolve seismic moment and thus moment magnitude of the rupture with very high accuracy, even if it is located several 100 km away from the trench. As a minimum, the GPS-Shield arrays placed along the trench will allow to estimate  $M_0$  and  $M_w$  for the corresponding sections of a rupture zone (partial magnitude) within just a few minutes of an earthquake.

Additionally, imaging synthetic aperture satellite radar interferometry (INSAR) (nowadays allows to derive reliable information about the areal distribution of surface deformation patterns that have been produced by earthquakes, even if the faults did not rupture through the surface. The interferometric fringe patterns are particularly well developed in vegetation-less arid areas (see Fig. 3.14).



**Fig. 3.14** INSAR Fresnell-zone patterns from the 26 December 2003, Mw 6.5 Bam earthquake, Iran, together with the teleseismic point-source fault-plane solution and the more detailed approximate fault trend (broken white line) with a step-over towards NE as inferred from the Fresnell-zone patterns. The maximum surface uplift and subsidence (in the line of sight to satellite) are 30 cm and 18 cm, respectively, the largest inferred fault slip 270 cm and the estimated total rupture length about 24 km. More than 80% of the seismic moment was released from the southern 13 km long fault segment. This preliminary figure has kindly be made available by R. Wang, GFZ German Research Centre for Geosciences. For more detailed final results and interpretation see Wang et al. (2004).

### 3.1.5 Summary and conclusions

The detailed understanding and quantification of the physical processes and geometry of seismic sources is one of the ultimate goals of seismology, be it in relation to understanding tectonics, improving assessment of seismic hazard or discriminating between natural and anthropogenic events. Earthquakes can be quantified with respect to various geometrical and



physical parameters such as time and location of the (initial) rupture and orientation of the fault plane and slip, fault length, rupture area, amount of slip, magnitude, seismic moment, radiated energy, stress drop, duration and time-history (complexity) of faulting, particle velocity, acceleration of fault motion etc. It is impossible, to represent this complexity with just a single number or a few parameters.

There are different approaches to tackle the problem. One aims at the detailed analysis of a given event, both in the near- and far-field, analyzing waveforms and spectra of various kinds of seismic waves in a broad frequency range up to the static displacement field, complemented by field investigations into macroseismic effects and surface expressions of fault rupture. Such detailed and complex investigations usually require a lot of time and effort. They may be feasible only for selected important events and are usually beyond the scope of routine seismological observatory practice. Nonetheless, we have deliberately dwelled on the issue of source complexity and the multitude of parameters required for describing them adequately. Because many of these parameter have to be derived by detailed analysis of high quality data collected by seismological observatories in a wider sense, permanent or temporary ones, physical or virtual seismic networks (Chapter 8) or seismic arrays (Chapter 9) of local, regional or even global scale or range of operation (see Chapter 8), measuring weak or strong motions.

With dense strong-motion networks deployed in source areas of potentially large earthquakes, as they become now increasingly available in many regions at risk, detailed pictures of the fracture process may already be obtained in near real-time. They are of greatest importance for more reliable shake-map calculations, and thus for a better guidance of relief efforts and as empirical references for both earthquake engineers, land-use planners and others involved in disaster preparedness and mitigation (see Chapter 15).

The second, more simple approach is usually taken at seismological observatories and data centers for the routine analysis of mass data. They describe the seismic source only by a limited number of parameters such as the origin time and (initial rupture) location, magnitude, intensity or acceleration of measured ground shaking, sometimes complemented by fault-plane solutions. These parameters can easily be obtained and have the advantage of rough but quick information being given to the public and concerned authorities. Furthermore, this approach provides standardized data for comprehensive earthquake catalogs which are fundamental for other kinds of research such as earthquake statistics and seismic hazard assessment. But we need to be aware that these simplified, often purely empirical parameters can not give a full description of the true nature and geometry, the time history nor the energy release of a seismic source. In the following we will focus on the most common procedures in routine seismological practice.

## **3.2 Magnitude of seismic events (P. Bormann)**

### **3.2.1 History, scope and limitations of the magnitude concept**

The concept of *magnitude* was introduced by Richter (1935) to provide an objective instrumental measure of the size of earthquakes. The term magnitude was recommended to Richter by H. O. Wood in distinction to the name *intensity* scale, which is based on the assessment and classification of shaking damage and human perceptions of shaking and thus depends on the distance and depth of the seismic source (see Chapter 13). In contrast, the

magnitude  $M$  uses instrumental measurements of the ground motion adjusted for epicentral distance and source depth. Standardized instrument characteristics were originally used to avoid instrumental effects on the magnitude estimates. Thus it was hoped that  $M$  could provide a single number to quantify earthquake size which was also approximately related to the released seismic energy. However, as outlined in 3.1 above, such a simple empirical parameter is not directly related to any physical parameter of the source. Rather, the magnitude scale is currently viewed as providing a quickly determined simple " ... parameter which can be used for first-cut reconnaissance analysis of earthquake data (catalog) for various geophysical and engineering investigations; special precaution should be exercised in using the magnitude beyond the reconnaissance purpose" (Kanamori, 1983).

In the following we will use mainly the magnitude symbols, sometimes with slight modification, as they have historically developed and are still predominantly applied in common practice. However, as will be shown later, such "generic" magnitude symbols are often not explicit enough to specify the type of seismographs, components and phases these magnitudes are based. This requires more "specific" magnitude names where higher precision is required (see IS 3.2). Moreover, in order to assure in future a unique nomenclature, at least for the recently recommended IASPEI standards for widely used magnitudes (see IASPEI 2005 and 2013 as well as IS 3.3), we will use in the following only this new nomenclature when the respective new procedures and so derived data are presented.

The original Richter magnitude,  $M_L$  or  $M_L$ , was based on maximum amplitudes measured on records of the standardized short-period Wood-Anderson (WA) seismometer network in Southern California that are displacement-proportional at periods less than 0.8 s. Such records were suitable for the classification of local shocks in that region. In the following we will name it  $M_l$  (with "l" for "local") in order to avoid on the one hand confusion with more specific names for magnitudes from surface waves where the phase symbol  $L$  stands for unspecified long-period surface waves (e.g., in  $M_{LV}$  or  $M_{LH}$ ), but also, in order to differentiate between local magnitudes scales in general with often unknown scaling, respectively calibration, to the Southern California standard, and those which have been derived and properly scaled according to the new IASPEI recommendations, which are written  $M_L$  or  $M_L$ , respectively.

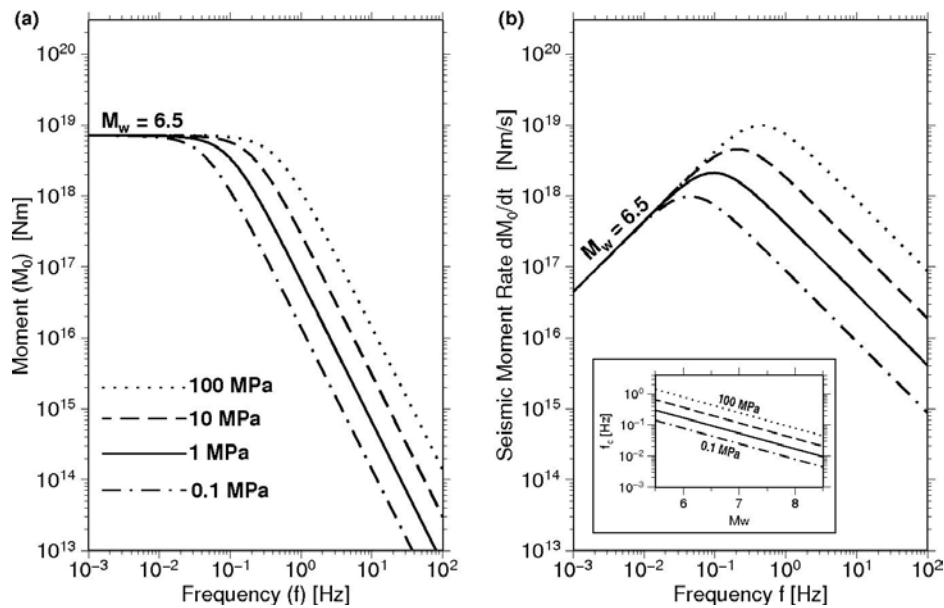
Gutenberg and Richter (1936) and Gutenberg (1945a, b and c) extended the magnitude concept so as to be applicable to ground motion measurements from medium- and long-period seismographic recordings of both surface waves (written  $M_s$  or  $M_s$ ) and different types of body waves ( $m_B$  or  $m_B$ ) in the teleseismic distance range. For the magnitude to be a better estimate of the seismic energy  $E_s$ , they proposed for  $m_B$  to divide the measured displacement amplitudes by the associated periods to obtain ground velocities and presented in Gutenberg and Richter (1956) the first energy-magnitude relationship  $\log E_s = 2.4 m_B - 1.2$  (when  $E_s$  is given in Joule). Although they tried to mutually scale the different types of magnitudes in order to match at certain magnitude values, it was realized that these scales are only imperfectly consistent with each other. Therefore, Gutenberg and Richter (1956a and b) provided correlation relations between various magnitude scales (see 3.2.9.2).

After the deployment of the World Wide Standardized Seismograph Network (WWSSN) in the 1960s it became customary to determine the body-wave magnitude only on the basis of short-period narrow-band vertical component P-wave recordings only. This short-period body-wave magnitude was termed  $m_b$  (or  $m_b$ ). The introduction of  $m_b$  increased the



inconsistency between the magnitude estimates from body and surface waves. The main reasons for this are:

- different magnitude scales use different periods and wave types which carry different information about the source process and the radiated source spectra;
- the spectral amplitudes radiated from a seismic source increase linearly with its seismic moment for frequencies  $f < f_c$  ( $f_c$  – corner frequency). This increase with moment, however, is reduced for  $f > f_c$  (see Fig. 3.5). This changes the balance between high- and low-frequency content in the radiated source spectra as a function of event size;
- the maximum seismic energy is released around the corner frequency of the displacement spectrum because this relates to the maximum of the ground-velocity spectrum (see Fig. 3.5). Accordingly,  $M$ , when supposed to be a measure of the seismic energy released, strongly depends on the position of the corner frequency in the source spectrum with respect to the pass-band of the seismometer used for the magnitude determination;
- for a given level of long-period displacement amplitude and thus seismic moment  $M_0$ , the position of the corner frequency is mainly controlled by the stress drop in the source (see section 3.1.2.5). Accordingly, high stress drop results in the excitation of more high frequencies (Fig. 3.15) and thus earthquakes with equal  $M_w$  may have rather different energy magnitudes  $M_e$  or short-period magnitudes such as  $m_b$  or  $M_L$ .



**Fig. 3.15** (a) Far-field displacement and (b) velocity source spectra scaled to seismic moment and moment rate, respectively, for a model earthquake with  $M_w = 6.5$  but different stress drop  $\Delta\sigma$  in units of MPa. The spectra were calculated based on the model assumptions explained for Fig. 3.5. The inset in Fig. 3.16(b) shows the variation of the corner frequency  $f_c$  obtained according to the Brune (1970) equation  $f_c = c v_s (\Delta\sigma/M_0)^{1/3}$  in a wider range of  $M_w$  for varying  $\Delta\sigma$  in increments of one order between 0.1 and 100 MPa. (Copy of Fig. 2, p. 415 in Bormann and Di Giacomo 2011: The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15** (2), 411-427; © Springer Publishers).

- seismographs with different transfer functions sample the ground motion in different frequency bands with different bandwidth. Therefore, no one-to-one agreement of the magnitudes determined on the basis of their records can be expected;
- additionally, band-pass recordings distort the recording amplitudes of transient seismic signals, the more so the narrower the bandwidth is. This can not be fully compensated by correcting only the frequency-dependent magnification of different seismographs based on their amplitude-frequency response. Although this is common seismological practice in order to determine so-called "true ground motion" amplitudes for magnitude calculation, it is not fully correct. The reason is that the instrument magnification or amplitude-frequency response curves are valid only for steady-state oscillation conditions, i.e., after the decay of the seismograph's transient response to an input signal (see section 4.5 in Chapter 4). True ground motion amplitudes can be determined only by taking into account the complex transfer function of the seismograph (see Chapter 5) and, in the case of short transient signals, by signal restitution in a very wide frequency band (Seidl, 1980; Seidl and Stammer, 1984; Seidl and Hellweg, 1988). Only recently a calibration function for very broadband P-wave recordings has been published (Nolet et al., 1998), however it has not yet been widely applied, tested and approved (see section 3.2.6.1).

Efforts to unify or homogenize the results obtained by different methods of magnitude determination into a common measure of earthquake size or energy have generally been unsuccessful (e.g., Gutenberg and Richter, 1956a; Christoskov et al., 1985). Others, aware of the above mentioned reasons for systematic differences, have used these differences for better understanding the specifics of various seismic sources, e.g., for discriminating between tectonic earthquakes and underground nuclear explosions on the basis of the ratio  $m_b/M_s$ . Duda and Kaiser (1989) recommend the determination of different spectral magnitudes, based on measurements of the spectral amplitudes from one-octave bandpass-filtered digital broadband velocity records (see section 3.2.6.1).

Another effort to provide a single measure of the earthquake size was made by Kanamori (1977). He developed the seismic moment magnitude  $M_w$ . It is tied to  $M_s$  but does not saturate for big events because it is based on seismic moment  $M_0$ , which is determined via the asymptotic plateau of low-frequency spectral displacement amplitudes for  $f \ll f_c$ . This level increases linearly with  $M_0$ . According to Eq. (3.1),  $M_0$  is proportional to the average static displacement and the area of the fault rupture and is thus a good measure of the total deformation in the source region. On the other hand it is (see the above discussion on corner frequency and high-frequency content) not as good a measure of earthquake size in terms of seismic energy release. This limits the usefulness of  $M_w$  for assessing and specifying the type and severity of seismic hazard and related risks when presented in so-called unified seismic catalogs or in near real-time earthquake alert messages as the only earthquake size parameter. Most earthquake damage is usually related to medium and low-rise structures with eigenfrequencies  $f > 0.5$  Hz (i.e., lower than about 20 stories) and mainly caused by high-frequency strong ground motion. Energy magnitude  $M_e$ , which may differ for some seismic events by one magnitude unit or even more from  $M_w$ , is then a very useful complement, as are the short-period magnitudes  $m_b$  and  $M_l$  that are measured at periods of particular relevance ( $T \approx 0.1$  s to 3 s) for earthquake engineers, engineering seismologists and disaster managers. However, there is no single number magnitude parameter available which could serve alone as a sufficiently reliable estimator of earthquake "size" and related potential

earthquake effects and risks in all their different aspects, tsunami threat included. Because the latter are not directly related to any specific type of the magnitude or the amount of stress drop in the source. Rather, they may be strongly modified by source directivity (see Fig. 3.112), source depth, attenuation conditions along the wave propagation path, wave focusing or defocussing by lateral heterogeneities, the very local site conditions (e.g., in the case of tsunami effects the detailed shallow water bathymetry and coastal configuration) etc. In any event, however, it is advisable to be guided by taking complementary magnitudes into account.

What is needed in practice are at least two parameters to characterize roughly both the source dimensions and the high-frequency energy of a seismic event, namely  $M_0$  and  $f_c$  or  $M_w$  together with  $m_b$  or  $M_l$ , respectively, or a comparison between the moment magnitude  $M_w$  and the energy magnitude  $M_e$  (e.g., Choy and Kirby, 2004; Di Giacomo et al, 2010a; Bormann and Di Giacomo, 2011; and IS 3.5 by G. Choy in this Manual).  $M_e$  can today be determined from direct energy calculations based on the integration of digitally recorded waveforms of broadband velocity (Seidl and Berckhemer, 1982; Berckhemer and Lindenfeld, 1986; Boatwright and Choy, 1986; Kanamori et al., 1993; Choy and Boatwright, 1995) (see section 3.2.7.2 and IS 3.6).

Despite their limitations, common magnitude estimates have proved to be suitable also for getting, via empirical relationships, quick but rough estimates of other seismic source parameters such as the seismic moment  $M_0$ , stress drop, amount of radiated seismic energy  $E_s$ , length  $L$ , radius  $r$  or area  $A$  of the fault rupture, as well as the intensity of ground shaking,  $I_0$ , in the epicentral area and the probable extent of the area of felt shaking (see sections 3.2.6.6 and 3.3.4).

Magnitudes are also crucial for the quantitative classification and statistical treatment of seismic events aimed at assessing seismic activity and hazard, studying variations of seismic energy release in space and time, etc. Accordingly, they are also relevant in earthquake prediction research. All these studies have to be based on well-defined and stable long-term data. Therefore, magnitude values – notwithstanding their inherent systematic biases as discussed above - have to be determined over decades and even centuries by applying rigorously clear and well documented stable procedures and well calibrated instruments. Any changes in instrumentation, gain and filter characteristics have to be precisely documented in station log-books or event catalogs and data corrected accordingly. Otherwise, wrong conclusions may be drawn from research based on incompatible data.

Being aware, on the one hand, of the inherent problems and limitations of the magnitude concept in general and specific magnitude estimates in particular and, on the other hand, of the urgent need to strictly observe reproducible long-term standardized procedures of magnitude determination, we will review below most of the magnitude scales applied in seismological practice. We will then introduce the newly recommended IASPEI measurement standards for some of the widely used magnitude scales. They have been outlined in more detail and commented with respect to their relationship to the classical magnitude scales in IS 3.3. Early comprehensive reviews of the complex magnitude issue have been published by Båth (1966, 1981), the most recent condensed ones were authored by Bormann and Saul (2009a) and Bormann (2011). Various special volumes with selected papers from symposia and workshops on the magnitude problem appeared in *Tectonophysics* (Vol. 93, No.3/4 (1983); Vol. 166, No. 1-3 (1989); Vol. 217, No. 3/4 (1993)).

### 3.2.2. General assumptions and definition of magnitude

Magnitude scales are based on a few simple assumptions, e.g.:

- for a given source-receiver geometry, "larger" events will produce wave arrivals of larger amplitudes at the seismic station. The logarithm of ground motion amplitudes  $A$  is used because of the enormous variability of earthquake generated amplitudes;
- magnitudes should, as preferred by Gutenberg when proposing the  $m_B$  scale, be a measure of radiated seismic energy and thus be proportional to the velocity of ground motion, i.e., to  $A/T$  with  $T$  as the period of the considered wave;
- the decay of ground displacement amplitudes  $A$  with epicentral distance  $\Delta$  and their dependence on source depth  $h$ , i.e., the effects of geometric spreading and attenuation of the considered seismic waves, is known at least empirically in a statistical sense. It can be compensated for by a calibration function  $\sigma(\Delta, h)$ . The latter is the log of the inverse of the reference amplitude  $A_0(\Delta, h)$  of an event of zero magnitude, i.e.,  $\sigma(\Delta, h) = -\log A_0(\Delta, h)$ ;
- the maximum value  $(A/T)_{\max}$  measured on the recorded waveform radiated by a point source of a seismic phase for which  $\sigma(\Delta, h)$  is known should provide the best and most stable estimate of the event magnitude;
- regionally variable preferred source radiation-pattern and *directivity* and/or focusing/defocusing conditions, which may be significant especially in the case of surface waves, may be corrected by (a) regional source correction term(s),  $C_r$ , and the influence of local site effects on amplitudes (which depend on local crustal structure, near-surface rock type, soft-soil cover and/or topography) may be accounted for by a station correction,  $C_s$ , which is assumed not to depend on azimuth.

Accordingly, the general form of all magnitude scales based on measurements of ground displacement amplitudes  $A_d$  and periods  $T$  is:

$$M = \log(A_d/T)_{\max} + \sigma(\Delta, h) + C_r + C_s. \quad (3.13)$$

**Problem 1: Calibration functions used in common practice do not consider a frequency dependence of  $\sigma(\Delta, h)$**

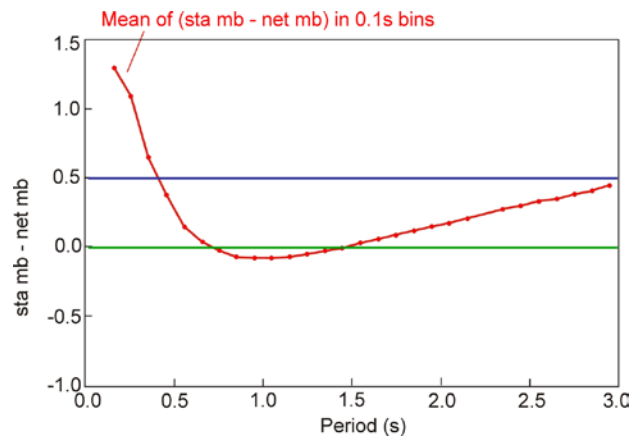
Theoretical calculations by Duda and Yanovskaya (1993) for different attenuation models and by Di Giacomo et al. (2008) for P waves, based on the assumption of  $Q$  values that do not depend on frequency, show that differences in  $A(\Delta, T)$  may become  $> 0.6$  m.u. for  $T < 3$  s, however they are  $< 0.2$  for  $T > 4$  s and thus they are more or less negligible for magnitude determinations in the medium- and long-period range. Yet, magnitude determinations based on the measured ratio  $A/T$ , respectively  $A \times f$ , largely compensate for the frequency-dependent attenuation of  $A$ . Accordingly,  $A/T$  based magnitudes are more stable in a wider range of periods than  $A$  based magnitudes.

In section 3.2.5.2 we will show that the differences in magnitude estimates between short-period  $m_b$ , when  $A$  is measured as the maximum amplitude in the whole P-wave train, and the more long-period magnitude estimates ( $m_B$  and  $M_s$ ) can fully be explained by differences in the frequency at which the amplitudes are measured with respect to the corner frequency of

the radiated source spectrum (see Figs. 3.5 and 3.15). We term this effect the *spectral component of saturation*. Accounting additionally for excessively large attenuation differences at higher frequencies, as proposed by some of the models, would result in even larger differences between teleseismic mb, mostly measured at 0.5 to 2 Hz, mB (mostly measured between  $0.03 \text{ Hz} < f < 0.3 \text{ Hz}$ ) and Ms measured around 20 s period. Therefore, the assumption of frequency-dependent Q according to the absorption model of Liu et al. (1976) seems better to explain both the differences in mb and mB as well as field observations by Der et al. (1982). For the time being, this relaxes the necessity to introduce into the general magnitude formula (3.13) an additional frequency-dependent attenuation term, neither for the two standard body-wave magnitudes nor for the broadband surface-wave magnitude Ms(BB), until better empirical data become available. Yet, this matter requires further studies.

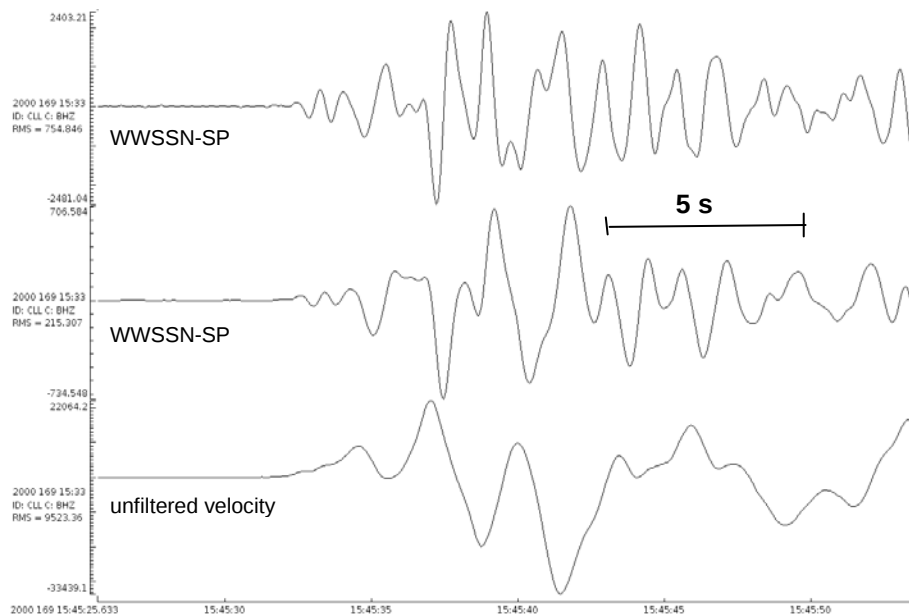
### **Problem 2: $(A/T)_{\max}$ or $(A_{\max}/T)$ ?**

For ease of record analysis it has become widespread practice to measure the easily recognizable largest record amplitude with its period T, to correct it for the period-dependent displacement response and only then to determine the ratio  $A_{\text{dmax}}/T$ . This, however, may not be the real  $(A_d/T)_{\max}$  intended by Gutenberg as a measure of maximum ground motion velocity and thus a close estimator of the radiated seismic energy. Especially, when T is measured in the period range of already steeply dropping amplification of the response curve differences in magnitude estimates may then reach several tenth of magnitude units with respect to the correct procedure (m.u.). This has been identified at the NEIC as a major problem when measuring in records of the WWSSN-SP type (see Fig. 3.19 below) the largest P-wave amplitude at period between about 2 to 3 s. The average deviations as compared to network averages may then reach about +0.5 m.u. (see Fig. 3.16), also when measuring at periods around 0.3 s, which is, however, rarely the case. This lead to the agency decision to calculate mb only from amplitudes measured at periods  $T < 2 \text{ s}$  instead of up to  $< 3 \text{ s}$ , as recommended by the classical mb measurement standard. The reason is that correcting the measured trace amplitudes for the period-dependent instrument response may in the range of the very strong decay of the WWSSN-SP response with 3<sup>rd</sup> order for periods larger than about 1.5 s yield much too large ground amplitudes at already slightly too large measured periods.



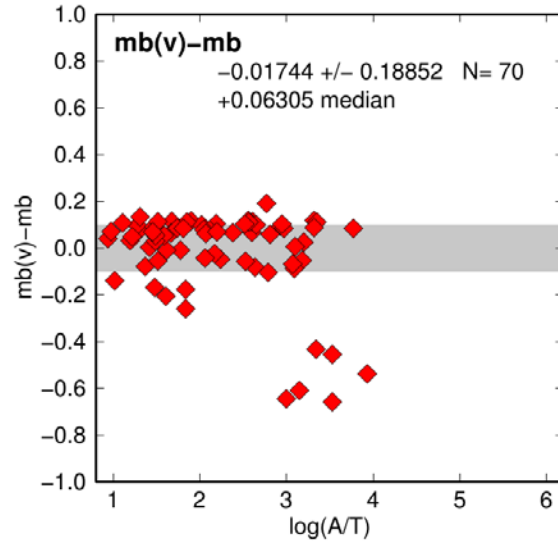
**Fig. 3.16** Mean difference between individual station mb and network mb in 0.1 m.u. bins as determined by the NEIC of the US Geological Survey as a function of period at which the maximum amplitude A for mb has been measured (simplified from a Figure by J. W. Dewey et al., 2011 ).

Although not yet fully understood in detail, this problem can obviously be reduced by measuring instead of  $A_{\max}$  on the WWSSN-SP filtered record trace the largest velocity amplitude  $V_{\max} = 2\pi(A/T)_{\max}$  in the time-differentiated WWSSN-SP record trace (see, e.g., Fig. 3.46).. One reason seems to be that - in comparison to WWSSN-SP records – their time-derivatives are more high-frequency with more harmonically looking oscillations and easier to identify related maximum peak-to-trough deflections that really relate to  $(A/T)_{\max}$ . In contrast, WWSSN-SP records tend to be more complex, making it sometimes more difficult to properly read the period which relates to the largest trace amplitude, respectively to the proper  $(A/T)_{\max}$ . This may, as in the case of Fig. 3.17, result in overestimating the  $T$  related to the largest trace amplitude and thus in strong overestimation of the ground motion amplitude  $A$  and thus  $mb$ . Another reason may be that the ground velocity magnification of the WWSSN-SP response decays only with the 2<sup>nd</sup> order towards longer periods. This reduces the adverse effect of measurement errors in  $T$  when converting the measured velocity trace amplitude into ground velocity amplitude.



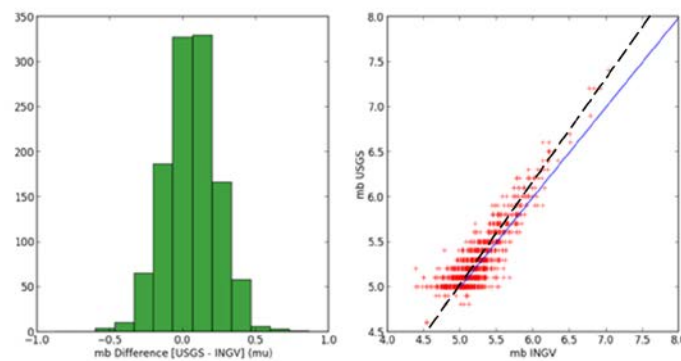
**Fig 3.17** Comparison of a P waveforms in an unfiltered velocity broadband record (bottom), in the respective WWSSN-SP simulated record (middle), and as a WWSSN-SP velocity trace. Figure amended according to a record plot kindly provided by J. Saul, 2012.

Fig. 3.18a plots the difference  $mb(V_{\max}) - mb(A_{\max}/T)$  for a test data set, both measured automatically. The cluster of strongly negative residuals between -0.4 and -0.65 m.u. relates to standard  $mb$  values measured at periods between 2.1 s and 2.8 s. For shorter periods the deviations between these two versions of  $mb$  measurement have been generally less than 0.2 m.u. with an average absolute deviation of only 0.07 m.u., but 0.11 m.u., when the outliers are included. The mean deviation is about -0.02 with the outliers and slightly positive, + 0.03 m.u., without them. This make  $mb(V_{\max})$  measurements a strong candidate procedure that holds promise to yields with greater ease (no period measurement required) on average  $mb$  values that are compatible with standard  $mb$  yet with greater stability and reduced data scatter. Further studies and documentation on this problem are encouraged.



**Fig. 3.18a** Difference between automatically measured mb values based on measuring the maximum velocity amplitude  $V/2\pi$  on time differentiated WWSSN-SP filtered traces minus mb based on A/T measurement in a wide range of  $\log(A/T)$  of 4 magnitude units. Note the strong outliers despite generally good agreement between these two ways. For explanation see text. (Figure by courtesy J. Saul, GFZ Potsdam).

As one step in this direction, Alberto Michelini from the Istituto Nazionale di Geofisica e Vulcanologia, Rome, Italy, compared on the request of the Editor for 1155 earthquakes between February 1<sup>st</sup>, 2012, and January 28<sup>th</sup>, 2013, mb(INGV), which is in fact an WWSSN-SP based  $mb(V_{\max})$ , with mb(USGS), which is supposed to be a IASPEI standard mb. The results have been plotted in Fig. 3.18b. One recognizes that the two magnitudes agree on average for the overwhelming majority of data with  $mb(\text{USGS}) < 6.0$  better than 0.1 m.u., yet the difference grows with magnitude up to about 0.3 m.u. The average difference as well as the median of  $mb(\text{USGS})-mb(\text{INGV})$  is 0.06 m.u., the standard deviation  $\pm 0.174$  m.u. and the average absolute deviation 0.145 m.u.



**Fig. 3.18b** **Left:** Histogram of the difference  $mb(\text{USGS})-mb(\text{INGV})$ ; **right:** data plot of  $mb(\text{USGS})$  over  $mb(\text{INGV})$  for 1155 earthquakes with  $4.5 < mb(\text{USGS}) < 7.5$ . The blue line is the 1:1 relationship, the broken black line an eye fit. Figure by courtesy of Alberto Michelini, INGV (2013).

### 3.2.3 General rules and procedures for magnitude measurement

#### 3.2.3.1 General procedures when working with analog data or on digital screen plots

Magnitudes can be determined on the basis of Eq. (3.13) by reading the displacement amplitudes  $A_d$  and their related periods  $T$  for calculating  $(A/T)_{\max}$  for body waves (e.g., P, S, PP) or surface waves (LQ or Lg, LR or Rg) for which calibration functions for either vertical (V) and/or horizontal (H) component records are available. If the period is measured on a seismogram recorded by an instrument whose response is in the considered period range already proportional to velocity, then  $(A_d/T)_{\max} = A_{v\max}/2\pi$ , i.e., the measurement can be directly determined from the maximum trace amplitude of this wave or wave group with only a correction for the velocity gain. In contrast, with displacement proportional records one may not know with certainty where  $(A/T)_{\max}$  is largest in the displacement waveform. Sometimes smaller amplitudes but associated also with shorter periods may yield larger  $(A/T)_{\max}$ . In the following we will always use  $A$  for  $A_d$ , and  $V$  for  $A_v$ , if not otherwise explicitly specified.

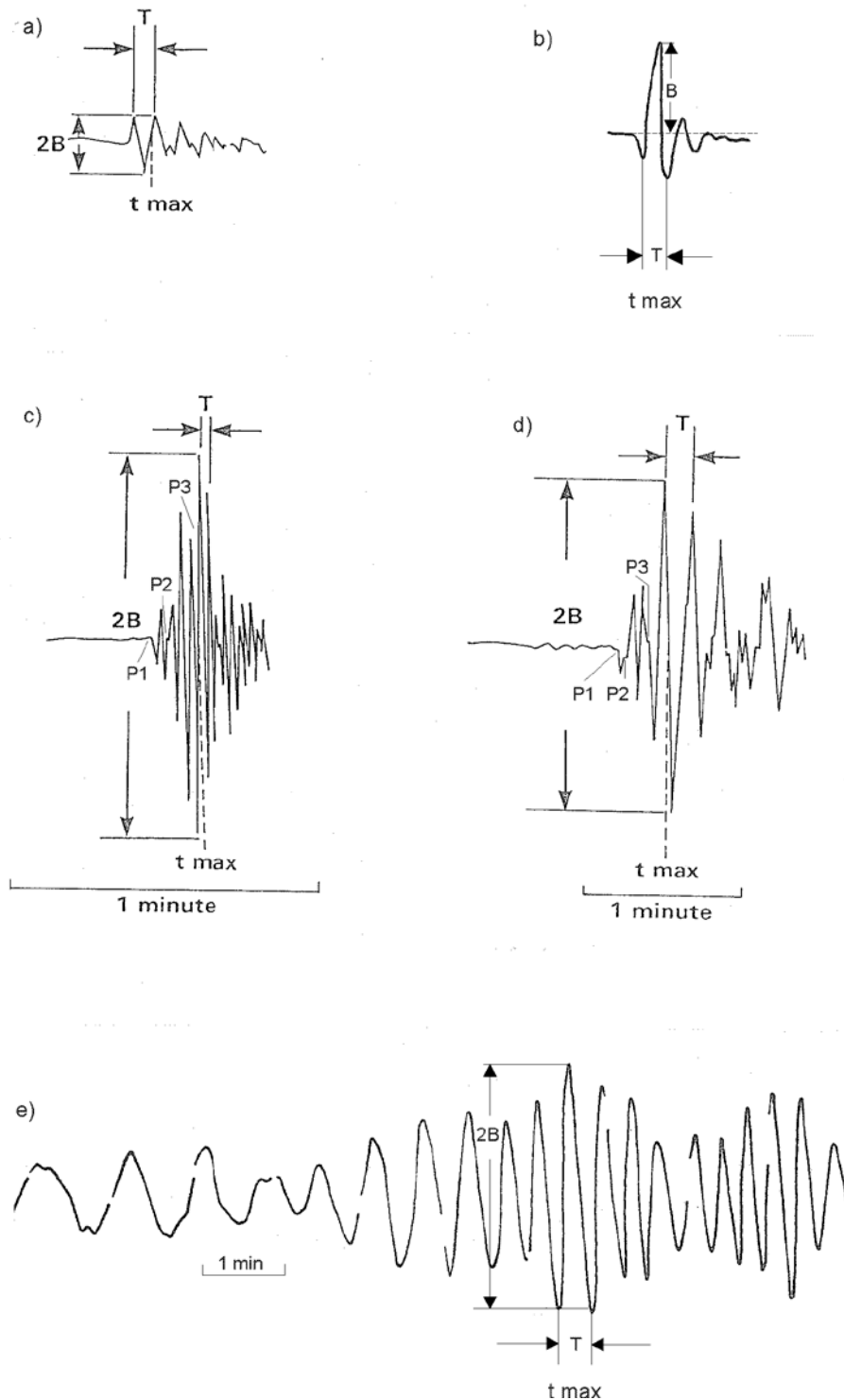
In measuring  $A$  and  $T$  from seismograms for magnitude determinations and reporting them to national or international data centers, **the following definitions and respective instructions** given in the Manual of Seismological Observatory Practice (Willmore, 1979) as well as in the recommendations by the IASPEI Commission on Practice from its Canberra meeting in 1979 (slightly modified and amended below) **should have been observed in the past**:

- The trace amplitude  $B$  of a seismic signal on a record is defined as its largest peak (or trough) deflection from the base-line of the record trace.
- For many phases, surface waves in particular, the recorded oscillations are more or less symmetrical about the zero line.  $B$  should then be measured either by direct measurement from the base-line or - preferably - by halving the peak-to-trough deflection (Figs. 3.19 a and c - e). For phases that are strongly asymmetrical (or clipped on one side)  $B$  should be measured as the maximum deflection from the base-line (Fig. 3.19 b).
- The corresponding period  $T$  is measured in seconds between those two neighboring peaks (or troughs) - or from (doubled!) trace crossings of the base-line - where the amplitude has been measured (Fig. 3.19);
- The trace amplitudes  $B$  measured on the record should be converted to ground displacement amplitudes  $A$  in nanometers (nm) or some other stated SI unit, using the  $A$ - $T$  response (magnification) curve  $\text{Mag}(T)$  of the given seismograph (see Fig.3.20); i.e.,  $A = B/\text{Mag}(T)$ . (**Note:** In most computer programs for the analysis of digital seismograms, the measurement of period and amplitude is done automatically after marking the position on the record where  $A$  and  $T$  should be determined).
- Amplitude and period measurements from the vertical component ( $Z = V$ ) are most important. If horizontal components ( $N$  - north-south;  $E$  - east-west) are available, readings from both records should be made at the same time (and noted or reported separately) so that the amplitudes can be combined vectorially, i.e.,  $A_H = \sqrt{(A_N^2 + A_E^2)}$ .
- When several instruments of different frequency response are available (or in the case of the analysis of digital broadband records filtered so as to match different standard responses),  $A_{\max}$  and  $T$  measurements from each should be reported separately and the



type of instrument used should be stated clearly (short-, medium- or long-period, broadband, Kirnos, Wood-Anderson, etc., or related abbreviations given for instrument classes with standardized response characteristics as in Tab. 3.1 and Fig. 3.20 left). For this the classification given in the old Manual of Seismological Observatory Practice (Willmore 1979) may be used.

- Broadband instruments are preferred for all measurements of amplitude and period.
- Note that earthquakes are often complex multiple ruptures. Accordingly, the time,  $t_{\max}$ , at which a given seismic body wave phase has its maximum amplitude may be quite some time after its first onset. Accordingly, in the case of P and S waves, the measurement should normally be taken within the first 25 s and 40-60 s, respectively, but in the case of very large earthquakes this interval may need to be extended to more than a minute. For subsequent earthquake studies it is also essential to report the measurement time  $t_{\max}$  (see Fig. 3.19).
- For teleseismic surface waves (i.e.,  $\Delta > 20^\circ$ ) the procedures are basically the same as for body waves. However,  $(A/T)_{\max}$ , often in the Airy phase of the dispersed surface wave train, occurs much later and should normally be measured in the period range between 16 and 24 s although both shorter and longer periods may be associated with the maximum surface wave amplitudes (see section 2.3 in Chapter 2 and Fig. 3.35 in section 3.2.5.1).
- Note that in displacement proportional records  $(A/T)_{\max}$  may not coincide in time with  $B_{\max}$ . Sometimes, in dispersed surface wave records in particular, smaller amplitudes associated with significantly smaller periods may yield larger  $(A/T)_{\max}$ . In such cases also  $A_{\max}$  should be reported separately. In order to find  $(A/T)_{\max}$  on horizontal component records it might be necessary to calculate  $A/T$  for several amplitudes on both record components and select the largest vectorially combined value. In records proportional to ground velocity, the maximum trace amplitude is always related to  $(A/T)_{\max}$ . Note, however, that as compared to the displacement amplitude  $A_d$  the velocity amplitude is  $A_v = V = A_d 2\pi/T$ .
- If mantle surface waves are observed, especially for large earthquakes (see section 2.3.4 of Chapter 2), amplitudes and periods of the vertical and horizontal components with the periods in the neighborhood of 200 s should also be measured.
- On some types of short-period instruments (in particular analog) with insufficient resolutions it is not possible to measure the period of seismic waves recorded from nearby local events and thus to convert trace deflections properly to ground motion. In such cases magnitude scales should be used which depend on measurements of maximum trace amplitudes only.
- Local earthquakes were often clipped in analog records of high-gain short-period seismographs with insufficient dynamic range. This made amplitude readings impossible. In this case magnitude scales based on record duration (see 3.2.4.5) might be used instead, provided that they have been properly scaled with magnitudes based on amplitude measurements.



**Fig. 3.19** Examples for measurements of trace amplitudes  $B$  and periods  $T$  in seismic records for magnitude determination: a) the case of a short wavelet with symmetric and b) with asymmetric deflections, c) and d) the case of a more complex P-wave group of longer duration (multiple rupture process) and e) the case of a dispersed surface wave train. Note: c) and d) are P-wave sections of the same event but recorded with different seismographs (classes A4 and C) while e) was recorded by a seismograph of class B3 (see Fig. 3.20).

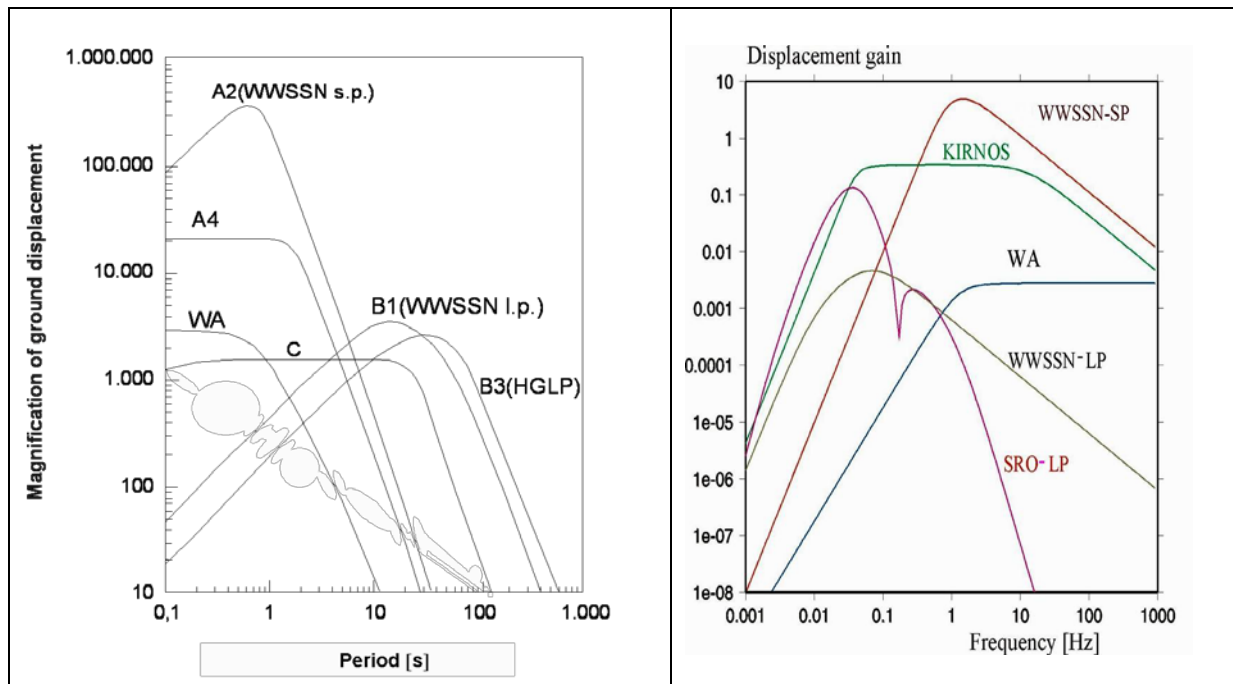
**Tab. 3.1** Example of a bulletin page of station Moxa (MOX), Germany, based on the analysis of analog photographic recordings. The event occurred on January 1967. Note the unambiguous annotation of the type of instruments and component used for the determination of onset times, amplitudes and periods. Multiple body wave onsets of distinctly different amplitudes, which are indicative of a multiple rupture process, have been separated. Types of analog standard seismographs used: A = A4 in Willmore (1979) = short-period with flat displacement response between 0.7 and 10 Hz, B = B3 = in Willmore (1979) = long-period 30-to-80 s seismograph, C = KIRNOS displacement broadband seismograph with flat response between 10 Hz and 20 s as in Fig. 3.20. Components: V = Z = vertical component; H = vectorially combined horizontal components; Lm - maximum of the long-period surface wave train.

| Day | Phase | Seismograph | h m s      | Remarks                                                                     |
|-----|-------|-------------|------------|-----------------------------------------------------------------------------|
| 5.  | +eiP1 | A           | 00 24 15.5 | <u>Mongolia</u> 48.08°N 102.80°E                                            |
|     | iP2   | A           | 24 21.5    | H = 00 14 40.4 h = normal MAG = 6.4                                         |
|     | iP3   | A,C         | 24 28.0    | $\Delta = 55.7^\circ$ Az = 309.6° (USCGS)                                   |
|     | Pmax  | C           | 24 31      |                                                                             |
|     | ePP2  | C           | 26 27.5    | PV1 A 1.2s 71.8nm MPV1(A)=5.6                                               |
|     | ePP3  | C           | 26 34      | PV2 A 1.8s 1120nm MPV2(A)=6.6                                               |
|     | eS2   | C           | 32 04      | PV3 A 1.6s 1575nm MPV3(A)=6.8                                               |
|     | i S3  | C           | 32 11      | PV3 C 8s 16.3 $\mu$ m MPV3(C)=7.1                                           |
|     | eiSS  | C           | 35 56      | SH3 C 18s 60 $\mu$ m MSH3(C)=7.3                                            |
|     | iSSS  | C           | 36 44      | LmV C 17s 610 $\mu$ m MLV(C) =7.8                                           |
|     | LmH   | C           | 48.0       | Note: P has a period of about 23s in the long-period seismograph of type B! |

**Note** in Tab. 3.1 the distinct differences between the different types of magnitude and the clear underestimation of short-period (type A  $\rightarrow$  mb) magnitudes. This early practice of specifying magnitude annotation has been officially recommended already by the IASPEI Sub-Committee on Magnitudes in 1977 (see Willmore, 1979) but has, regrettably, not become global standard. However, current deliberations in IASPEI stress again the need for more specific magnitude measurements and reports to databases along these lines (see IS 3.2) if the measured data deviate more than 0.1 m.u. from the ones derived for the same type of magnitude calculated by adherence to the newly recommended IASPEI measurement standards. The latter are explained in detail in IS 3.3 of this Manual. Some essential deviations from the above outlined recommendations of the IASPEI Canberra meeting in 1979 are presented in the relevant subsections below with respect to the specific type of magnitude. Determining magnitudes according to more modern and physically based concepts, such as radiated energy or seismic moment, requires special procedures (see IS 3.6 and 3.8-3.10).

Global or regional data analysis centers calculate mean magnitudes on the basis of many A, T and/or M data reported by seismic stations from different distances and azimuths with respect to the source. This will more or less average out the influence of regional, source and local station conditions. Therefore, A, A/T or M data reported by individual stations to such centers should not yet be corrected for  $C_r$  and  $C_s$ . These corrections can be determined best by network centers themselves when comparing the uncorrected data from many stations (e.g., Hutton and Boore, 1987). They may then use such corrections for reducing the scatter of individual readings and thus improve the average estimate.

When determining new calibration functions for the local magnitude  $M_L$ , station corrections have to be applied before the final data fit in order to reduce the influence of systematic biases on the data scatter. According to the procedure proposed by Richter (1958) these station corrections for  $M_L$  are sometimes determined independently for readings in the N-S and E-W components (e.g., Hutton and Boore, 1987). When calculating network magnitudes some centers prefer the median value of individual station reports of  $M_L$  as the best network estimate. As compared to the arithmetic mean it minimizes the influence of widely diverging individual station estimates due to outliers or wrong readings (Hutton and Jones, 1993).



**Fig. 3.20** Relative magnification curves for ground displacement for various types of standardized classical seismographs. **Left:** as a function of period according to Willmore (1979) and **right:** as a function of frequency as in Chapter 11.

**WA** = Wood-Anderson seismograph, which was instrumental in the definition of the local magnitude scale;

**C** = KIRNOS – short-to-medium-/long-period Russian broadband seismograph operated at all first-rate stations of the former Soviet Union and its allied countries as well as in China, which was instrumental in developing the Moscow-Prague  $M_s$  magnitude calibration function applicable in a wide range of surface-wave periods;

**A4** = Kirnos SKM short-period seismograph of relatively broad bandwidth;

**A2** = WWSSN-SP = short-period and

**B1** = WWSSN-LP = longperiod seismograph of the World Wide Standardized Seismograph Network (WWSSN), set up by the United States Geological Survey (USGS) in the 1960s and 1970s, which were instrumental for measuring short-period P-wave magnitude  $m_b$  and the long-period  $M_s$  around 20 s periods;

**B3** = US High Gain Long-Period seismograph with maximum magnification around 40 s;

**SRO-LP** – long-period high-gain seismograph of the U.S. Seismological Research Observatories (SRO) for measuring 20 s surface-wave amplitudes and magnitudes with best signal-to-noise ratio by strongly filtering out 6 s ocean microseisms.

### 3.2.3.2 Aim and specified general procedures for IASPEI standard magnitudes and possible modifications in automated procedures (P. Bormann, J. Saul, and S. Wendt)

IASPEI has established in 2002 within its Commission on Seismic Observation and Interpretation (CoSOI) a Working Group on Magnitude Measurements (in the following for short named the WG Magnitude) in an effort to assure

- homogeneity and long-term global compatibility of magnitude data and their usefulness for seismic hazard assessment and research and therefore
- best possible agreement with magnitudes of the same type that have been measured for decades from analog seismograms according to original definitions, as well as
- unambiguity in the nomenclature of magnitude data of similar type but measured with different procedures

aiming at

- minimizing bulletin magnitude biases that result from procedure-dependent single-station or network magnitude biases,
- increasing the number of seismological stations and networks with well-defined procedures and
- promoting the best possible use of the advantages of digital data and processing.

As of now, this WG has agreed on standard measurement procedures for five widely used magnitudes: ML, mb, mb(Lg), Ms(20), mB(BB) and Ms(BB). For details see the latest version IASPEI (2013) of the recommendations of the Working Group of Magnitude Measurements as well as a more elaborated version with comments in IS 3.3.

Of general importance is that these standard magnitudes have to be measured on either simulated records of classical analog seismographs or on unfiltered digital broadband records:

- ML on Wood-Anderson (WA) records;
- mb and mb-Lg on WWSSN-SP records;
- Ms(20) on WWSSN-SP records.
- mB(BB) and Ms(BB) on unfiltered broadband records with a velocity-proportional response over at least the period range in which mB(BB) ( $T = 0.2 \text{ s}$  to  $30 \text{ s}$ ) and Ms(BB) ( $T = 3 \text{ s}$  to  $60 \text{ s}$ ) should be measured.

For the approximate shape of the response curves of WA, WWSSN-SP and –LP see Fig. 3.20 above, for the more precise normalized response curves and their respective poles and zeros see Figure 1 and Table 1 in IS 3.3.

Note that for the international data exchange and archiving the nomenclature of the standard magnitudes should be written in the (for magnitude data maximum of) 5-character IASPEI Standard Format (ISF; see Chapter 10), which would be ML, mb, mb\_Lg, Ms\_20, mB\_BB and Ms\_BB.

In contrast to the classical magnitude formulas, which are based on displacement amplitudes  $A$  measured in units of  $\mu\text{m}$ , amplitude readings for standard magnitudes should all be reported

in units of nm (for A) or in nm/s (for velocity amplitudes V measured on unfiltered BB records). Therefore, the commonly known classical calibration relationships have been modified for the standard magnitudes to be consistent with these new units.

The measured amplitudes have to be reported following their so-called **amplitude phase names**, which are specified for the respective magnitudes as: IAML, IAmb, IAmb\_Lg, IAMs\_20, IVmB\_BB and IVMs\_BB. Tab. 10.5 in Chapter 10 gives an example how these amplitude data, also for other non-standard magnitudes, appear now in IMS1.0/ISF parameter format in the event and station parameter plots produced by the USGS/NEIC HYDRA automatic location and analysis system.

The current recommendations for amplitude and period measurements for the new **IASPEI magnitude standards** orient themselves essentially on what has been described above with Fig. 3.18 and is still dominating practice when analyzing analog records or digital screen plots. According to IS 3.3 these recommendations read as follows:

*“The amplitudes used in the magnitude formulas ... are in most circumstances to be measured as one-half the maximum peak-to-adjacent-trough (sometimes called “peak-to-peak”) deflection of the seismogram trace. None of the magnitude formulas presented in this article are intended to be used with the full peak-to-trough deflection as the amplitude. The periods normally are to be measured as twice the time-intervals separating the peak and adjacent-trough from which the amplitudes are measured. The amplitude-phase arrival-times are normally to be measured as the time of the zero-crossing between the peak and adjacent-trough from which the amplitudes are measured.”*

Although the calculation of the broadband magnitudes mB\_BB and Ms\_BB, now based on direct measurement of  $V_{\max}$ , do no longer require that the periods are measured, the WG Magnitude strongly encourages their complementary measurement and reporting, since the period at which  $V_{\max}$  appears in broadband records is closely related to the corner frequency of the radiated source spectrum and thus is of interest also for other users of these data.

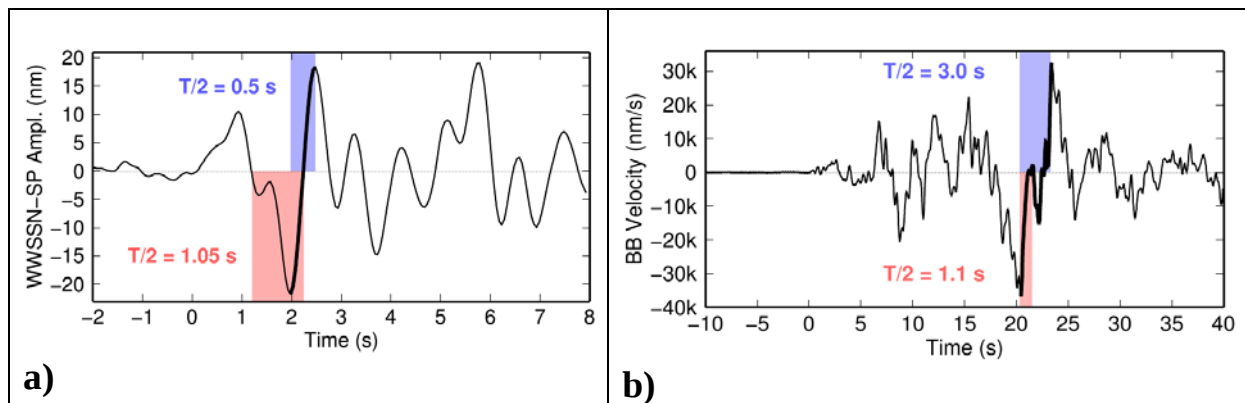
Furthermore, **we have to accept that their might be agency-specific circumstances that may lead to the modification of standard procedures, e.g., of measuring amplitudes and periods or using modified types of responses. But then certain rules should be obeyed (see sections 3.5 and 5 in IS 3.3) and a careful documentation of station/agency magnitude procedures be assured (see questionnaire in Annex 2 to IS 3.4).** The latter should be posted on the station/agency website and also be deposited and kept updated at the ISC for the information of data users.

While the visual recognition and picking of the dominating amplitudes and related periods is relatively easy interactively for an experienced analyst on a computer screen or record plot, even in the case of rather complex superimposed waveforms as in Fig. 3.17, or 3.21b below, it is much more difficult for a definite algorithm to handle properly all possible ramifications of such a measurement problem. Agencies which run automatic procedures of data analysis, may therefore prefer to measure as amplitude just the maximum positive or negative deflection between two zero crossings of the record trace and to estimate the zero position from the long-term average prior to the respective wave onset. The related period is then taken, e.g., as twice the timedifference between the respective zero crossings. This may be easier to handle automatically. However, in order to check whether such a modified automatic measurement procedure, but also an automatic procedure which is supposed to mimic best the

recommended measurement standards, yield indeed magnitude estimates within  $\pm 0.1$  m.u. of standard magnitudes measured manually or interactively by an expert analyst, this has to be tested and documented by comparative measurements. They should use an identical representative test data set in a wide range of magnitudes or  $\log(A/T)$  and  $\log(V/2\pi)$  values, respectively. Only when the results agree with an average absolute deviation less than about 0.1 m.u., the automatic data, or those derived with any other modified measurement procedure, can be accepted as being in agreement with the standards and thus can be reported to international agencies or in agency bulletins with the IASPEI amplitude-phase-name nomenclature.

We demonstrate this by way of example for period and amplitude measurements for mb, mB\_BB, Ms\_20 and Ms\_BB. The interactive analysis has been carried out by S. Wendt, CLL observatory of the University Leipzig, Germany, and the automatic analysis by J. Saul, GFZ German Research Centre for Geosciences at Potsdam. The latter tested two versions of rather simple but very fast and stable test algorithms, intended to be used with masses of globally retrieved waveform data in the tsunami early warning context. This sets limits with respect to algorithm flexibility and sophistication.

Fig. 3.21 gives an example of period measurements on a WWSSN-SP waveform for mb and on a velocity broadband waveform for mB\_BB. Fig. 3.22 summarises the median and average ratios  $T_{\text{manual}}/T_{\text{automatic}}$  and their standard deviations when measuring mb, mB\_BB, Ms\_20 and Ms\_BB.



**Fig. 3.21** P waveforms on **a)** a simulated WWSSN-SP record and **b)** on an unfiltered velocity broadband record to explain some of the possible complications of automatic amplitude and period measurements as compared to manual readings. The width of the **blue column** corresponds to  $T/2$  when measuring as amplitude half of the vertical distance between the maximum peak and the maximum adjacent trough (for short peak-to-trough = ptp; marked as **bold black record trace**) and as period twice the time difference between ptp, as recommended by the standard. In contrast, the width of the **red column** corresponds to reading  $T/2$  between the two adjacent zero crossings related to the largest (here negative) swing, when measuring as amplitude the largest zero-to-peak (or trough) deflection, in the following termed ztp for short. For comments on Fig. 3.21 see text.

When looking at the two waveforms in Fig. 3.21 one realizes that for an analyst there is no difficulty at all to properly determine  $A$  and  $T$  according to the standard recommendation based on ptp readings, even on the broadband trace with a distinct superposition of various

periods. The dominant one, related to the absolute  $V_{\max}$ , is without any doubt. Yet, in the BB velocity record of Fig. 3.21b T/2 manual peak-to-trough (= ptp in blue) is about 3 times T/2 ptp automatic (in red), because the simple algorithm got trapped in a secondary adjacent peak just after the zero crossing from below. A more sophisticated algorithm, which would recognize the difference between a secondary adjacent and the **maximum** adjacent peak would surely have got about the same period as an analyst.

Luckily, differences in T at which  $V_{\max}$  is measured do not matter when calculating mB\_BB. Only asymmetry between the corresponding largest positive and negative half cycle amplitudes makes a difference when comparing ptp/2 with zero-to-peak (ztp) amplitude measurements. In our case, however, the difference between **manual** ptp/2 and automatic ztp is minimal (**0.02 m.u.**). In contrast, the difference between the tested **automatic** ptp/2 and automatic ztp results is nearly **0.3 m.u.**, which would be the theoretical maximum for single-sided swings. Yet, on the smoother P waveform in Fig. 3.21a the current ptp/2-algorithm had not difficulties to correctly read A and T. In contrast, the automatic ztp-algorithm measured on this waveform a much too large period between the related two zero crossings. In view of the steep WWSSN-SP response drop for periods larger than 1.5 s, a much too large A was calculated, resulting in a **0.6 m.u. larger mb value than mb(IASPEI)** (see section 3.2.2, **Problem 2**, Figs. 3.16 and 3.18 and related discussion).

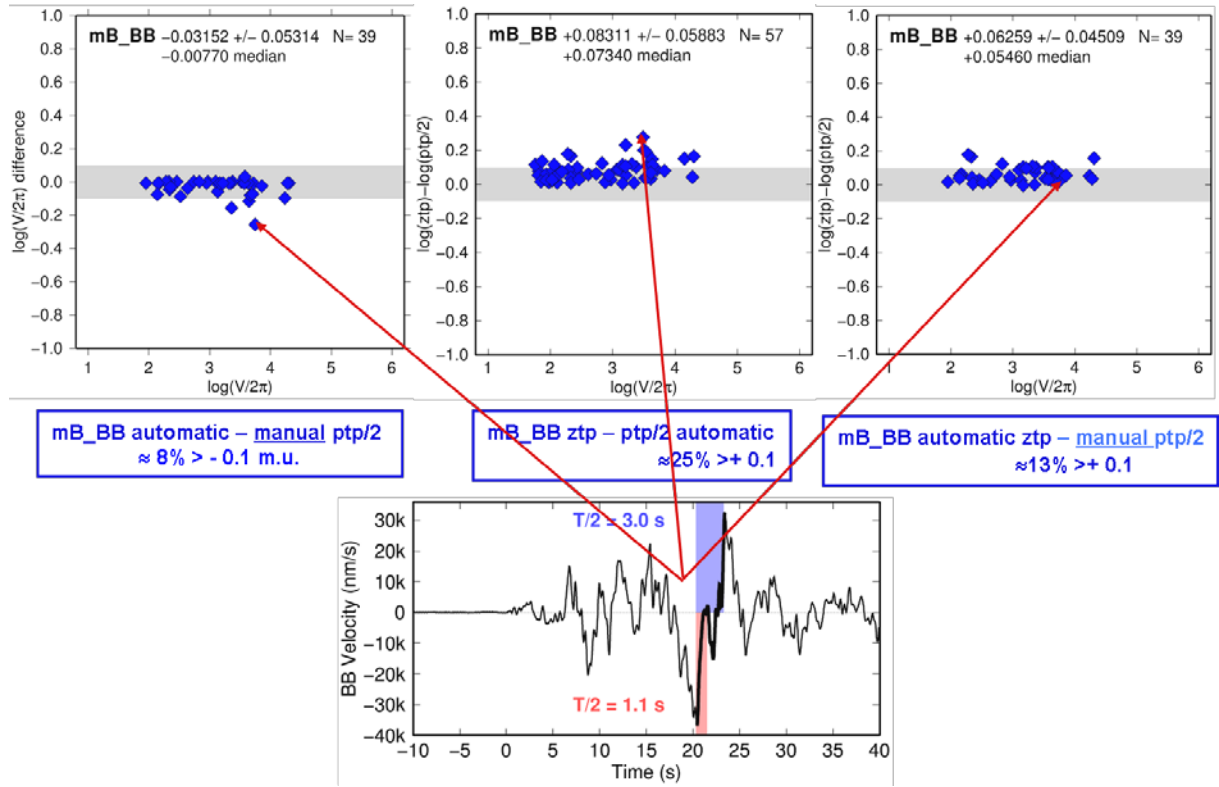
Fig. 3.22 summarizes the results of our comparative period measurements for all four teleseismic standard magnitudes. The average ratios  $T_{\text{automatic}}/T_{\text{manual}}$  are close to 1 for mb, Ms\_20 and Ms\_BB with very small standard deviation SD. Worst are the mean and SD for mB\_BB. On the other hand the median ratio is best for the two broadband magnitudes mB\_BB and Ms\_BB. This indicates that outliers in period estimates in conjunction with mB\_BB measurements, which can not be prevented with the here tested automatic ptp/2 algorithm, seem to bias both the mean and SD significantly. However, the generally good agreement between manual and automatic period determinations according to the standard recommendation is very encouraging.

| Period ratios: |                                            |                                            |
|----------------|--------------------------------------------|--------------------------------------------|
| mb:            | $T_{\text{automatic}} / T_{\text{manual}}$ | mean = $1.025 \pm 0.004$<br>median = 1.007 |
| mB_BB:         |                                            | mean = $0.850 \pm 0.023$<br>median = 1.000 |
| Ms_20:         |                                            | mean = $0.996 \pm 0.001$<br>median = 0.997 |
| Ms_BB:         |                                            | mean = $1.027 \pm 0.003$<br>median = 1.000 |

**Fig. 3.22** Period ratios  $T_{\text{automatic}}/T_{\text{manual}}$  (median, mean and standard deviation) for the teleseismic standard magnitudes mb, mB\_BB, Ms\_20 and Ms\_BB when measured according to the 2 times peak-to-trough rule.

Next we compare in Fig. 3.23 various combinations of automatic and manual mB\_BB measurements.





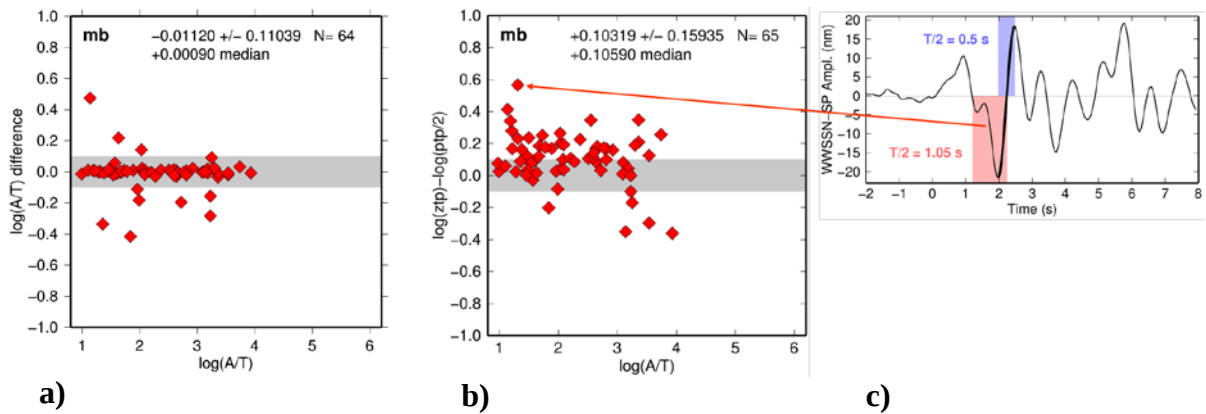
**Fig. 3.23** Differences of mB\_BB values (ordinates of the upper three figures) between automatic and manual ptp/2 procedures (upper left), fully automatic ztp and ptp/2 procedure (upper middle), and automatic ztp compared with manual ptp/2 procedure (upper right). The analyzed records cover a wide range of earthquakes with magnitudes  $5.3 \leq mB\_BB \leq 8.3$ . In the middle row the percentage of differences  $> \pm 0.1$  m.u. is given. The long red arrows point to the respective magnitude difference values in the uppermost three diagrams that result from applying the respective procedure combination to the waveform depicted in the bottom diagram.

From Fig. 3.23 the following conclusions can be drawn:

- Different algorithms of automatic mB\_BB determination may yield results which significantly differ from each other or from competent manual mB\_BB measurements.
- Best agreement between automatic and manual mB\_BB calculations is achieved with the current ptp/2 test algorithm. The single outlier of nearly  $-0.3$  m.u. is due to the algorithm getting stuck in a secondary peak near to the zero line. Only in two more cases the difference between automatic and interactive readings is slightly larger than  $0.1$  m.u., probably again due to taking some secondary maxima/minima as largest adjacent peak or trough, yet further away from the zero line. In summary, it is most likely, that the tested automatic ptp/2 procedure yields overwhelmingly standard compatible mB\_BB data. The median difference in our test with 39 events is  $+0.01$  m.u. and the average absolute difference  $0.03$  m.u., reducing to  $0.02$  m.u. without the single large outlier.
- If the algorithm could be upgraded so as to avoid getting trapped in secondary extrema of different polarity adjacent to the primary amplitude maxima or minima then the automatic mB\_BB measurements would be perfectly compatible with manual standard measurements.

- The agreement between automatic ztp and manual ptp/2 measurements is somewhat inferior. In about 13% of the studied cases residuals are  $> 0.1$  m.u., but always  $< 0.2$  m.u. with the average absolute deviation 0.04 m.u.
- The disagreement is largest between fully automatic ztp and ptp/2 derived mB\_BB values. In about 25% of the cases the difference is  $> 0.1$  m.u. and may reach about 0.3 m.u. with a median of 0.07 and an average absolute deviation of almost 0.05 m.u. This is not compatible with IASPEI standard requirements.

Fig. 3.24 shows a similar comparison of automatic and manual mb procedures.



**Fig. 3.24** Difference in m.u. between automatically and manually determined mb values in the magnitude range  $5.0 \leq mb \leq 7.6$ : **a)** automatic – manual ptp/2 measurement; **b)** automatic ztp – manual ptp/2 measurement; **c)** WWSSN-SP waveform with the ztp negative half swing marked in red from which A and T were measured that produced the data point with the largest difference of almost 0.6 m.u. in diagram b).

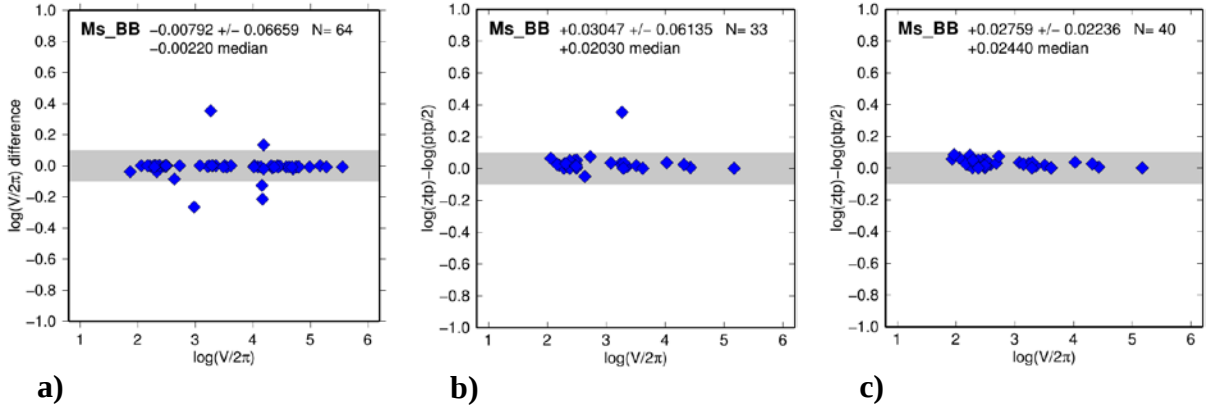
From Fig. 3.24 the following conclusions can be drawn:

- mb yields more data than mB\_BB for smaller events but with much larger scatter.
- For about 84% of the analyzed waveforms, **the ptp/2 test algorithm yielded mb values that agreed within 0.1 m.u. with those determined by manual expert analysis.** Both arithmetic mean and median difference are within 0.01 m.u. from zero. Therefore, the ptp/2 algorithms hold promise to produce overwhelmingly mb values that are compatible with IASPEI standard mb. Despite sometimes large outliers the median difference is zero and the average absolute deviation between automatic and manual ptp/2 measurements for mb only 0.05 m.u.
- Yet, outliers up to 0.6 m.u. are not acceptable for standard magnitude data that aim at drastically reducing procedure-dependent errors. Therefore, for assuring IASPEI compatible amplitude, period and mb values in the agency bulletins and data reports to international data centers either the algorithm has to be improved or expert control of the final data is indispensable.
- The negative outliers of mb are again due to the ptp/2 algorithm getting occasionally trapped in secondary adjacent maxima or minima (see above discussion for mB\_BB).
- Significant positive outliers may be due to overestimating the period T (see the 2-3s problem discussed above), usually at rather small  $\log(A/T)$  values due to superposition

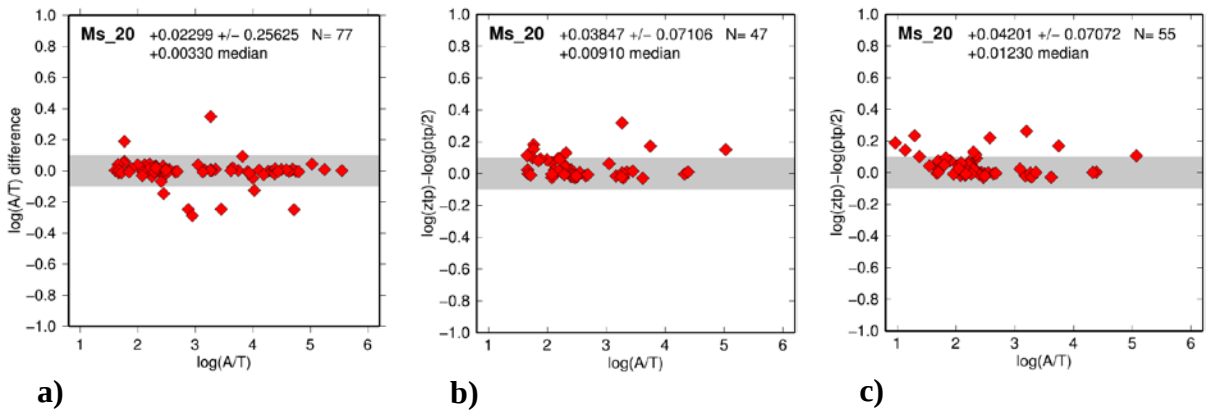
with more long-period noise amplitudes which is not recognized and properly accounted for by the automatic algorithm.

- **The ztp-algorithm**, which still worked reasonably well for mB\_BB, **does not yield standard compatible results for mb**. Average and median difference are already about 0.1 m.u., the standard deviation is  $\pm 0.16$  m.u. and the average absolute difference to ptp/2-based manual analysis 0.11 m.u. Accordingly, for about  $\frac{3}{4}$  of the analyzed waveforms the difference to manual ptp/2 measurement was larger than 0.1 m.u., with positive and negative outliers up to about 0.4-0.6 m.u.

Figs. 3.25 and 3.26 show such comparisons also for broadband Ms\_BB and narrowband Ms\_20.



**Fig. 3.25** Difference in m.u. between automatically and manually determined velocity broadband Ms\_BB values in the magnitude range  $4.6 \leq \text{Ms\_BB} \leq 9.0$ : **a)** automatic – manual ptp/2 measurement; **b)** automatic ztp – manual ptp/2 measurement; **c)** automatic ztp – ptp/2 measurement.



**Fig. 3.26** Difference in m.u. between automatically and manually determined velocity broadband Ms\_20 values in the magnitude range  $4.8 \leq \text{Ms\_20} \leq 9.0$ : **a)** automatic – manual ptp/2 measurement; **b)** automatic ztp – manual ptp/2 measurement; **c)** automatic ztp – ptp/2 measurement.

Comments and conclusions related to Fig. 3.25 and 3.26 are the following:

- For both  $M_s$  magnitudes automatic and manual A/T and  $V_{\max}$  readings based on ptp/2 yield in most cases comparable values, yet the average agreement is clearly better and the data scatter less for  $M_s_{\text{BB}}$  (mean difference -0.01 m.u., SD =  $\pm 0.07$  m.u. and average absolute deviation of 0.02 m.u. only, as compared to  $M_s_{20}$  with -0.02,  $\pm 0.25$  and 0.07 m.u. respectively).
- For only about 8% of the data automatic ptp/2  $M_s_{\text{BB}}$  differed more than  $\pm 0.1$  m.u. from the manually measured ones, for  $M_s_{20}$  this applied to about 10% of the data.
- **The differences between automatic ztp and manual ptp/2 are even less.** For  $M_s_{\text{BB}}$  the agreement is even perfect, with the exception of a single but relatively large (+0.35 m.u.) outlier, which is in fact the same for  $M_s_{20}$ .
- The reason for this single large outlier turned out to be an event association error by the automatic algorithm. It searches for the maximum surface-wave amplitude within a given time window after the P-wave onset. The time window depends on distance and a fixed group-velocity window assumed for the surface waves. In the current algorithm, the group velocity window covers the range between 2.5 and 4.5 km/s. With this wide window, applied to two Chile earthquake aftershocks, following each other within 17 min, the algorithm measured as the largest surface-wave amplitude for both  $M_s_{\text{BB}}$  and  $M_s_{20}$  that of the about 0.3 m.u. larger later aftershock, the analyst, however, the surface-wave maxima measured by the analyst, however, related correctly to the earlier aftershock. This wrong association accounts also for the identically large positive outlier when comparing the automatic and manual ptp/2 results for  $M_s_{\text{BB}}$  and  $M_s_{20}$  in Figs. 3.25 and 3.26. The lesson is, that proper association of surface-wave maxima related to closely following events are generally more difficult for an algorithm which has to operate blind with a fixed set of parameters than for an off-line working intelligent analyst. Therefore, before final data cataloguing and reporting to international data centers automatically calculated magnitude data should go through an expert review.
- While there are no negative residuals at all for  $M_s_{\text{BB}}$  when comparing automatic ztp with manual ptp/2, there are 3 negative ones when comparing automatic and manual ptp/2, and even 6 in the case of  $M_s_{20}$ . One of the negative outliers for both  $M_s_{\text{BB}}$  (-0.27 m.u.) and  $M_s_{20}$  (-0.25 m.u.) was again due to the wrong association of the automatically measured surface-wave maximum to another event that occurred within the same hour but at rather different distance and thus arrived at the recording station again close to the surface-wave maximum of another event. And one more of the 3 negative  $M_s_{\text{BB}}$  residuals (-0.22 m.u.) was caused by missing the later arriving absolute  $V_{\max}$  for  $M_s_{\text{BB}}$  within the preset group velocity and measurement time window, which, however, worked fine for this event for the earlier arriving  $M_s_{20}$  amplitude maximum..
- **Only for  $M_s_{\text{BB}}$  the tested automatic ztp and ptp/2 algorithms yielded practically identical results within  $< 0.1$  m.u.**

Tab. 3.2 summarizes the essential statistical results of the above procedure comparison.

**Tab. 3.2** Statistic parameters of comparing automatic magnitude calculations based on ztp and ptp/2 amplitude measurements with manual ptp/2 measurements according to IASPEI standard procedures. **Mauto** = type of magnitude, measured automatically, **N** = number of measured differences between automatic and manual magnitude determinations; **Median** = median difference, **Mean** = arithmetic mean of differences, **SD** = standard deviation from the mean, **Avad** = average absolute difference, **% > 0.1** = approximate percentage of differences larger than 0.1 m.u., **Max outliers**. **Note:** all calculated median, mean, SD and avad values have been rounded to the nearest second decimal. Since the distribution of differences is not Gaussian the median and avad values are more relevant to assess the agreement with the strict standard procedure. Mauto procedures in best agreement with the manual standard procedure have been written in bold letters.

| <b>Mauto</b>        | <b>N</b> | <b>Median<br/>m.u.</b> | <b>Avad<br/>m.u.</b> | <b>Mean<br/>m.u.</b> | <b>SD<br/>m.u.</b> | <b>%<br/>&gt; 0.1.u.m</b> | <b>Max outliers<br/>m.u.</b> |
|---------------------|----------|------------------------|----------------------|----------------------|--------------------|---------------------------|------------------------------|
| mb(ztp)             | 65       | +0.10                  | 0.11                 | +0.10                | ±0.16              | 55                        | +0.6 and -0.4                |
| <b>mb(ptp/2)</b>    | 64       | 0.00                   | 0.05                 | -0.01                | ±0.11              | 18                        | +0.5 and -0.5                |
| mB_BB(ztp)          | 39       | +0.05                  | 0.04                 | +0.06                | ±0.05              | 13                        | +0.3                         |
| <b>mB_BB(ptp/2)</b> | 39       | -0.01                  | 0.03                 | -0.03                | ±0.05              | 8                         | -0.15                        |
| Ms_20 (ztp)         | 47       | +0.01                  | 0.05                 | +0.04                | ±0.07              | 18                        | +0.3                         |
| Ms_20(ptp/2)        | 77       | 0.00                   | 0.07                 | +0.02                | <b>±0.26</b>       | 10                        | +0.4 and -0.3                |
| <b>Ms_BB(ztp)</b>   | 33       | +0.02                  | 0.02                 | +0.03                | ±0.06              | 3                         | +0.4                         |
| <b>Ms_BB(ptp/2)</b> | 64       | 0.00                   | 0.03                 | -0.01                | ±0.07              | 8                         | +0.4 and -0.3                |

Summarizing our procedure comparison test we conclude:

- The current automatic GFZ algorithm for determining teleseismic magnitudes of the types mb, mB\_BB, Ms\_20 and Ms\_BB have been developed in the first instance for application in the tsunami early warning systems, i.e., for very fast and stable applications to great amounts of globally accessible waveform data. Therefore, ztp has so far been given preference for ptp/2 in operational applications.
- Yet, our investigations showed, that the ztp algorithm is – with the exception for Ms\_BB – inferior (often not much but worst for mb) to the ptp/2 algorithm, which is closer to the preferred IASPEI standard procedure.
- However, with the exception for mb, both algorithms yield (when taking into account the median, respectively the mean differences, standard deviations, and average absolute differences as well as the real distribution of residuals) magnitude estimates which differ in less than 20%, for the ptp/2 version even in less than 10% of the cases, magnitude values that are within 0.1 m.u of manual expert estimates.
- The differences could be further reduced, for body-wave magnitude estimates in particular, by using a more sophisticated ptp/2 algorithm that avoids getting trapped in a secondary maximum or minimum by searching for the **largest adjacent** amplitude with opposite polarity to the primary maximum deflection from the zero line..

- Automatic algorithms for surface waves have to search for the maximum amplitude within a flexible time window, depending on distance and a preset group velocity window. If the latter is chosen too small it may fail or yield to small values when applied to Ms\_BB measurements in a wide range of periods and distances. And when the group velocity window is chosen too wide, the algorithm may measure instead of the searched for surface-wave maximum of an earlier and/or weaker event the later following larger maximum of an overlapping event, closely spaced in time of arrival. Our data have been affected by such cases.
- Event association errors, and likely also mistakes in proper peak-to-trough association in the case of complex waveforms, can hardly be excluded for sure in automatic procedures of magnitude determination that are based on preset algorithm parameters. Therefore, automatically determined magnitude data should be critically checked by expert analysts before they enter final bulletins or are reported to international data centers for final use and archiving.
- Generally, in any such comparison of different magnitude measurement procedures, differences beyond the acceptable level of measurement errors of about 0.10-0.15 m.u. should not be considered as random noise. Usually they have a non-random cause. Therefore, one should strive to identify their reasons. This requires, however, that one goes back for the specific data points to their respective waveforms and compare which amplitudes and periods have been measured by what method and at which time. If the reasons for significant residuals have become clear this may also offers the chance to reduce the data scatter by modifying the algorithm. This we have tried to demonstrate by way of example.
- Another lesson from our demonstration is that one should avoid in such comparisons that the selected reference waveform data contain overlapping wave groups belonging to different events in order to avoid large outliers due to wrong associations.

In the following we will outline the origin, general features, formulae and specific differences of various magnitude scales that are currently in use or recommended by the IASPEI WG on Magnitudes. We will highlight which of these scales are at present already accepted as world-wide standards and will also spell out related problems which still require further consideration, clarifying discussion, recommendations or decisions by CoSOI. Data tables and diagrams on calibration functions used in actual magnitude determinations are given in Datasheet 3.1.

### 3.2.4 Magnitude scales for local and regional events

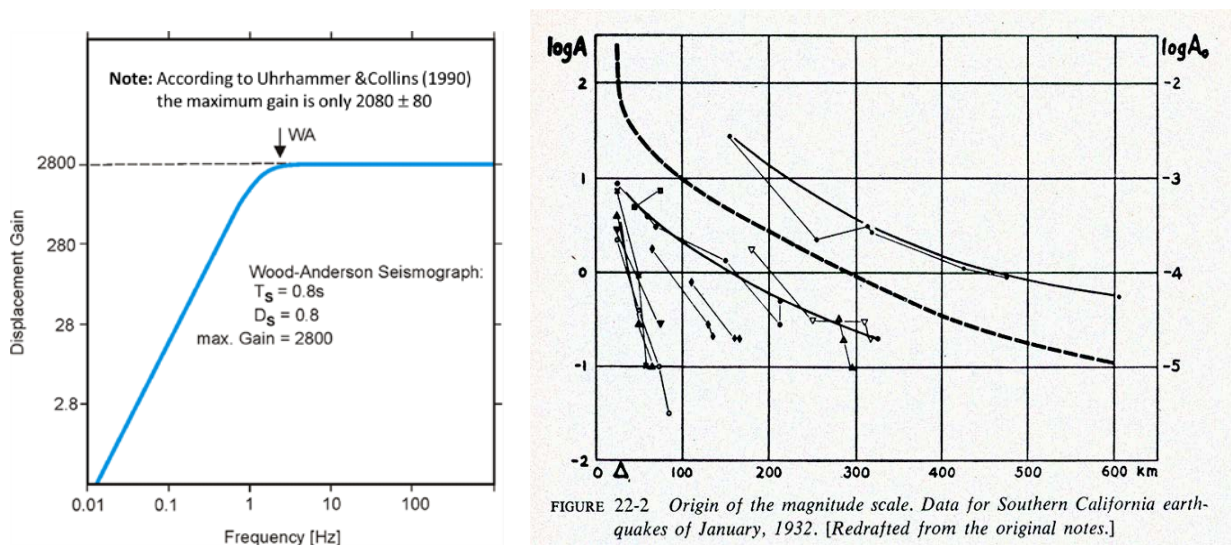
The large variability of velocity and attenuation structure of the crust does in fact not permit the development of a unique, internationally standardized calibration function for local events. However, the original definition of magnitude by Richter (1935) did lead to the development of a local magnitude scale  $M_L$  (original nomenclature  $ML$ ) for California.  $M_L$  scales for other areas are usually scaled to Richter's or Hutton and Boore's (1987) definition and also the procedure of measurement is more or less standardized. With slight modification, the Hutton and Boore formula has now been accepted by IASPEI as the reference formula for standard  $ML$  to which other regional  $M_L$  scales should be scaled in order to yield compatible results within about 0.1 m.u. for equal measurement values at different distances.

### 3.2.4.1 The original Richter magnitude scale $M_L$

In his 1935 landmark paper Richter stated:

*„...it is frequently desirable **to have a scale** for rating these shocks in terms of their original energy, independently of the effects which may be produced at any particular point of observation.“*

For his measurements he used Wood-Anderson (WA) horizontal component torsion seismometers which had become available and widely deployed as a standard instrument in the 1930s in Southern California. According to the manufacturer the **WA seismometers** had the following parameters: natural period  $T_S = 0.8$  s, damping factor  $D_S = 0.8$ , **static magnification**  $V_{\max} = 2800$ . The resulting response is sketched in Fig. 3.27 left..



**Fig. 3.27 Left:** Wood-Anderson response according to manufacturer's instrument data; **Right:** Plot of original data measured by Richter; copy of Figure 22-2 in Richter (1958).

Following a recommendation by Wadati, Richter plotted the logarithm of maximum **trace amplitudes**,  $A_{\max}$ , measured on WA records as a function of epicentral distance  $\Delta$ . He found that  $\log A_{\max}$  decreased with distance along more or less parallel curves for earthquakes of different size (Fig. 3.20 (right)). This led him to propose the following definition for the magnitude as a quantitative measure of earthquake size (Richter 1935, p. 7):

*"The magnitude of any shock is taken as the logarithm of the maximum trace amplitude, expressed in microns, with which the standard short-period torsion seismometer ... would register that shock at an epicentral distance of 100 km".*

This local magnitude was later given the symbol  $M_L$  (Gutenberg and Richter, 1956b). In order to calculate  $M_L$  also for other epicentral distances,  $\Delta$ , between 30 and 600 km, Richter (1935) provided attenuation corrections. They were later complemented by attenuation corrections for  $\Delta < 30$  km assuming a focal depth  $h$  of 18 km (Gutenberg and Richter, 1942; Hutton and Boore, 1987). Accordingly, one gets

$$M_L = \log A_{\max} - \log A_0 \quad (3.14)$$

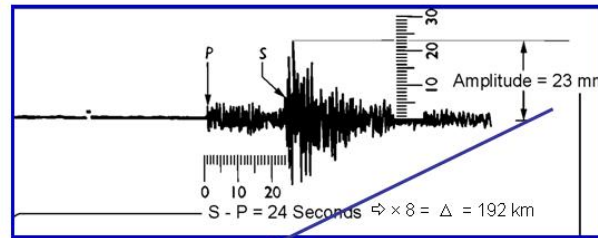


with  $A_{\max}$  in mm of measured zero-to-peak trace amplitude in a Wood-Anderson seismogram. The respective calibration values  $-\log A_0$  were published in tabulated form by Richter (1958) (see Table 1 in DS 3.1). Fig. 3.28 summarizes the classical procedure of ML determination.

According to this original definition of the local magnitude scale  $ML = ML$  by Richter (1935), only the *maximum trace amplitude*  $B$  in mm as recorded in standard records of a Wood-Anderson seismometer is measured, i.e.:

$$ML = \log B(WA) - \log A_0(\Delta).$$

⇒ no period  $T$  is measured, and **NO CONVERSION** to “ground motion” amplitude  $A$  by correcting for the frequency-dependent seismograph magnification is made.



$$ML = \log B(WA) - \log A_0(\Delta)$$

$$ML = 1.36 + 3.46 = 4.82$$

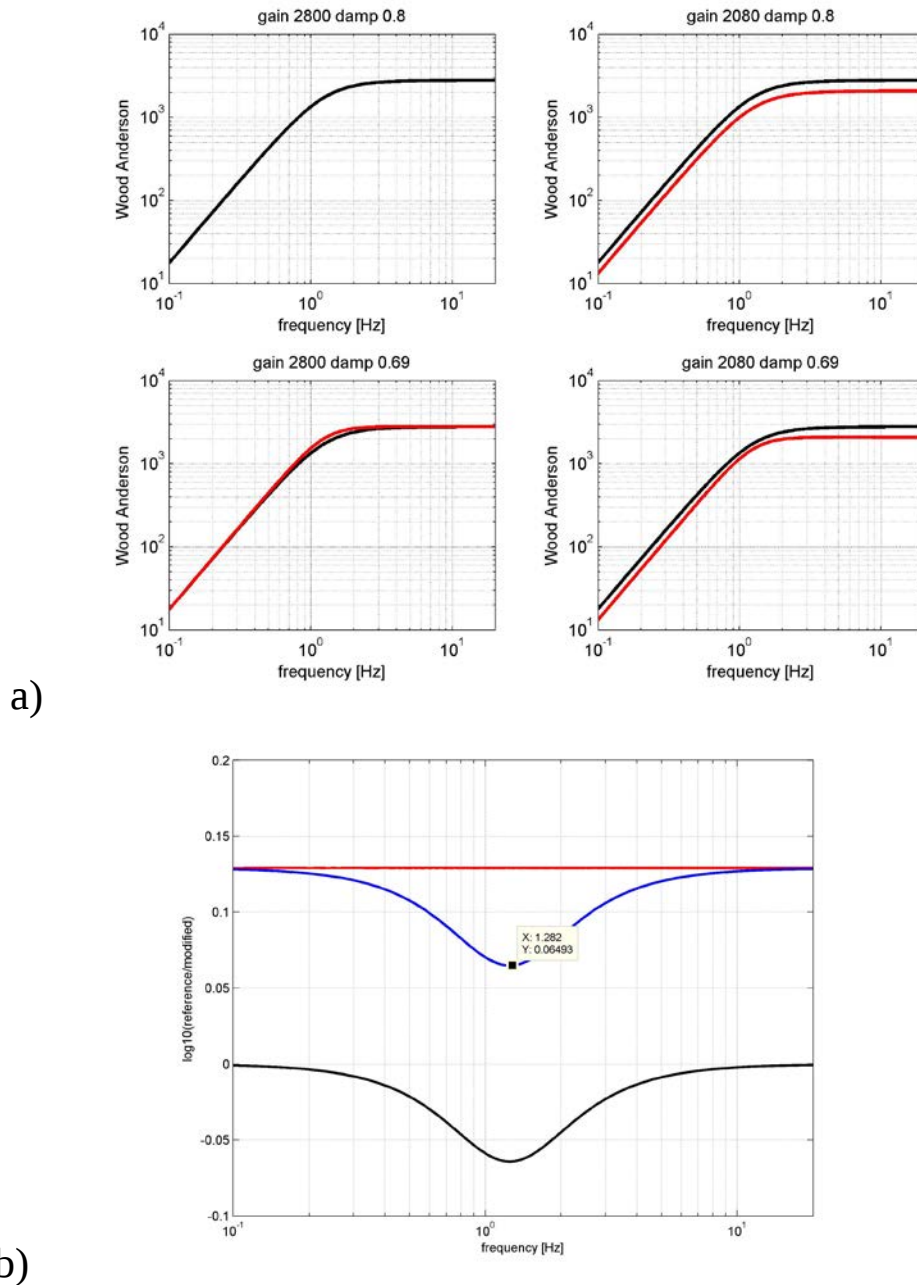
Calibration function  $\sigma_1(\Delta) = -\log A_0$  for local magnitudes  $ML$  according to Richter (1958).  $A_0$  are the trace amplitudes in mm recorded by a Wood-Anderson Standard Torsion Seismometer from an earthquake of  $ML = 0$ .  $\Delta$  - epicentral distance in km.

| $\Delta$ (km) | $\sigma_1(\Delta)$ | $\Delta$ (km) | $\sigma_1(\Delta)$ | $\Delta$ (km) | $\sigma_1(\Delta)$ | $\Delta$ (km) | $\sigma_1(\Delta)$ |
|---------------|--------------------|---------------|--------------------|---------------|--------------------|---------------|--------------------|
| 0             | 1.4                | 90            | 3.0                | 260           | 3.8                | 440           | 4.6                |
| 10            | 1.5                | 100           | 3.0                | 280           | 3.9                | 460           | 4.6                |
| 20            | 1.7                | 120           | 3.1                | 300           | 4.0                | 480           | 4.7                |
| 30            | 2.1                | 140           | 3.2                | 320           | 4.1                | 500           | 4.7                |
| 40            | 2.4                | 160           | 3.3                | 340           | 4.2                | 520           | 4.8                |
| 50            | 2.6                | 180           | 3.4                | 360           | 4.3                | 540           | 4.8                |
| 60            | 2.8                | 200           | 3.5                | 380           | 4.4                | 560           | 4.9                |
| 70            | 2.8                | 220           | 3.65               | 400           | 4.5                | 580           | 4.9                |
| 80            | 2.9                | 240           | 3.7                | 420           | 4.5                | 600           | 4.9                |

**Fig. 3.28** Summary of the classical procedure of measuring  $ML$ . Note that the letter  $B$  was given here to record trace amplitudes, highlighting that they were not been converted to ground motion displacement amplitudes  $A$ .

**Note 1:** According to Uhrhammer and Collins (1990) the magnification of 2800 of WA seismometers had been calculated on the basis of wrong assumptions on the suspension geometry. **More correct values** for the static magnification and the damping of the seismometer are  $2080 \pm 60$  and  $D_s = 0.7$ , respectively (see also Uhrhammer et al., 1996). Accordingly, **magnitude estimates based on synthesized WA records** assuming the original WA parameters systematically underestimate the size of the event. This underestimation depends on the frequency at which the amplitude is measured. It is on average at typical periods about 0.1 m.u., has its minimum around the corner frequency of 0.128 Hz ( $T = 0.8$  s; 0.065 m.u.) and is largest (up to 0.13 m.u.) at significantly lower and higher frequencies (Fig. 3.29). However, as long as records of original WA seismographs are used this bias plays no role.





**Fig. 3.29** **a)** Response curves of Wood-Anderson seismometers calculated for different static magnification and damping. Upper left: Original manufacturer's response; upper right: **red** response curve according to Uhrhammer and Collins (1990) when correcting only for the difference in static magnification; lower right: red curve: as upper right but additionally accounting for the difference in seismometer attenuation. **b)** differences in ML (m.u.): red line – correcting only for difference in static magnification; blue curve – accounting additionally for the difference in seismometer damping; black curve – difference between red and blue. (Figure by courtesy of D. Bindi, 2011).

**Note 2:** In contrast to the general magnitude formula (3.13), Eq. (3.14) considers only the maximum displacement-proportional record amplitudes but not their periods. Reason: WA instruments are short-period and their traditional analog recorders had a limited paper speed. Proper reading of the period was difficult. It was assumed, therefore, that the maximum

amplitude phase (in case of local events generally Sg, Lg or Rg) had always about the same dominant period within the plateau range of the response. Also,  $-\log A_0$  does not consider a depth dependence of  $\sigma(\Delta, h)$ . Seismicity in southern California is generally shallow (mostly less than 15 km). Eq. (3.14) also does not give regional or station correction terms. They were already taken into account when determining  $-\log A_0$  for southern California.

**Note 3:** In the following we use as the general NMSOP nomenclature for local magnitude data  $M_l$  (with  $l$  = local), unless the respective data have been measured and calibrated for sure according to the new IASPEI  $M_l$  standard or they relate directly to the original Richter scale.

**Note 4:** The smallest events recorded in local microearthquake studies have negative values of  $M_l$  while the largest  $M_l$  is about 7, i.e., the  $M_l$  scale also suffers “saturation” (see Fig. 3.47). Despite these limitations,  $M_l$  estimates of earthquake size are relevant for earthquake engineers and risk assessment since they are closely related to earthquake damage. The main reason is that many structures have natural periods close to that of the WA seismometer (0.8s) or are within the range of its pass-band (about 0.1 - 1 s).

### Problems and conclusions:

- 1) In 3.2.3 it was said that in the case of horizontal component recordings, as a general IASPEI rule,  $A_{Hmax}$  should be the maximum vector sum amplitude measured at  $t_{max}$  in both the N and E component. Deviating from this, Richter (1958) says: “... *In using ...both horizontal components it is correct to determine magnitude independently from each and to take the mean of the two determinations. This method is preferable to combining the components vectorially, for the maximum motion need not represent the same wave on the two seismograms, and it even may occur at different times.*” In most investigations aimed at deriving local  $M_l$  scales  $A_{Hmax} = (A_N + A_E)/2$  has been used instead to calculate  $M_l$  although this is not fully identical with  $M_l = (M_{lN} + M_{lE})/2$  and might give differences in magnitude of up to about 0.1 m.u.
- 2) The Richter  $M_l$  from arithmetically averaged horizontal component amplitude readings will be smaller by at least 0.15 magnitude units as compared to  $M_l$  from  $A_{Hmax}$  vector sum. In the case of significantly different amplitudes  $A_{Nmax}$  and  $A_{Emax}$  this difference might reach even several tenths of magnitude units. However, the method of combining vectorially the N and E component amplitudes, as generally practiced in other procedures for magnitude determination from horizontal component recordings, is hardly used for  $M_l$  because of reasons of continuity in earthquake catalogs, even though it would be easy nowadays with digital data.
- 3) According to Hutton and Boore (1987) the *distance corrections developed by Richter for local earthquakes ( $\Delta < 30$  km) are incorrect*. Bakun and Joyner (1984) had come to the same conclusion for weak events recorded in Central California at distances of less than 30 km. This leads to magnitude estimates from nearby stations that are smaller than those from more distant stations. Therefore, Hutton and Boore (1987) proposed that local magnitude scales be redefined such that  $M_l = 3$  corresponds to 10 mm of motion on a WA record at 17 km hypocentral distance. This is consistent with the original definition of the scale for Southern California and will also allow a better comparison of earthquakes in regions with very different wave attenuation within the first 100 km.

- 4) According to Hutton and Jones (1993) redetermination of local magnitudes recorded by the Southern California Seismographic Network (SCSN) from 1932 to 1990 has shown that the magnitudes have not been consistently determined over that period of time. The amplitudes recorded on WA instruments were systematically overestimated prior to 1944 compared to present reading procedures, leading to a significant overestimation of ML. Moreover, the change to computerized estimation in 1975 led to slightly lower event magnitudes for the time after. These obviously not well documented, thought over and agreed upon changes contributed to an apparently higher rate of seismicity in the 1930s and 1940s than later in the catalog, which had been misinterpreted as a decrease in seismicity rate after the 1952 Mw7.5 Kern County earthquake. Rereading WA amplitudes for moderate earthquakes ( $M_l \geq 4.5$ ) for the whole time span resulted in magnitudes which proved to be Poissonian distributed with no significant change in seismicity above the 90% level.
- 5) Similar experiences with other local and global catalogues led Habermann (1995) to state: “... the heterogeneity of these catalogues makes characterizing the long-term behaviour of seismic regions extremely difficult and interpreting time-dependent changes in those regions hazardous at best.... Several proposed precursory seismicity behaviors (activation and quiescence) can be caused by simple errors in the catalogues used to identify them. ....Such mistakes have the potential to undermine the relationship between the seismological community and the public we serve.”

The points stated above highlight the urgency for standardization of measurement procedures and of careful documentation of introduced changes and their reasons. Only this can assure the unambiguous reproduction of so important event parameters such as magnitude and enable - if required – their correction for assuring long-term stability and representativeness of earthquake catalogs.

For a review of the development and use of the Richter scale see Boore (1989).

### 3.2.4.2 The new IASPEI standard ML

In order to avoid several of the above mentioned problems and inconsistencies in ML determination and to reduce related data scatter and/or level discrepancies the IASPEI/CoSOI Working Group on Magnitude Measurements recommends the following for **IASPEI standard ML** (for more details and discussions see IS 3.3):

*“For crustal earthquakes in regions with attenuative properties similar to those of Southern California, the proposed standard equation is*

$$ML = \log_{10}(A) + 1.11 \log_{10}R + 0.00189 \cdot R - 2.09, \quad (3.15)$$

*where  $A$  = maximum **trace** amplitude in **nm** that is measured on output from a **horizontal-component** instrument that is filtered so that the response of the seismograph/filter system replicates that of a **Wood-Anderson standard seismograph** but with a static magnification of 1 (see Table 1 and Figure 1 in IS 3.3) and  $R$  = **hypocentral distance in km**, typically less than 1000 km.”*

Note that equation (3.15) is an expansion of that proposed by Hutton and Boore (1987). The constant term -2.09 is based on the experimentally determined attenuation and static magnification of the Wood-Anderson instrument by Uhrhammer and Collins (1990), rather than on the respective parameters specified earlier by the manufacturer (see discussion and Fig. 3.29 in 3.2.4.1). Further, the WG states:

*“For seismographic stations containing two horizontal components, amplitudes are measured independently from each horizontal component, and each amplitude is treated as a single datum. There is no effort to measure the two observations at the same time, and there is no attempt to compute a vector average.*

*For crustal earthquakes in regions with attenuative properties that are **different** than those of coastal California, and for measuring magnitudes with vertical-component seismographs, the standard equation is of the form:*

$$ML = \log_{10}(A) + C(R) + D, \quad (3.16)$$

*where A and R are as defined in equation (here 3.15), except that A may be measured from a **vertical-component** instrument, and where C(R) and D have been **calibrated** to adjust for the different regional attenuation and to adjust for any systematic differences between amplitudes measured on horizontal seismographs and those measured on vertical seismographs.”*

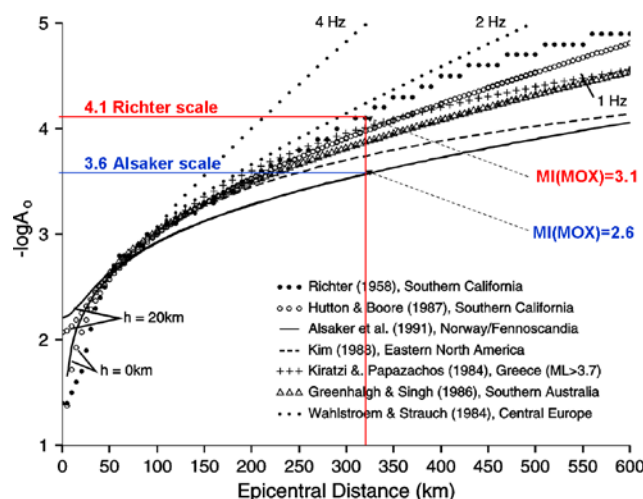
Procedures to synthesize the responses of other seismographs from digital broadband recordings are meanwhile common knowledge (e.g., Plešinger et al., 1995 and 1996). Therefore, WA seismographs are no longer required for carrying out ML determinations. Savage and Anderson (1995) and Uhrhammer, et al. (1996) demonstrated the ability to determine an unbiased measure of local magnitude from synthetic WA seismograms. Thus, a seamless catalog of ML could be maintained at Berkeley, California. In a first approximation (although not identical) this may also be achieved by converting record amplitudes from another seismograph with a displacement frequency response  $\text{Mag}(T_i)$  into respective WA trace amplitudes by multiplying them with the ratio  $\text{Mag}_{\text{WA}}(T_i)/\text{Mag}(T_i)$  for the given period of  $A_{\text{max}}$  (see EX 3.1). Sufficient time resolution of today’s high-frequency digital records is likewise no longer a problem.

**Note:** See sections 4.1 and 5.1.1 in IS 3.3 for differences between Richter’s (1935) and the now standard formula (3.15) as well as for admissible modifications of the WA response for improving the SNR and still enabling unbiased ML estimates of very weak earthquakes.

### 3.2.4.3 Other ML scales based on amplitude measurements

As highlighted by the IASPEI recommendation for an ML standard, Richter’s and also Hutton and Boore’s attenuation corrections can only be assumed valid for Southern California. More recently, a new calibration function has been agreed upon for all California and vicinity (Uhrhammer et al., 2011). But shape and level of these California calibration functions may be different from those applicable in other regions of the world with different velocity and attenuation structure, crustal age and composition, heat-flow conditions and depth distribution of earthquakes. Accordingly, when determining ML calibration functions for other regions, the amplitude attenuation law has to be determined first and then this curve has to be scaled to the

original definition of ML at 100 km epicentral distance (or at closer distance as recommended by Hutton and Boore, 1987; see Problem 3 in 3.2.4.1). Some examples for other regional ML calibration functions have been plotted in Fig. 3.30. Formulas for these and many more regional calibration functions are given in Table 2 of DS 3.1 together with comments on their scaling, components to be used and the distance range of applicability. Those for continental shield areas revealed significantly lower body-wave attenuation when compared with Southern California (e.g., the calibration curve of Alsaker et al., 1991, in Fig. 3.30. Therefore, despite proper scaling to California ML at or near to  $\Delta = 100$  km, their calibration and thus derived ML values for equal amplitude input data may already be several tenths of magnitude units smaller at several hundred km distance.



**Fig. 3.30** Calibration functions for ML determination in different regions. Some of them extend well beyond the 600 km distance plotted here, e.g., that of Alsaker et al. (1991) up to 1500 km. At the latter distance  $-\log A_0$  differs by 1.7 magnitude units from the extrapolated calibration curve for southern California. Even at much shorter distance, such as 320 km for a demonstration event recorded at station MOX, the difference is already to 0.5 m.u.

The problem of vector summing of amplitudes in horizontal component records or of arithmetic averaging of independent ML determinations in N and E components can be avoided by measuring the maximum amplitude on vertical component recordings instead, provided that the respective  $-\log A_0$  curves are properly scaled to the original definition of Richter at  $\Delta = 100$  km or at shorter distances, as proposed by Hutton and Boore (1987) and done, e.g., for Tanzania (Langston et al., 1998), for better compensation of near range differences in attenuation properties and in the depth distribution of earthquakes. Several vertical component formulas have been proposed also for other regions (see. Table 2 of DS 3.1). At large distances, sometimes well beyond the distance of 600 km for which  $-\log A_0$  was defined by Richter (1958), these formulas mostly pertain to Lg waves. Alsaker et al. (1991) and Greenhalgh and Singh (1986) showed that  $A_{Zmax} \approx 1$  to 1.2 times  $A_{Hmax} = 0.5 (A_{Nmax} + A_{Emax})$  and thus yields practically the same magnitudes.

**Note 1:** Station corrections in some of these studies varied between -0.6 to +0.3 magnitude units (Bakun and Joyner, 1984; Greenhalgh and Singh, 1986; Hutton and Jones, 1993) and correlated broadly with regional geology. This points to the urgent need to determine both calibration functions and station corrections for ML on a regional basis.

**Note 2:** Since sources in other regions may be significantly deeper than in southern California, either  $\sigma(\Delta, h)$  should be determined or at least the epicentral distance  $\Delta$  be replaced in the magnitude formulas by the "slant" or hypocentral distance  $R = \sqrt{(\Delta^2 + h^2)}$ . The latter is now common practice. E.g., the  $M_L$  scale for Kamchatka is based on the 1968 modified version of Fedotov's (1963) K-scale (see IS 3.7, Figure 5) and is defined as  $M_L = 0.5 K_S - 0.75$ . It uses the maximum ratio  $A/T$  for S-waves scaled against the S-P traveltime (which is approximately proportional to  $R$ ). The calibration curve  $\sigma(S-P)$  for the K-scale is the averaged one over three depth bins in the depth range 0-200 km.  $M_L$  defined in this way is well correlated en masse with event  $m_b$  (A. Gusev, personal communication 2013).

**Note 3:** There have been efforts to develop frequency-dependent regional calibration functions, e.g., for Central Europe (Wahlström and Strauch, 1984; see Fig. 3.30). But this breaks with the required continuity of procedures and complicates the calibration relationship for  $M_L$ .

**Note 4:** Increasing availability of strong-motion records and their advantage of not being clipped even by very strong nearby events have led to the development of (partially) frequency-dependent  $M_L^{SM}$  scales for strong-motion data (Lee et al., 1990; Hatzidimitriou et al., 1993). The technique to calculate synthetic Wood-Anderson seismograms from strong-motion accelerograms was first introduced by Kanamori and Jennings (1978). In any event, the best way to make  $M_L^{SM}$  values best compatible with standard  $M_L$ , the broadband strong-motion records have to be pre-processed by applying double integration so as to produce the respective broadband displacement equivalent (with the uncertainty of the unknown integration constants). Then a WA simulation filter should be applied (see section 3.1 in IS 3.3 and for a general application example Margaritis and Papazachos, 1999) before measuring the amplitudes. And, most importantly, a local/regional displacement calibration function carefully scaled to that in (3.15) has to be available.

#### 3.2.4.4 $m_b$ \_Lg

Short-period Lg waves with periods  $T < 3$  s (see Chapter 2, 2.3.3 and definition in IS 2.1) die out even along short oceanic travel path but they may travel far with comparably low attenuation (e.g., in comparison with Sg) through the continental crust. This applies especially to travel paths of Lg in cratonic platform areas, such as those in North America and Eurasia. Accordingly, while Sg may no longer be recognizable Lg may become the prominent phase at distances above a few hundred km (see Fig. 2.17 in Chapter 2) and well recordable up to about 30° epicentral distance.

The Lg phase is a complicated seismic signal. It is comprised of many surface-wave higher modes having group velocities around 3.5 km/s. Lg energy is strongly scattered by crustal heterogeneities (Knopoff et al., 1973; Bouchon, 1982; Kennett, 1985 ; Xie and Lay, 1994 and 1995). This complexity of Lg results in a very effective averaging of both the source radiation and crustal wave propagation effects. Therefore, statistical measures of Lg, such as rms amplitudes over a specified group velocity window, yield for given propagation paths very stable and accurate relative estimates of source strength, e.g., Lg magnitude values, which are useful in nuclear test discrimination and yield estimation (Xie and Lay, 1995). However, regional calibration for different travel paths is a prerequisite.

Lg magnitudes are calibrated either with respect to (or in a similar way as) Ml or to teleseismic mb. In the latter case they are usually termed mbLg or Mn (Ebel, 1982). Thus, when scaled to match with teleseismic mb at larger distances, a respective mb\_Lg scale, covering the range between 4° and 30° (as originally proposed by Nuttli (1973) can bridge the gap between more local Ml scales and teleseismic (D > 20°) mb.

Moreover, mb\_Lg has proven to yield rather stable magnitude estimates, even for single stations (Mayeda, 1993), with low scatter and thus good input data for estimating the yield of underground nuclear explosions (UNE) (e.g., Nuttli, 1986, Hansen et al., 1990). Further, since teleseismic P-wave records from earthquakes with magnitudes mb around 4 or less usually suffer of very low signal-to-noise ratio, mb\_Lg permits to extend the discrimination capability between UNE and natural earthquakes on the basis of the Ms-mb criterion down to Ms = 2.5 (Patton and Schlittenhardt, 2005; Richards, 2002). This has made mb\_Lg to a preference magnitude in the test-ban control community (e.g., Zhao et al., 2008). Yet, mb\_Lg has also become a favorite regional magnitude in general seismological practice of North America, southern Asia and in several northern European states such as Denmark, Finland and Norway (see **EMSC-CSEM Newsletter of Nov. 15, 1999** via <http://www.emsc-csem.org/Documents/?d=newsl>). One should note, however, that some of the so-called Lg scales are often not proper mb\_Lg - as shortly sketched below - but rather extensions of Ml, scaled to California Ml and not to teleseismic mb, yet measuring Lg amplitudes up to much greater distances than common for chiefly Sg-based Ml.

Nuttli (1973) was first to propose a conventional type of formulas for mb(Lg) for the eastern United States that have been routinely used at the NEIC:

$$\text{mb(Lg)} = 3.75 + 0.90 \cdot \log_{10} \Delta + \log (A/T) \text{ for } 0.5^\circ < \Delta < 4^\circ, \quad (3.17)$$

and

$$\text{mb(Lg)} = 3.30 + 1.66 \cdot \log_{10} \Delta + \log (A/T) \text{ for } 4^\circ < \Delta < 30^\circ, \quad (3.18)$$

where  $\Delta$  = distance in degrees, A = ground motion amplitude of Lg waves in microns, and T = period in seconds in the range 0.5 to 1.5 s.. A is the magnification-corrected, sustained (3<sup>rd</sup> largest peak) amplitude in the Lg train, measured on vertical-component, short-period (SP) WWSSN seismograms and corrected for period-dependent magnification.

The two log-linear formulas, which match at  $\Delta = 4^\circ$ , were required in order to approximate the empirical (A/T) observations in the Central U.S., which the exponential decay laws of Ewing et al. (1957), sufficiently well. And the constants had to be chosen such that mb(Lg) agrees for moderate size earthquakes in the U.S. with teleseismic mb(P).

Båth et al. (1976) developed a similar Lg scale for Sweden which is widely used in Scandinavia. Street (1976) recommended a unified mbLg magnitude scale between central and northeastern North America. Herrmann and Nuttli (1982) showed (later also Kim, 1998) that mbLg values are commonly similar to Ml when based on amplitude readings with periods around 1 s. They also proposed to define regional attenuation relations so that mbLg/Mn from different regions predict the same near source ground motions, i.e., that the mbLg scale becomes “transportable”. This new magnitude equation employs two terms to correct for amplitude decay (2<sup>nd</sup> term = geometrical decay and 3<sup>rd</sup> term = attenuation) and it requires that the attenuation  $\gamma$  of 1-Hz Lg waves is known beforehand. The formula reads:



$$mb(Lg) = 3.81 + 0.83 \cdot \log_{10} \Delta + \gamma \cdot (\Delta - 0.09^\circ) \cdot \log e + \log A, \quad (3.19)$$

where  $e = 2.718$ . Note that in this formula just amplitude  $A$  and not  $(A/T)$  is measured, again on WWSSN-SP records but in an even narrower range of period (0.8 s to 1.2 s). The constant in this formula was also determined by scaling  $mb(Lg)$  to teleseismic  $mb(P)$ . Specifically, for earthquakes in central United States, an  $Lg$  amplitude  $A$  of 115  $\mu m$  at distance  $0.09^\circ$  had to yield an  $mb = 5.0$  to be consistent with teleseismic magnitudes. This differs from the conventional  $mb(Lg)$  formula which yields at this distance an  $A/T$  of 155  $\mu m/s$  for  $mb = 5$ .

Herrmann and Kijko (1983) developed a frequency-dependent scale  $mLg(f)$  in order to broaden the frequency domain within which  $mbLg$  is applicable. Ebel (1994) proposed  $mLg(f)$ , calibrated to  $mb$  and computed with appropriate  $Lg$  spatial attenuation functions, to become the standard for regional seismic networks in northeastern North America. Ambraseys (1985) published calibration  $Q_g$  (for  $Sg$  and  $Lg$ ) and  $Q_R$  (for crustal Rayleigh waves), respectively that are applicable for northwestern European earthquakes in the distance range  $0.5^\circ < D < 11^\circ$ .

The final form of an amplitude-based  $mb(Lg)$  formula has been proposed by Nuttli (1986) and is written in two parts: First a corrected “hypothetical”  $Lg$  amplitude at distance 10 km =  $0.09^\circ$  is calculated on the basis of the general Ewing et al. (1957) decay law that fits best the empirical regional attenuation curves and then, in a second step,  $mb(Lg)$  using this hypothetical amplitude, normalized to 110  $\mu m$ , and the calibration constant. These formulas read:

$$A(10) = A(\Delta) \cdot (\Delta/10)^{1/3} \cdot [\sin(\Delta/111.1) / \sin(10/111.1)]^{1/2} \cdot \exp[\gamma \cdot (\Delta - 10)] \quad (3.20)$$

$$\text{and} \quad mb(Lg) = 5.0 + \log [A(10)/110], \quad (3.21)$$

where  $A(\Delta)$  is sustained ground motion amplitude in  $\mu m$  and  $\Delta$  is distance in km. The minor deviations from the earlier Herrmann and Nuttli (1982) formulas are, besides the scaling to 110  $\mu m$  instead of 115  $\mu m$ , that the acceptable period range has been somewhat relaxed (0.7 s to 1.3 s) and that only amplitudes in the group velocity window of 3.6 to 3.2 km/s should be measured.

Based on the formulas developed by Nuttli (1986) and Herrmann and Nuttli (1982) the CoSOI Working Group on Magnitude Measurement recommended now (see IASPEI, 2013 and/or IS 3.3) the following **IASPEI standard formula for  $mb(Lg) = mb\_Lg$** :

$$mb\_Lg = \log_{10}(A) + 0.833 \log_{10}[r] + 0.4343 \gamma (r - 10) - 0.87 \quad (3.22)$$

where  $A$  = “sustained ground-motion amplitude” in **nm**, defined as the third largest amplitude in the time window corresponding to group velocities of 3.6 to 3.2 km/s, in the period ( $T$ ) range 0.7 s to 1.3 s;  $r$  = **epicentral distance in km**;  $\gamma$  = coefficient of attenuation in  $km^{-1}$ .  $\gamma$  is related to the quality factor  $Q$  through the equation  $\gamma = \pi/(Q \cdot U \cdot T)$ , where  $U$  is group velocity and  $T$  is the wave period of the  $L_g$  wave.

*A and T are measured on output from a **vertical-component** instrument that is filtered so that the frequency response of the seismograph/filter system replicates that of a WWSSN **short-period** seismograph. Arrival times with respect to the origin of the seismic disturbance are used, along with epicentral distance, to compute group velocity  $U$ .”*



**Note:** “ $\gamma$  is a strong function of crustal structure and should be determined specifically for the region in which the mb\_Lg is to be used.”

Benz et al. (1997) describe in detail the procedure applied and the values derived in a comprehensive regional Lg attenuation survey for the continental United States.

The approximate equations (3.17) and (3.18) of Nuttli (1973) yield mb\_Lg that are about 0.1 m.u. smaller than those of the proposed standard procedure when  $\gamma = 0.00063 \text{ km}^{-1}$  in equation (3.22). Yet, preference is given to the proposed standard formula because of its transportability to regions with attenuation different than that of eastern North America (Nuttli, 1986; Patton, 2001).

**Recent researches** on mb\_Lg and regional variation of Lg attenuation **commonly use root-mean-square (RMS) amplitude measurements** rather than measurements made at a single peak (e.g., Patton and Schlittenhardt, 2005; Phillips and Stead, 2008). The use of RMS amplitudes offers the promise of more precise mb\_Lg measurements (according to Mayeda, 1993, with roughly 25% smaller residuals as compared to single amplitude mb\_Lg), but will require the development of a new formula, which should then be calibrated with respect to equation (3.22). See in this context also the last paragraph in the next section on Md and Mc. Developing the mb(Lg) scale for central Europe, Patton and Schlittenhardt (2005) had to calibrate all GRSN stations for site terms and for differences in Lg attenuation, because lateral variations in Lg Q turned out to be significant across the study area. Therefore, a regional Q model consisting of constant-Q partitions north, south and in the central Alps had to be developed.

#### 3.2.4.5 Duration and coda magnitudes (Md or Mc)

The pioneer in proposing earthquake magnitude determination via total signal duration has been the Hungarian seismologist Bisztricsany (1958). He scaled the time difference from the first P onset until then end of the surface-wave train in teleseismic records to magnitude. His method becomes now again of interest in estimating roughly the seismic moment of strong historical earthquakes from their duration in records of well defined historical instruments. Widely applied, however, have been duration magnitudes Md only since the late 1960s for local earthquakes. The simplest one is the coda magnitude (Mc), developed as a complement or alternative to single-amplitude based Ml. Signal codas are formed by scattered waves, i.e., waves which did not travel the minimum paths between seismic source and receiver, being reflected and diffracted by lateral inhomogeneities of the propagation medium. The stronger the primary signal, the larger the signal-to-noise ratio and thus the coda duration d. Different procedures are applied for determining signal or coda duration such as:

- duration from the P-wave onset to the end of the coda, i.e., where the signal disappears in the seismic noise of equal frequency, as in Fig. 3.31);
- duration from the P-wave onset to that time when the coda amplitudes have decayed to a certain threshold level, given in terms of average or RMS signal-to-noise ratio or of absolute signal amplitudes or signal level. This is objective, reproducible and best suited for automatic coda length determination, not, however, for manual ones.
- total elapsed time = coda threshold time minus origin time of the event.

Thus  $d$  becomes a good estimator of signal strength or event magnitude, which is in the local range, in agreement with single scatter theories, even independent on recording distance when the total elapse time is taken (Aki and Chouet, 1975). Yet, it is also possible to estimate the magnitude of large earthquakes even at teleseismic distances by using the P-wave coda (Houston and Kanamori, 1986). Another distinct advantage of  $M_d$  is that coda amplitudes depend much less than maximum phase amplitudes on source radiation pattern and directivity (Masuda, 1992).

One of the main reasons, however, for developing  $M_d$  scales was that analog paper or film recordings had a very limited dynamic range of only about 40 dB and analog tape recordings of about 60 dB. Even the dynamic range of early digital recorders with 12 or 16 bit A-D converters and thus 66 dB or 90 dB, respectively, was not yet enough to assure unclipped seismic records and  $A_{\max}$ -based magnitude determinations in the case of strong earthquakes recorded at local and regional distances.

In contrast,  $M_d$  (or  $M_c$ ) scales are based on either the total signal duration or the duration of the most pronounced coda of the Sg and/or Lg phase of an event. Coda length can easily be estimated even on clipped records of low dynamic range. However, with 24 bit A-D converters and  $\approx 140$  dB usable dynamic range now being standard, clipping is usually no longer a pressing problem. It is rare that an event is not considered for analysis because of clipped record traces.

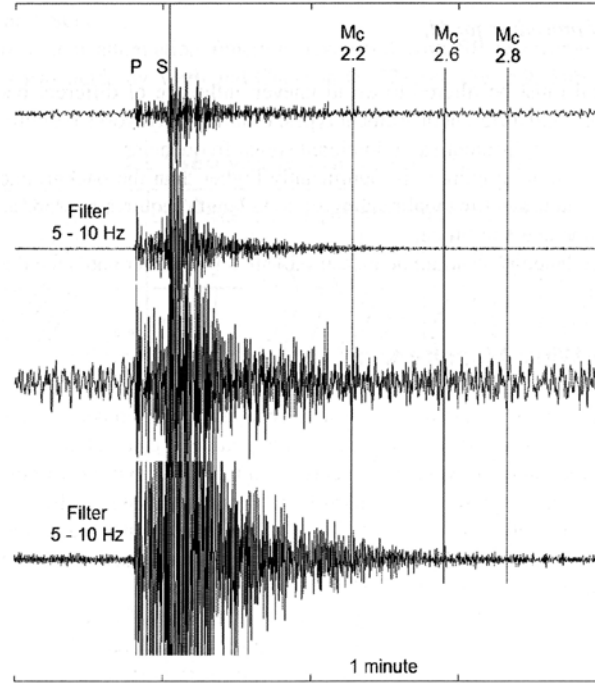
Note that scattering efficiency on small-scale heterogeneities depends on wavelength. Codas of seismic phases (also termed *signal-generated noise*) are well developed in short-period but not in long-period records and more pronounced in bandlimited than in broadband records (see Figs. 4.12-14 in Chapter 4). Moreover, SNR conditions at station and network sites influence the Coda-SNR and thus the measured coda length above noise level. Therefore,  $M_d$  scales should be developed for networks with given instrumentation (gain and bandpass range), noise conditions and practice a uniquely specified way of measuring signal duration and choosing the screen scaling appropriately. Fig. 3.31 illustrates to some extent, how variations in these parameters and related uncertainty in picking the coda end may influence the calculated  $M_d$  values.

Moreover, one should also know that short coda length ( $< 10$  s) are not a stable measure of earthquake size since they do not yet contain the backscattered energy from more distant heterogeneities (Aki and Chouet, 1975). Thus, one can not expect a generally valid relationship for  $M_d$  calculation for both small (in the near range only recorded) and large events (recorded also at larger distances) (e.g., Bakun and Lindh, 1977).

According to Aki and Chouet (1975) coda waves from local earthquakes are commonly interpreted as back-scattered waves from numerous heterogeneities uniformly distributed in the crust and that only one scatterer has been met by not-minimum path rays before reaching the station. Therefore, for a given local earthquake **at epicentral distances shorter than 100 km** the total duration of a seismogram is almost independent of distance and azimuth and of structural details of the direct wave path from source to station. Also the shape of coda envelopes, which decay exponentially with time (see Fig. 3.22), remains practically unchanged. The dominating factor controlling the amplitude level of the coda envelope and signal duration is then the earthquake size. This allows development of duration magnitude scales without a distance term if coda duration is measured from the origin, i.e.:

$$M_d = a_0 + a_1 \log d. \quad (3.23)$$

Thus, quick magnitude estimates from local events are feasible even without knowing the exact distance of the stations to the source.



**Fig. 3.31** Unfiltered (1<sup>st</sup> and 3<sup>rd</sup> trace) and filtered record traces of a local event with  $M_L = 2.8$  at station FOO in Norway in 117 km epicentral distance, with the 3<sup>rd</sup> and 4<sup>th</sup> traces being amplified for better noise resolution. Note the significant differences in  $M_c = M_d$  estimates using formula (3.27) below, depending on where one picks the end of the coda. High-pass filtering and record amplification may aid a more realistic picking. (Copy of Fig. 6.7 in Havskov and Ottemöller (2010); © Springer publishers).

At larger distances and with multi-scattering involved, also distance comes into play. A theoretical description of the coda envelopes as an exponentially decaying function with time and linear dependence on distance was presented by Herrmann (1975). He proposed a duration magnitude formula of the more general form:

$$M_d = a_0 + a_1 \log d + a_2 \Delta. \quad (3.24)$$

Most coda wave magnitude relationships are of this general type and use measured coda lengths larger than 30 s.

Havskov and Ottemöller (2010) show by way of example that the large variations in the formula parameters determined for different regions and thus the great variability of calculated  $M_d$  values for equal values of  $d$  are much larger than common for  $A_{\max}$ -based  $M_L$  scales. They also argue that most of these differences are due to different ways of measuring coda length and to differences in SNR conditions but not so much due to different tectonics and attenuation conditions.

A very early formula for the determination of local magnitudes based on signal duration was developed for earthquakes in Kii Peninsula in **Central Japan** by Tsumura (1967) and scaled to the magnitudes  $M_{JMA}$  reported by the Japanese Meteorological Agency:

$$M_d = 2.85 \log (F - P) + 0.0014 \Delta - 2.53 \quad \text{for } 3 < M_{JMA} < 5 \quad (3.25)$$

with  $P$  as the onset time of the  $P$  wave and  $F$  as the end of the event record (i.e., where the signal has dropped down to be just above the noise level),  $F - P$  in s and  $\Delta$  in km (also in all later following relationships)

Another duration magnitude equation of the same structure has been defined by Lee et al. (1972) for the **Northern California** Seismic Network (NCSN). The event duration,  $d$  (in s), is measured from the onset of the  $P$  wave to the point on the seismogram where the coda amplitude has diminished to 1 cm on the Develocorder film viewer screen with its 20 times magnification. With  $\Delta$  in km these authors give:

$$M_d = 2.00 \log d + 0.0035 \Delta - 0.87 \quad \text{for } 0.5 < M_l < 5. \quad (3.26)$$

The location program HYPO71 (Lee and Lahr, 1975) employs Eq. (3.25) to compute duration magnitudes, called FMAG. But it was found that Eq. (3.25) yields seriously underestimated magnitudes of events  $M_l > 3.5$ . Havskov and Ottemöller (2010) also calculated the greatest differences of this formula as compared to other formulas developed for other regions. Therefore, several new duration magnitude formulae have been developed for the **NCSN**, all scaled to  $M_l$ . One of the latest versions by Eaton (1992) uses short-period vertical-component records, a normalization of instrument sensitivity, different distant correction terms for  $\Delta < 40$  km,  $40 \text{ km} \leq \Delta \leq 350 \text{ km}$  and  $\Delta > 350 \text{ km}$ , as well as a depth correction for  $h > 10 \text{ km}$ .

Formulas according to (3.24) were also derived for **Norway** by Havskov and Sørensen (2006):

$$M_d = 3.16 \log d + 0.0003 \Delta - 4.28 \quad \text{for } 0.5 \leq M_l \leq 5, \quad (3.27a)$$

for **Mexico** by Havskov and Macias (1983):

$$M_d = 2.40 \log d + 0.00046 \Delta - 1.59 \quad \text{for } 3 \leq M_l \leq 6, \quad (3.27b)$$

for **East Africa** Dindi et al. (1995):

$$M_d = 1.9 \log d + 0.0004 \Delta - 1.2 \quad \text{for } 3 \leq M_l \leq 5, \quad (3.27c)$$

and for Northwestern Italy by Bindi et al. (2005) with significantly different coefficients for each stations of a network of 12 stations, ranging for the constant term between -1.7 and -2.7, for the distance term between  $(2.8 \text{ and } 4.9) \times 10^{-3}$  and for the duration term between 1.77 and 2.61.

In contrast, other formulas, mostly for smaller earthquakes recorded at shorter distances, were developed according to the general relationship (3.23), i.e., without a distance term for scaling smaller earthquakes recorded at closer distances, such as

by Bakun and Lindh (1977) for **California**:

$$M_d = 0.71 \log d + 0.28 \quad \text{for } 0.5 \leq M_l \leq 1.5, \quad (3.28a)$$

by Viret (1980) for the **eastern USA**:

$$M_d = 2.74 \log d - 3.38, \quad (3.28b)$$

and by Castello et al. (2007) for **Italy**, with mostly large station corrections  $Sc_j$  for stations in Southern Italy:

$$M_d = 2.49 \log d - 2.31 + Sc_j \quad \text{for } 1.5 \leq M_l \leq 4.5, \quad (3.29)$$

Other  $M_d$  formulas used in **Albania, France, Italy, Portugal** and former **Yugoslavia** can be found in the **EMSC-CSEM Newsletter of Nov. 15, 1999** via <http://www.emsc-csem.org/Documents/?d=newsl>), for **Greece** in Kiratzi and Papazachos (1985), for **Sweden** in Wahlström (1979), for the Vogtland swarm earthquakes in **Germany** in Klinge (1989), for the **Caucasus** area in Rautian et al. (1979) and for **Japan** in Masuda (1992).

According to Eaton (1992), average station residuals between amplitude- and duration-based local magnitude estimates are practically independent of distance from the epicenter to at least 800 km. Moreover, when properly scaled mutually, the difference between these two independent estimates is, when averaged over 0.5 m.u. intervals, less than 0.05 in the magnitude range  $M = 0.5$  to  $M = 5.5$ .

**Note:** Crustal structure, scattering and attenuation conditions vary from region to region and even from station to station (local site effects). The latter seem to dominate both amplitude-based and duration-based magnitude residuals. Older well-consolidated rocks produce negative residuals (due to smaller amplitudes and shorter durations) and younger unconsolidated rocks produce positive residuals (Eaton, 1992). No general formulas can therefore be given. They must be determined locally for any given station or regionally for every network and be properly scaled to the best available amplitude-based  $M_l$  scale. In addition, the resulting specific equation will depend on the chosen definition for  $d$ , the local noise conditions, the sensor response and gain operated at the considered seismic station(s) of a network.

Moreover, with  $mb\_Lg$  scales using increasingly coda envelopes instead of third peak or RMS amplitudes, they become a version of coda magnitudes scales  $Mc$ . Mayeda (1993), using 1-Hz  $Lg$ -coda envelopes, thus made very stable single-station estimates of magnitudes from Nevada test site underground nuclear explosions. Rautian et al. proposed already in 1981 the use of coda amplitude, not of duration, in the definition of coda-based magnitudes, since coda spectral amplitudes are closely related to the earthquake source spectrum (Rautian and Khalturin, 1978). Coda-amplitude based  $Lg$ -magnitude estimates have generally a five times smaller scatter. Rautian et al. (1981) designed two particular scales based on the records of short-period (SP) and medium-period (MP) instruments. A scale of this kind is still used routinely by the Kamchatka seismic network (Lemzikov and Gusev, 1989 and 1991; see also IS 3.7). The main advantage of such magnitude scales is their unique intrinsic accuracy; even a single-station estimate has a root-mean-square (RMS) error of only 0.1 or even less.

### 3.2.4.6 The Russian K-class system for classifying earthquakes

For quantifying the size of local and regional earthquakes in terms of released seismic energy the energy class (K-class) system has been developed since the late 1950s in the former Soviet Union (FSU). For small and moderate earthquakes in the catalog “*Earthquakes in the USSR*” only K values are given, supplemented by magnitude values based on body and surface-wave readings only for larger, world-wide recorded earthquakes (see section 3.2.5 on common teleseismic magnitude scales and Table 4 in DS 3.1).

Nature, origin, and methodology of the K-class system are poorly known to western seismologists studying Soviet and Russian seismological data, yet they are of great interest and importance to those conducting detailed research on the seismicity of the former USSR (FSU) with the aim to make it also quantitatively compatible with other seismicity data and maps on a regional and global scale (Rautian et al., 2007; Mackey et al., 2010). We have, therefore, added to NMSOP-2 an elaborate new information sheet **IS 3.7**. It outlines in detail the conditions to be met and the assumptions made in developing the K-scale, looks into the conversion relationships between different regional K-scales and ISC mb and Ms as well as Mw and Me magnitude.

K-class was defined as  $K = \log_{10} E_S$  with  $E_S$  in units of Joules. Yet, available instrumentation and analysis tools as well as simplifications introduced in the interest of mass routine measurements by technical personal resulted in the development of various regional K-scales which mostly do not measure  $(A/T)_{\max}$  (proportional to ground motion velocity) but displacement amplitudes A only. This, however, made in fact K to be a size estimator which is more closely related to Ml and for small to moderate size earthquakes also to Mw rather than being a good estimator of released seismic energy  $E_S$  or modern energy magnitude Me. An exception is the Kamchatka K, which is based on measuring  $(A/T)_{\max}$  (see Figure 5 in IS 3.7). Therefore, it scales also reasonably well with teleseismic Me based on the integration of squared velocity amplitudes on broadband records (see Figure 11 in IS 3.7).

Generally, there exist significant differences between the various local to regional K-scales, such as exist for any local-regional magnitude scale (see sections 3.2.4.3 to 3.2.4.5 for Ml, Md and mb(Lg)). This makes a simple and globally valid conversion of published K values into relevant magnitude equivalents impossible. For a very rough orientation, however, which K value would approximately correspond to a given “western” magnitude, or vice versa, Bormann derived, on the basis of a figure published by Riznichenko (1992) (see Figure 8 in IS 3.7), the following “conservative” formulas

$$\log E_S \approx K = 1.8 M + 4.1, \quad (3.30a)$$

or, when resolved for M,

$$M = 0.556 K - 2.3 \quad (3.30b)$$

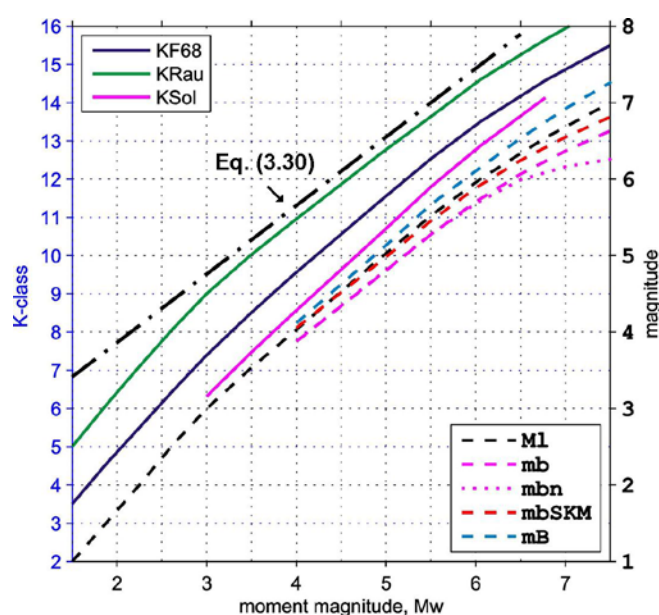
where M, depending on distance and event size, is supposed to relate either to Ml (for values << 4 to about 6), mb (for > 4 to about 5.5), mB (between about 5.5 to 7.5) or Ms (> 6 to about 8.5).

However, according to Fig. 3.32, the Eqs.(3.30a) and (3.30b) generally estimates for a given

Mw a higher K than any of the other K or magnitudes scales, or, for a given K underestimates Mw or the respective other magnitude values.. The differences in K are smallest and in the range  $3 < Mw < 6.5$  almost negligible when calculating K according to the linear Rautian (1960) formula for Tadshikistan, which is practically identical with Eq. (3.30b).:

$$M = (M_L) = 0.56 K - 2.22. \quad (3.31)$$

Yet, when lowering in Eq.(3.30) K by about 1.5 units this would result in a better **average** M to K conversion estimate, although with a large average deviation in the order of 1 unit in K with respect to other specific regional conversion relationships.

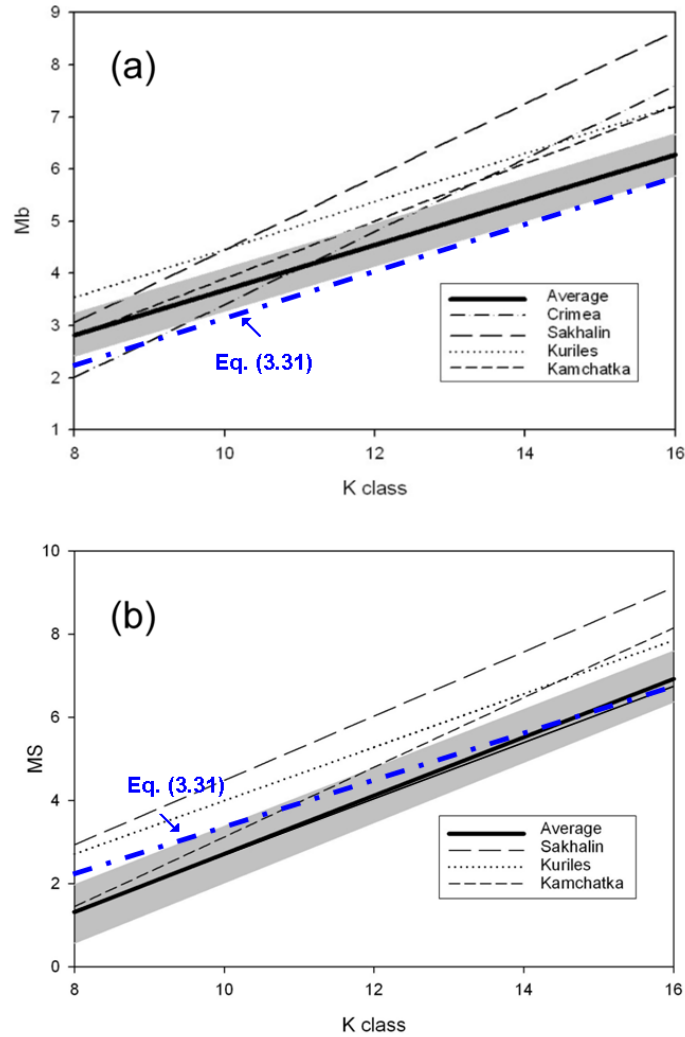


**Fig. 3.32** Non-linear relationships between K-class, Mw, Ml and different kinds of P-wave magnitudes according to Figure 9 in IS 3.7, provided by A. Gusev, in comparison with the linear M-K relationship of Eq. (3.30). For explanation of legend and references see IS 3.7.

This is confirmed when comparing Eq. (3.31) with the linear regressions between K and short-period  $mb = Mb$  in Figure 7a of IS 3.7, however Eq. (3.31) is for the various linear K-Ms relationships a somewhat better average estimator (Fig. 3.33). Thus, one should be aware of the uncertainty of K to M and M to K conversions which depend on the type of magnitude and the sometimes large differences between the various regional K scales. The causes of these differences may not even be of pure seismological reasons, such as differences in prevailing source and wave propagation processes. E.g., the discrepancy between KF68 and KSol in Fig. 3.32 is mainly due to an error in the absolute scaling constant of KSol, which result in K values that are 0.6 units too low (Fedotov, 1972). Also assumptions made about the shear wave velocity for calculating the wave energy density on the reference sphere have been somewhat different, e.g., for the Rautian and Fedotov K scales (see related discussion in IS 3.7). Such inconsistencies in K procedures and scaling may significantly biases the seismological conclusions which could be drawn from the data and should be eliminated.

None the less, the authors of IS 3.7 see great potential for modernizing and homogenizing the K-class procedures with the aim to establish an empirically founded simple and very useful

energy-related complement to both teleseismic  $M_e$  as well as to local and regional  $M_L$ . Proper  $M_e$  calculations, which are chiefly based on model assumptions and full-fledged energy integration, are hardly possible on a mass routine basis and at the higher frequencies at which small to moderate local and regional earthquakes are recorded. On the other hand  $M_L$  is, in contrast to **what K is supposed to be**, more closely related to seismic moment and  $M_w$  than to **released seismic energy**. But, in order to bring up K to what it was originally intended for, it is necessary that future K is consistently calculated only on the basis of maximum measured ground motion velocity amplitudes  $V_{\max}$  instead of displacement amplitudes  $A_{\max}$ . And  $V_{\max}$  should be measured directly on standardized, in all regions compatible short-period records with identical (real or simulated) transfer functions that are velocity proportional in a sufficiently wide bandpass range that covers the expected corner frequencies of the source spectra of all earthquakes in the K or magnitude range of interest. Similarly, a unique model and elastic parameter set used for calculating energy density as well as the procedures for calibrating regional K-scales to a standard reference K-scale have to be developed, agreed upon and carefully documented, as they exist for standard ML (see 3.2.4.3). Otherwise, no interregional compatibility of K values can be achieved.



**Fig. 3.33** Linear orthogonal regression relationships between (a) ISC  $m_b = M_b$  and (b) ISC  $M_s = M_S$  with different regional K scales (according to Figure 7 in IS 3.7) and their



comparison with Eq. (3.31).

### 3.2.4.7 The Japanese $M_{JMA}$ scale

The Japan Meteorological Agency (JMA) uses routinely its own magnitude scale.  $M_{JMA}$  magnitudes are determined from the maximum ground displacement amplitude  $A$  of the total seismic event record trace and measured at any station at epicentral distance  $\Delta < 2000$  km. The maximum amplitude may be either that of a surface wave or of a body wave recorded by a seismograph with an eigenperiod of 5 s. According to Katsumata (1996) two calibration functions are used:

a) for earthquakes at focal depth  $h < 60$  km Tsuboi's (1954) formula, which is said to be well calibrated with the Gutenberg and Richter (1954) magnitudes. It reads

$$M_{JMA} = \log \sqrt{(A_N^2 + A_E^2)} + 1.73 \log \Delta - 0.83 \quad (3.32a)$$

where  $A_N$  and  $A_E$  are half of the maximum peak-to-trough amplitudes in  $\mu m$  with periods  $T \leq 10$  s in the two horizontal components and  $\Delta$  is given in km, and

b) for earthquakes deeper than 60 km Katsumata's (1964) formula

$$M_{JMA} = \log A + K \quad (3.32b)$$

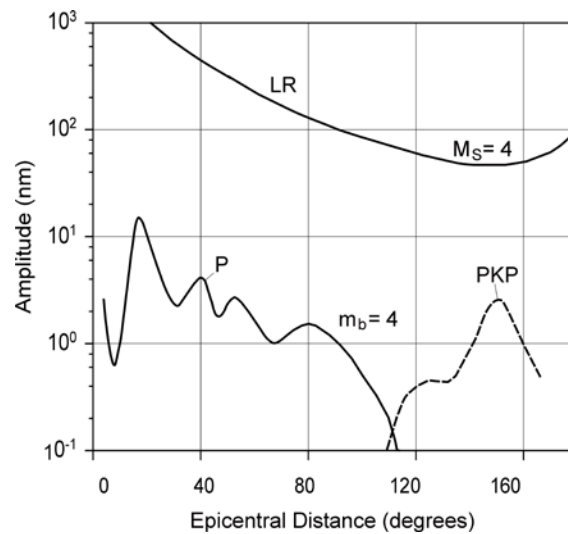
where  $K$  is a function of  $\Delta$  and  $h$  and given as a table by Katsumata (1964). Also this formula was scaled so that it gives almost the same magnitude values as that of Gutenberg and Richter (1954). Both Koyama et al. (1982) and Nuttli (1985) give rather similar (average graphical) relationships between  $M_s$  and  $M_{JMA}$ . The agreement is almost perfect at magnitude 7, however  $M_{JMA}$  tends to be increasingly larger than  $M_s(20)$  for smaller magnitudes, up to about 1 m.u. at  $M_{JMA} = 5$ .

Interestingly, the Tsuboi formula (3.32a) was derived on the basis of records of the astatic medium-period Wiechert horizontal pendulum records (see *Dedication to Wiechert and Galitzin* via the NMSOP-2 cover page), which had a flat displacement response for periods between about 0.1 and 5 s. All later introduced instruments or simulation filter responses used for measuring  $M_{JMA}$  mimic approximately the original Wiechert response. However, for earthquakes with  $M_w < 5.5$  nowadays also velocity proportional responses, flat between about 1 and 30 Hz may be used for measuring an  $M_{JMA}(\text{velocity})$ . Further, station magnitudes that deviate from the average network magnitude more than 0.5 m.u. are not taken into account in calculating the event magnitudes and mean values of the latter with standard deviations  $\geq 0.35$  m.u. are not given as event magnitudes (personal communication by Yasuhiro Yoshida, 1996).

## 3.2.5 Common teleseismic magnitude scales (P. Bormann and S. Wendt)

Wave propagation in deeper parts of the Earth is, within a few percent of lateral velocity perturbations (see Chapter 2, Fig. 2.80), more regular than in the crust and can for most purposes be described sufficiently well by 1-D velocity and attenuation models. This permits derivation of globally applicable teleseismic magnitude scales. Fig. 3.34 shows smoothed  $A-\Delta$

relationships for short-period P and PKP waves as well as for long-period surface waves for teleseismic distances, normalized to a magnitude of 4.



**Fig. 3.34** Approximate smoothed amplitude-distance functions for P and PKP body waves (at about 1 Hz) and of long-period Rayleigh surface waves LR at periods  $T \approx 20$  s for an event of magnitude 4.

From Fig. 3.34 the following general conclusions can be drawn:

- Surface waves and body waves have a different geometric spreading and attenuation. While the former propagate in two dimensions only, with amplitudes  $A$  decaying with distance  $\Delta$  roughly  $\sim \sqrt{\Delta}$ , the latter propagate three-dimensionally, with amplitudes decaying in a homogeneous full space  $\sim \Delta$ . Therefore, for shallow seismic events of the same magnitude, surface waves have generally larger amplitudes than body waves.
- Surface wave amplitudes change smoothly with distance. They generally decay up to about  $140^\circ$  and increase again beyond about  $150^\circ$ - $160^\circ$ . The latter is due to the increased geometric focusing towards the antipodes of the spherical Earth's surface which then overwhelms the amplitude decay due to attenuation.
- In contrast to surface waves, the  $A$ - $\Delta$  relations for first arriving longitudinal waves (P and PKP) show significant amplitude variations. The latter are mainly caused by energy focusing and defocusing due to velocity discontinuities in deeper parts of the Earth. Thus the amplitude peaks at around  $20^\circ$  and  $40^\circ$  are related to discontinuities in the upper mantle at 410 km and 670 km depth, the rapid decay of short-period P-wave amplitudes beyond  $90^\circ$  is due to the strong velocity decrease at the core-mantle boundary (“core shadow”), and the amplitude peak for PKP near  $145^\circ$  is caused by the focusing effect of the outer core (see Fig. 11.59??).

Other body wave candidates for magnitude determinations again behave differently, e.g. PP which is reflected at the Earth's surface half way between the source and receiver. PP does not have a core shadow problem and is well observed up to antipode distances. Furthermore, one has to consider that body waves are generated efficiently by both shallow and deep earthquakes. This is not the case for surface waves. Accordingly, the different  $A$ - $\Delta$ - $h$  behavior

of surface and body waves requires different calibration functions if one wants to use them for magnitude determination.

### 3.2.5.1 Surface-wave magnitude scales $M_s$ and the new IASPEI standards

Gutenberg (1945a) developed the magnitude scale  $M_s$  for teleseismic surface waves:

$$M_s = \log A_{H_{\max}}(\Delta) + \sigma_s(\Delta). \quad (3.33)$$

It is based on measurements of the maximum horizontal ground motion displacement amplitudes  $A_{H_{\max}} = \sqrt{(A_N^2 + A_E^2)}$  of the surface wave train at “periods of about” 20 s although Gutenberg used occasionally also periods as low as 12 s and as high as 23 s (Abe, 1981a; Lienkaemper, 1984; see section 4.4 in IS 3.3). The surface-wave maximum usually corresponds to the Airy phase, a local minimum in the group velocity dispersion curve of Rayleigh surface waves which arises from the existence of a low-velocity layer in the upper mantle (see Fig. 2.10 in Chapter 2).

It should be noted, however, that according to Geller and Kanamori (1977) Gutenberg did not measure the true maximum horizontal motion vector with both components measured at the same time. Rather, he combined the largest amplitude in the  $A_N$  component with the largest amplitude of  $A_E$ , although they might occur at different time and belong to different types of surface waves (see Figure 5 in Section 4.4 of IS 3.3). There was no corresponding formula given for using vertical component surface waves, which would have avoided this problem, because no comparably sensitive and stable vertical component long-period seismographs were available at that time.

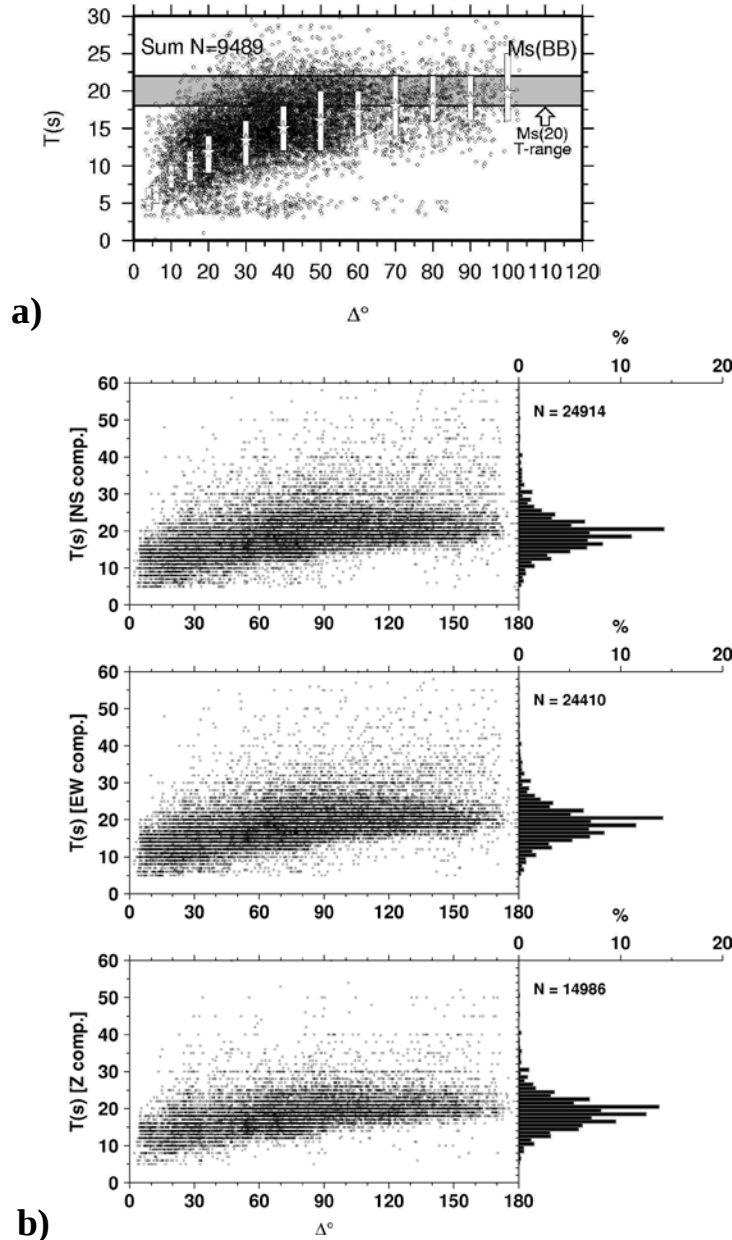
The calibration function  $\sigma_s(\Delta)$  is the inverse of a semi-empirically determined  $A$ - $\Delta$ -relationship scaled to an event of  $M_s = 0$ , thus compensating for the decay of amplitude with distance. The specified Gutenberg (1945a) formula, applicable between  $20^\circ \leq \Delta < 130^\circ$  epicentral distance, reads:

$$M_s = \log A + 1.656 \log \Delta^\circ + 1.818 \quad (3.34)$$

where  $\Delta^\circ$  is in degrees and  $A$  the ground motion displacement amplitudes in  $\mu\text{m}$ . Differences in the calibration term are  $<0.05$  m.u. as compared to respective tabulated calibration values published by Richter (1958) (see DS 3.1). The latter, however, account better than the simple formula for the energy focusing of surface waves towards the antipodes. The differences between  $M_s$  values calculated for distances  $>130^\circ$  according to (3.33) or by using the tabulated calibration values are 0.07 m.u. at  $140^\circ$ , 0.12 m.u. at  $160^\circ$  and 0.55 m.u. at  $180^\circ$ .

The Gutenberg relationship was further developed by Eastern European scientists. Soloviev (1955) proposed to use instead of the maximum ground displacement  $A_{\max}$  the maximum ground particle velocity  $V_{\max} = 2\pi(A/T)_{\max}$  since the latter is more closely related to seismic energy. Moreover,  $(A/T)_{\max}$  also better accounts for the large variability of periods at which surface-waves have their largest amplitudes, depending on distance of travel, crustal structure but also on magnitude and source depth, i.e., on the primary conditions of wave excitation. And finally, measuring  $(A/T)_{\max}$  permits rather stable magnitude estimates in a wider range of distances than the Gutenberg relationship which requires to read only periods around 20 s at teleseismic distances. Although for most continental Rayleigh waves the Airy phase periods

are around 10 to 15 s and for dominantly oceanic travel paths periods around 20 s and more (Bullen and Bolt, 1985; Marshall and Basham, 1972), the actual range of periods at which  $(A/T)_{\max}$  in the surface-wave trains is observed is much larger, ranging between  $2 \text{ s} < T < 60 \text{ s}$ . Periods, measured at the China Earthquake Network Center (CENC), have been plotted over epicenter distance  $\Delta$  in Fig. 3.35a and respective bulletin data prior to 1971 in Fig. 3.35b. Note the obvious distance trend of average  $T$  for  $\Delta < 70^\circ$ .



**Figure 3.35 a)** Variability of periods associated with  $V_{\max}$  of the Rayleigh wave group depending on distance  $\Delta$ . The gray shaded stripe marks the period range for which traditional  $M_s$  measurements at the U.S. Geological Survey's National Earthquake Information Center (NEIC) are carried out. Light gray columns give the observed range of average periods as published by Vaněk et al. (1962) and Willmore (1979), and the stars give the respective average periods of the plotted open circle data points within the related magnitude segment. Copy of Figure 8 in Bormann et al. (2009), p. 1874, © Seismological Society of America.

**b)** 3-component distance-period plots for  $(A/T)_{\max}$  surface wave readings retrieved at the ISC from printed station/network bulletins for earthquakes that occurred before 1971 and for stations shown in Figure 7 of Di Giacomo et al. (2013a). Figure by courtesy of Di Giacomo, 2013.

Based on early observations of this kind Soloviev and Shebalin (1957) and Karnik (1956), a team of researchers from Moscow and Prague proposed a new  $M_s$  scale and calibration function (Karnik et al., 1962 ; Vanek et al., 1962):

$$M_s = \log (A/T)_{\max} + \sigma_s (\Delta) = \log (A/T)_{\max} + 1.66 \log \Delta + 3.3. \quad (3.35)$$

It is applicable at epicentral distances  $2^\circ < \Delta < 160^\circ$ , for sources that are less than 60 km deep and to surface-wave amplitude readings with periods above 3 s. The IASPEI Committee on Magnitudes recommended at its Zürich meeting in 1967 the use of “the Moscow-Prague 1962 formula” as **standard for  $M_s$  determination**.

The recommendation to use the Prague formula as standard for  $M_s$  calculation has generally been accepted by the global seismological community. However, as outlined by Bormann et al. (2009), there have been widespread misunderstanding and wrong statements in publications by other authors about how and in which ranges of distance and periods this formula should be used. No wonder that the concerned U.S. agencies decided to use it only at teleseismic distances and for amplitude readings in the period range 18s to 22 s in keeping with Gutenberg’s original formula and recommendations. This, however, made the deliberate introduction of  $(A/T)_{\max}$  in the Prague formula practically meaningless, because differences in period readings in this narrow period range make for less than 0.1 m.u. difference. Yet, this U.S. practice, based on amplitude readings on records of the long-period WWSSN-LP instrument (see response in Fig. 3.20), became standard throughout the western seismological community.

In contrast, seismological observatories in the former Soviet Union (FSU) and its allied countries used formula (3.35), as recommended in the original publications and in the Willmore (1979) Manual of Seismological Practice. And A and T values were read on records of type Kirnos seismographs with displacement proportional response between about 0.1 and 10 to 20 s (see response Fig. 3.20), as most of the earlier data readings which had led to the derivation of the Prague formula.

In China, however,  $M_s$  values were routinely determined according to both  $M_s$  concepts. This provided to the IASPEI/CoSOI WG on Magnitudes a unique opportunity to compare the respective data measured on the same stations of the China National Seismic Network and analyzed by the same personnel at the China Earthquake Network Center (CENC). The results, published by Bormann et al. (2007), revealed both similarities and relevant differences of the respective  $M_s$  values and encouraged the WG to propose the following two **IASPEI standards for  $M_s$**  (see IASPEI 2005 and its updates of 20011 and 2013):

$M_s(20)$ , or  $M_{s\_20}$  in the ISF format (see Chapter 10) for data exchange and archiving, and  $M_s(BB)$ , or  $M_{s\_BB}$ , respectively. Their abridged definitions read as follows (for more details see IS 3.3):

$$M_{s\_20} = \log_{10}(A/T) + 1.66\log_{10}\Delta + 0.3, \quad (3.36)$$

“where  $A$  = **vertical-component** ground displacement in **nm** measured from the maximum trace-amplitude of a surface-wave phase having a period between **18 s and 22 s** on a waveform that has been filtered so that the frequency response of the seismograph/filter system replicates that of a World-Wide Standardized Seismograph Network (WWSSN) **long-period seismograph** ... with  $A$  being determined by dividing the maximum trace amplitude by the magnification of the simulated WWSSN-LP response at period  $T$ ,  $\Delta$  = **epicentral** distance in degrees,  $20^\circ \leq \Delta \leq 160^\circ$ ”.

Equation (3.36) is formally equivalent to the  $M_s$  equation proposed by Vaněk et al. (1962) but is here applied to vertical motion measurements in a narrow range of periods. Moreover,  $M_{s\_20}$  is expected to significantly underrepresent the energy released by intermediate- and deep-focus earthquakes due to their less effective generation of 20 s surface waves. Therefore, some agencies generally compute  $M_{s\_20}$  only for shallow-focus earthquakes, typically those whose confidence-intervals on focal-depth would allow them to be shallower than 50 or 60 km. For deeper earthquakes adjustment would need to be made, e.g., in accordance with Herak et al. (2001) (see also Question 4 below) in order to assure compatibility of  $M_{s\_20}$  data within the acceptable uncertainty limit of about 0.1 m.u. for IASPEI standard magnitudes.

For  $M_{s\_BB}$  the standard formula reads:

$$M_{s\_BB} = \log_{10}(V_{\max}/2\pi) + 1.66 \log_{10}\Delta + 0.3, \quad (3.37)$$

“where  $V_{\max}$  = ground **velocity in nm/s** associated with the maximum trace-amplitude in the surface-wave train, as recorded on a **vertical-component** seismogram that is **proportional to velocity**, where the period of the surface-wave,  $T$ , should satisfy the condition  $3 \text{ s} < T < 60 \text{ s}$ , and where  $T$  should be preserved together with  $V_{\max}$  in bulletin data-bases;  $\Delta$  = **epicentral** distance in degrees,  $2^\circ \leq \Delta \leq 160^\circ$ ”

Formula (3.37) is based on the  $M_s$  equation proposed by Vaněk et al. (1962), but is here applied to vertical motion measurements and is used with the  $\log_{10}(V_{\max}/2\pi)$  term replacing the  $\log_{10}(A/T)_{\max}$  term of the original. (3.37) could be further simplified by including  $-\log_{10}2\pi = -0.8$  into the constant, thus sparing one operation and reading

$$M_{s\_BB} = \log_{10}V_{\max} + 1.66 \log_{10}\Delta - 0.5. \quad (3.37a)$$

Also  $M_{s\_BB}$  should be calculated for shallow focus earthquakes only, unless appropriate depth corrections are made (see Question 4 below).

As of now (2012), both the ISC and NEIC use Eq. (3.36) for the determination of  $M_s$  from events with focal depth  $h < 60$  km without specifying the type of waves or components considered. The ISC accepts both vertical or resultant horizontal amplitudes of surface waves, with periods between 10 - 60 s from stations in the distance range  $5^\circ - 160^\circ$  but calculated the representative average  $M_s$  only from observations between  $20^\circ - 160^\circ$ . In contrast, the NEIC calculated so far only  $M_{s\_20}$  from vertical component readings of stations between  $20^\circ \leq \Delta \leq 160^\circ$  and for reported periods of  $18 \text{ s} \leq T \leq 22 \text{ s}$ . This limitation in period range is not necessary and limits the possibility of  $M_s$  determinations from weaker and regional earthquakes.

Recently, there has been again a tendency to determine the surface-wave magnitude by specifying the type of the waves and/or components used, e.g., MLRH or MLRV from

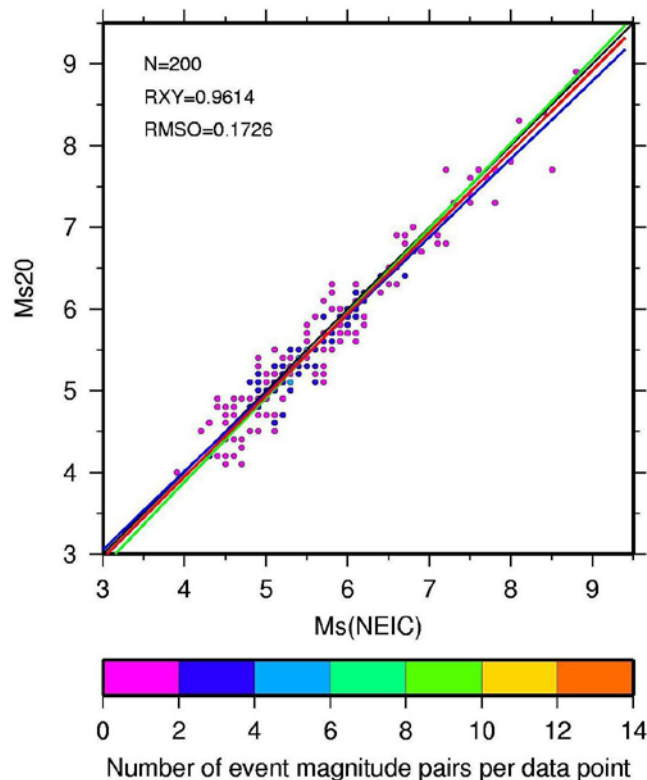
Rayleigh waves and MLQH from Love waves or simply MLH and MLV as was the practice in Eastern Germany in the 1960's (see Tab. 3.1) and recommended already in 1967 by the IASPEI Committee on Magnitude at Zürich. Since the newly proposed IASPEI Seismic Format (see 10.2.5) accepts such specifications (in the case that they require more than 5 characters in the metadata reported to data centers), the IASPEI WG on Magnitudes intends to elaborate recommendations for unambiguous standards and “specific” magnitude names. For a preliminary version see IS 3.2.

**Question 1: Are the new IASPEI Ms standards compatible with Gutenberg Ms?**

For 20 s surface waves of the same amplitudes Eqs. (3.35) and (3.36) yield about 0.18 m.u. (0.19 m.u. at  $\Delta = 100^\circ$ ) units larger Ms values than the original Gutenberg formula (3.34). Therefore, Abe (1981a) gave the following relationship between Ms determinations by NEIC using Eq. (3.35) and  $M_{GR}$  values published by Gutenberg and Richter (1954):

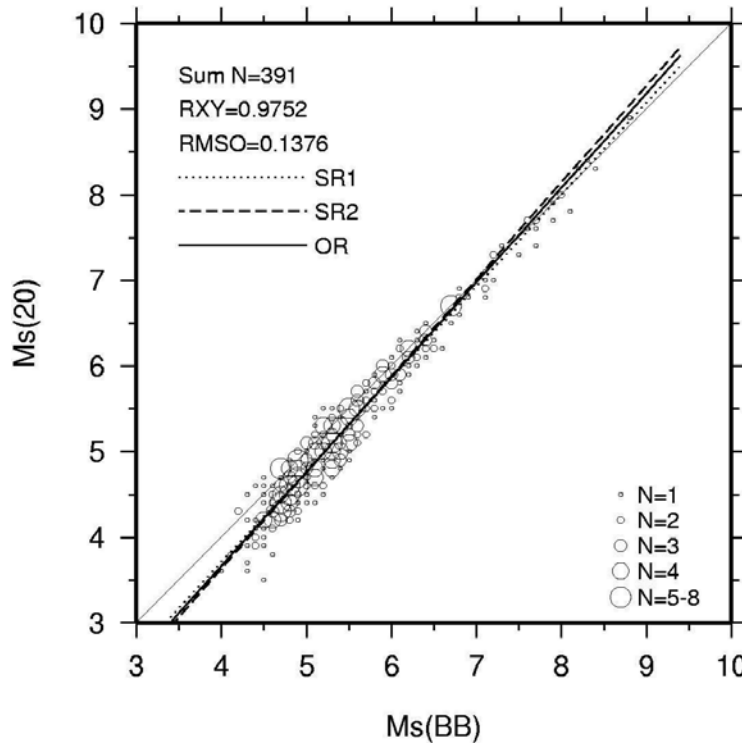
$$Ms(\text{“Prague”}, \text{NEIC}) = M_{GR}(\text{Gutenberg-Richter}) + 0.18. \quad (3.38)$$

This conclusion, however, could **not be confirmed** by Lienkaemper (1984). Recomputations with the Moscow-Prague formula for the same events were only 0.03 m.u. higher on average than  $M_{GR}$ . In section 4.4 of IS 3.3, Bormann and Dewey give additional reasons for the good agreement. Moreover, Bormann et al. (2009) compared  $Ms(\text{NEIC})$  with strict  $Ms_{20}$  determinations and found an excellent agreement (Fig. 3.36). Thus the **compatibility with earliest available and future Ms determinations at periods around 20 s is quantitatively assured** despite several changes in instrumentation and in the procedure of reading amplitudes on horizontal and later vertical-component recordings.



**Fig. 3.36** Data and linear regression relations (standard and orthogonal) between the new IASPEI standard  $M_s_{20}$ , measured at the China Earthquake Network Centre (CENC) by using China network data alone for earthquakes in the years 2001-2007, and the respective event  $M_s$ (NEIC) based on global data.  $R_{XY}$  – correlation coefficient;  $RMSO$  – orthogonal rms error, based on data published in Bormann et al. (2009).

NEIC never measured  $M_s$  in a wider range of period and distance. Therefore, a similar comparison cannot be made with  $M_s_{BB}$ . But Fig. 3.37 shows an excellent overall correlation between  $M_s_{20}$  and  $M_s_{BB}$ , yet with a slight systematic difference in trend: For magnitudes less than about 6.5  $M_s_{BB}$  tends to be increasingly larger, reaching on average 0.3 m.u. at  $M_s_{BB} = 4.5$  (Fig. 3.37). The main reason is that the average period at which the surface waves have their largest  $V_{max}$ , respectively  $(A/T)_{max}$ , depends also on magnitude, dropping from about 16 s at  $M_s_{BB} = 6.5$  down to about 11 s at  $M_s_{BB} = 4.5$ , i.e., well below the 18 to 22 s period range at which  $M_s_{20}$  is measured (see Figure 7 in Bormann et al., 2009). Accordingly, at  $T = 11$  s the ratio  $A/T$  is twice as large as at  $T = 22$  s. This makes for a difference of 0.3 m.u.



**Fig. 3.37** Data and linear regression relations (standard and orthogonal) between the new IASPEI standard magnitudes  $M_s_{20}$  and  $M_s_{BB}$ , measured at CENC.  $R_{XY}$  = correlation coefficient,  $RMSO$  = orthogonal root mean square error. Figure compiled from data published in Bormann et al. (2009).

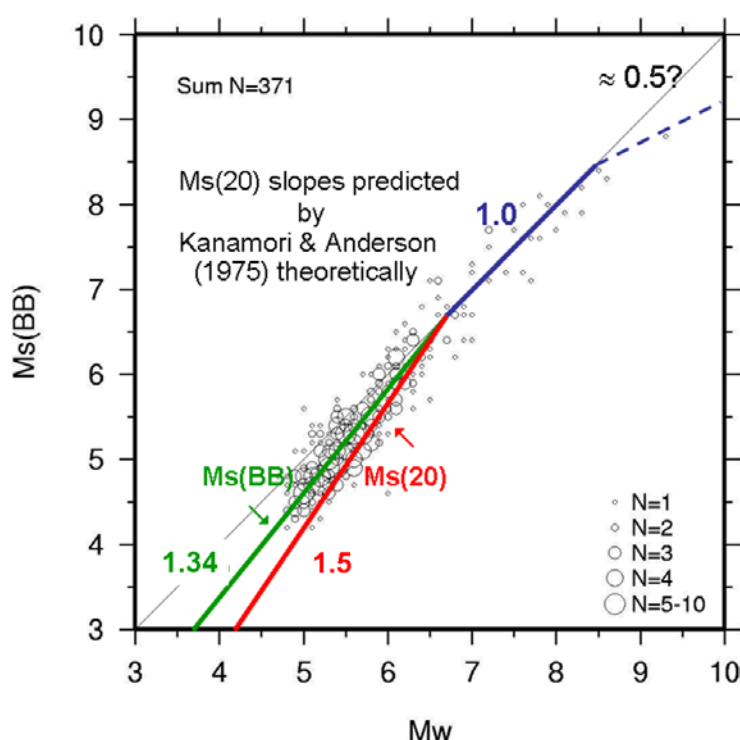
Figure 7 in IS 3.3 illustrates how in general (although not in each single case) the difference between  $M_s(BB)$  and  $M_s(20)$  grows with growing deviation of the periods at which  $V_{max}$  is measured in BB records from the narrow period range of  $M_s_{20}$ . Since surface waves of weaker earthquakes are recorded only at shorter distances with dominatingly shorter periods this magnitude dependence on the measured surface-wave periods translates also into a



distance dependence (see Fig. 3.35). One consequence is that  $M_s$ \_20 data systematically underestimate  $M_w$  at magnitudes below about 6.5 at which energy is systematically radiated at shorter periods than 20s (see Fig. 3.5 right; Kanamori and Anderson, 1975; Ekström and Dziewonsky, 1988). The then larger  $M_s$ \_BB values reduce this difference by almost 50% (Fig. 3.38).

While Ekström and Dziewonsky (1988) consider the significant slope change in the relationship between  $M_w$  and  $M_s = M_s$ \_20 for  $M_w <$  about 6.5 as an  $M_s$  bias in estimating earthquake size, others (J. Dewey, personal communication 2013) are even inclined to consider this as an advantage of  $M_s$ \_20, because then, in the range  $M_w \approx 4.5$  to 6.5,  $M_s$  is approximately proportional to  $\log M_0$ . However, as long as  $M_w$  is officially considered as being **the authoritative magnitude** that best represents earthquake size, even down to negative magnitudes (see, e.g., Kwiatek et al., 2010), and not  $M_s$  because of its near one-to-one relationship with  $\log M_0$  in a limited magnitude range, it is preferably to judge the earthquake size representativeness of other magnitude scales with respect to their relationship to  $M_w$  and not to  $M_s$  or  $\log M_0$ . When accepting this argument, then  $M_s$ \_BB is from all very easy to measure classical magnitudes the one which differs on average least from  $M_w$  ( $\leq 0.3$  m.u., Fig. 3.38) in the wide magnitude range from about  $M_w$  4.5 to 8.5. In contrast, the deviations are  $\leq 0.5$  m.u. for  $mB$ \_BB (see Fig 3.74 and Fig. 3.79 upper right), may reach about 0.6 m.u. or even more for  $M_s$ \_20 (Figs. 3.70a, 3.79 lower left, 3.81), 1.0 m.u. for  $ML$  and 1.0 -1.5 for  $mb$  (see Utsu, 2002, Fig. 3.70a; and for  $mb$  also Fig. 3.79 upper left and Fig. 3.81).

Moreover,  **$M_s$ \_BB**, is in tune with the  $M_s$  standard calibration equation (3.35), permits  $M_s$  determination down to epicentral distances of only  $2^\circ$  and tends to have smaller measurement errors than  $M_s$ \_20, both when measured interactively (Table 5 in IS 3.3) or in automatic mode (Tab. 3.2 and Figs. 3.25 and 3.26). Thus,  **$M_s$ \_BB** is a robust surface-wave magnitude which yields in some distance and magnitude ranges **valuable complementary information to  $M_s$ \_20**.



**Fig. 3.38** Data and linear regressions between the new IASPEI standard  $M_s$ \_BB, measured at CENC, and  $M_w$ (GCMT) as well as the approximate respective regression line of  $M_s$ \_20 for  $M_w < 6.5$ . Figure compiled according to data published in Bormann et al. (2009).

**Question 2: *Is  $M_s$  measured on horizontal and vertical components compatible?***

The calibration terms in both the Gutenberg and the Moscow-Prague formula for  $M_s$  determination had been derived for the correction of amplitude readings in horizontal (H) component records only. The introduction of stable long-period vertical component (V) seismometers into global seismological practice in the 1960s introduced another change in  $M_s$  measurement practice. According to Hunter (1972) the differences between  $M_s(V)$  and  $M_s(H)$  are negligible. This was confirmed by Bormann and Wylegalla (1975). According to their orthogonal regression relationship

$$MLV = 0.97MLH + 0.19, \quad (3.39)$$

with a correlation coefficient of 0.98 and a standard deviation of  $\pm 0.11$  m.u. only, the two magnitudes differ on average in the magnitude range between 4 to 8.5 less than 0.07 m.u. Thus, both findings **confirm the continuity and compatibility of old and modern  $M_s$  data**. Therefore, from May 1975 onwards, the USGS decided to calculate their  $M_s$  exclusively from vertical component readings. Since that time,  $M_s$ (NEIC) is in fact identical with the IASPEI confirmed standard magnitude  $M_s$ \_20.

**Question 3: *How reliably do the calibration formulas for  $M_s$  determination compensate for the distance dependence of amplitudes?***

Both the original Gutenberg relationship (3.33) and the Moscow-Prague IASPEI standard relationship (3.35) are rather simple and supposed to be used up to  $140^\circ$ , respectively  $160^\circ$  epicentral distance only. Reason: Both simple formulas do not well account for the amplitude increase beyond  $140^\circ$  due to the focusing of surface-waves near the antipode (see Fig. 3.34). This, however, can be avoided by using the respective tabulated calibration values instead (see DS 3.1., Tables 3 and 4). For  $M_s$ \_BB determination the tabulated Moscow-Prague calibration values agree at epicentral distances between  $1^\circ$  and  $140^\circ$  within 0.05 magnitude units with the values calculated from the calibration term in formula (3.35). For larger distances, however, this formula overestimates the magnitude between 0.05 and 0.55 (at  $180^\circ$ ) m.u. About the same applies when comparing values calculated with the Gutenberg formula (3.33) with the related tabulated calibration values published by Richter (1958). Therefore, **preference should be given to the use of the tabulated values, at least for distances  $>140^\circ$** .

However, several authors realized that using standard formula (3.35) for calculating 20 s  $M_s$  results in systematic distance-dependent biases (von Seggern, 1977; Herak and Herak, 1993; Rezapour and Pearce, 1998). Therefore, they proposed revised formulas for  $M_s$ \_20 calculation. The revised formula of Herak and Herak for  $M_s$ \_20 is:

$$M_s = \log (A/T)_{\max} + 1.094 \log \Delta + 4.429. \quad (3.40)$$

It is based on USGS data, i.e., on amplitude readings in the period range 18 to 22 s. It provides distance-independent estimates of  $M_s$  over the whole distance range  $4^\circ < \Delta < 180^\circ$ .  $M_s$  values according to Eq. (3.40) are equal to those from Eq. (3.35) at  $\Delta = 100^\circ$ , larger by 0.39 magnitude units at  $\Delta = 20^\circ$  and smaller by 0.12 units for  $\Delta = 160^\circ$ . Eq. (3.40) is practically equal to the formulae earlier proposed by von Seggern (1977) and similar to the more recent formula (18) obtained by Rezapour and Pearce (1998) on the basis of all ISC  $M_s$  data between 1978 and 1993:

$$M_s = \log (A/T)_{\max} + 1/3 \log(\Delta) + 1/2 \log(\sin\Delta) + 0.0046\Delta + 5.370. \quad (3.41)$$

Formula (3.41) makes allowance for the theoretically known contribution of both dispersion and the distance-dependent geometrical spreading, thus providing an improved overall distance correction, especially beyond  $\Delta = 145^\circ$  and reduces for intermediate size earthquakes the scatter against  $\log M_0$  as compared to  $M_{s\_20}$  calculated with the Moscow-Prague formula.

Developing new calibration formulas for  $M_{s\_20}$  is principally correct, because the IASPEI standard Moscow-Prague formula had not been developed for calibrating displacement amplitudes at periods around 20 s but for calibrating  $(A/T)_{\max}$  in a much wider range of periods and distances. Already Gutenberg (1945a) wrote in his  $M_s$  paper, that the 20 s condition “is not fulfilled for distances less than  $20^\circ$ .” Therefore, reported distance-dependent biases may reach at near regional distances up to about 0.5 m.u. and are in perfect agreement with the observed trend to shorter surface-wave periods as presented in Fig. 3.35, whereas for stronger earthquakes recorded at larger distances the difference between  $M_{s\_20}$  and  $M_{s\_BB}$  is in most cases negligible (Fig. 3.37). Yet, with the exception of formula (18) in Rezapour and Pearce (1998), which has become the standard  $M_s$  formula at the International Data Center (IDC) of the CTBTO (see Chapter 15), none of these alternative  $M_{s\_20}$  calibration relationships has already been accepted as a global standard. However, there is a need for it and it will surely reduce the currently existing systematic difference between  $M_{s\_20}$  and  $M_{s\_BB}$  in certain distance and magnitude ranges. There is hope that the now beginning global collection of standardized  $M_{s\_20}$  data with likely reduced procedure-dependent errors will finally result in a globally accepted new calibration function for standard  $M_{s\_20}$ .

#### **Question 4: Why is there no depth term in the $M_s$ calibration functions?**

The amplitudes of the fundamental surface-wave mode decay exponentially with depth (see Chapter 2) and routine depth determinations for shallow earthquakes ( $< 70$  km) are still rather uncertain (at present with errors in the order of 10-20 km, at Gutenberg’s time even worse or undetermined). This limited both the derivation and the meaningful application of depth corrections. On the other hand, more than 95% of the global seismic moment is released by shallow earthquakes. Therefore, Gutenberg (1945a) proposed to use his formula only for earthquakes at source depths not exceeding 40 km and to use only surface waves around 20 s which are also less sensitive to uncertainties in source depth than those with shorter periods. He even specified: “...it seems that for shocks with a depth of focus of about 35 km, 0.1 should be added to the magnitude calculated from (1) [here Eq. (3.33)] to bring the resulting  $M$  into agreement with the original zero point of Richter’s scale, and that the values given in table 4 are not affected by more than  $\pm 0.2$  by variations in focal depth so long as this does not exceed 40 km.” And in Gutenberg (1945b) he even wrote: “The determination of the magnitude of shocks with a focal depth in excess of about 30 km must be based on the

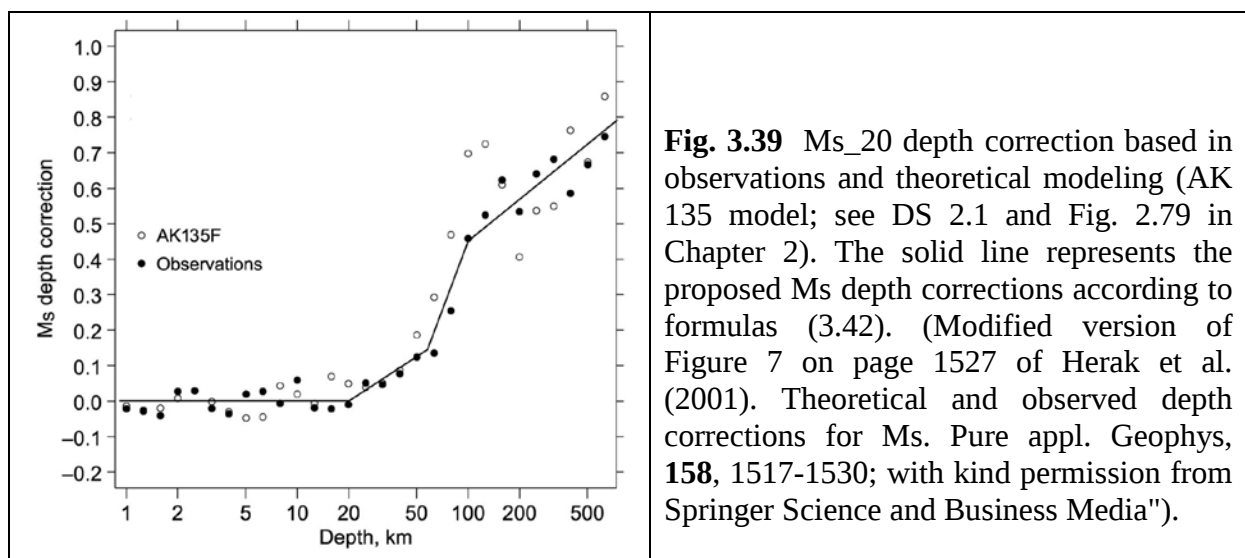
*amplitudes of the body waves*". Theoretical calculations by Panza et al. (1989) indicate that the depth correction may already exceed 1 m.u. even for shallow sources. This is confirmed by an empirical formula of Ochozinskaya (1974) used at seismic stations in Russia for determining the depth of shallow earthquakes ( $h < 70$  km) from the relationship between  $m_B$  and  $M_s$ :  $h(\text{in km}) = 54m_B - 34M_s - 107$  with a correlation coefficient of 0.88. For the same reasons also Vaněk et al. (1962) did not include depth corrections in their formula but relaxed somewhat the upper depth limit being aware of the uncertainties both of the initial hypocenter depth determinations and their limited relevance with respect to the hypocenter position of maximum seismic energy release in the case of large finite source ruptures.

#### **Question 5: Are depth correction terms for $M_s$ available and feasible?**

Empirically derived corrections for intermediate and deep earthquakes were published by Båth (1985). They range between 0.1 and 0.5 magnitude units for focal depths of 50 - 100 km and between 0.5 and 0.7 units for depths of 100 - 700 km. Based on model calculation and their comparison with real observations of 20 s surface waves from earthquakes down to  $h = 530$  km depth, Herak et al. (2001) proposed the following corrections  $\Delta M_s(h)$  for 20 s surface waves, which are in the same range as the empirical data by Båth (1985), i.e., up to about 0.75 m.u. near 600 km depth (see also Fig. 3.39):

$$\begin{aligned} \Delta M_s(h) &= 0 & \text{for } h < 20 \text{ km} & \quad (3.42) \\ \Delta M_s(h) &= 0.314 \log(h) - 0.409 & \text{for } 20 \text{ km} \leq h < 60 \text{ km} \\ \Delta M_s(h) &= 1.351 \log(h) - 2.253 & \text{for } 60 \text{ km} \leq h < 100 \text{ km} \\ \Delta M_s(h) &= 0.400 \log(h) - 0.350 & \text{for } 100 \text{ km} \leq h < 600 \text{ km}. \end{aligned}$$

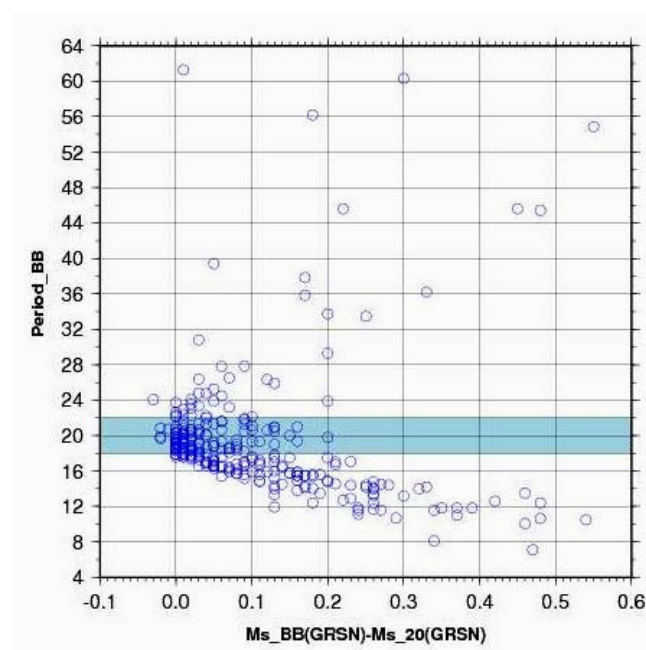
It would be desirable to test the feasibility of these correction terms globally based on type IAMs\_20 amplitude readings also for deeper earthquakes, maybe put in brackets, and to investigate their compatibility with  $m_b$ ,  $m_B$  and  $M_w$  estimates for these earthquakes. If such comparisons demonstrate the reliability, stability and thus feasibility of these depth corrections for  $M_s$  they may in future be integrated in a new  $M_s$ \_20 standard.



If  $V_{\max}$  is also found at periods around 20 s, than the Herak et al. (2001) corrections could be applied to  $M_s\_BB$  as well. However, the penetration depth of surface waves and therefore also the efficiency of surface-wave excitation depends on wavelength and thus period. If  $V_{\max}$  is measured at  $T \gg 20$  s (up to about 60 s) depth corrections may not be needed for source depth up to about 100 km. For periods of only about 5 s, however, the underestimation may become significant already for source depth larger than about 10 km. Regrettably, period-dependent source-depth correction terms have not yet been derived. With standard  $V_{\max}$  and related period data expected to become available en mass in future in the bulletins of international data centers it should be possible to determine such period-dependent empirical corrections in relation to those for  $M_s\_20$ .

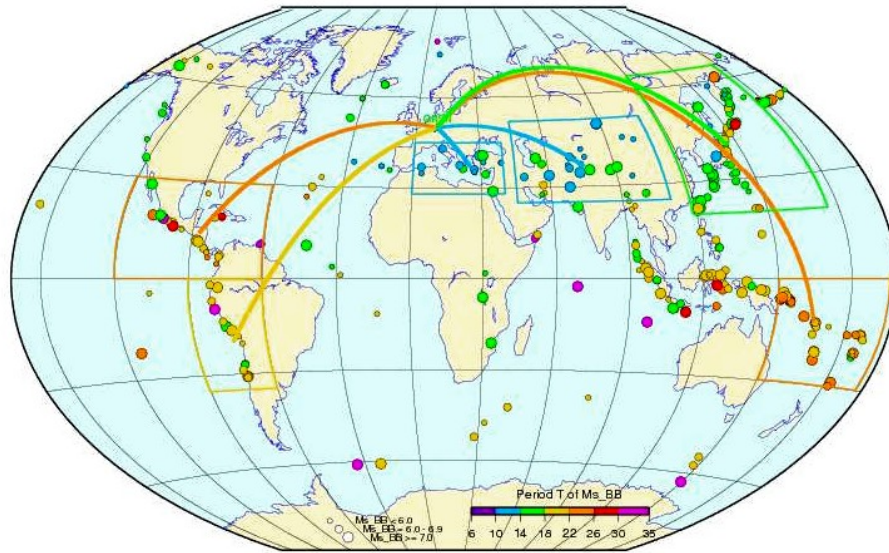
**Question 6: How does travel paths affect the period at which  $M_s\_BB$  is measured?**

Marshall and Basham (1972) investigated how much the travel paths of Rayleigh (LR) surface waves affect their measured displacement amplitudes at different periods. They found “...that for an impulsive input into a North American path the amplitudes at 10 secs are eight times larger than the amplitudes at 20 seconds, due to the dispersion characteristics of the path alone;...” In contrast: “...over a Eurasian path the amplitude difference between 10 and 20 seconds is only a factor of two. The predicted amplitude at 20 seconds period for both paths is however almost equal, ...” Moreover they showed that both oceanic and mixed intercontinental paths LR waves exhibit relatively low amplitudes at the shorter periods below 20 s, yet, for  $20 \text{ s} < T < 45 \text{ s}$  oceanic path LR amplitudes dominate those with continental or mixed intercontinental paths. Therefore, the periods at which the largest LR velocity amplitudes were measured at stations of the China National Seismic Network (CNSN) up to epicenter distance of  $100^\circ$  on LR waves with dominantly continental or mixed paths were in 79% of the cases outside of the 18-22 s period band at which  $M_s\_20$  is measured (see Fig. 3.35). In contrast,  $M_s\_BB$  determined at stations of the German Regional Seismic Network (GRSN) up to  $D = 160^\circ$ , which include a larger share of LR of mixed or even dominantly oceanic paths than in records of the CNSN, “only” 62% of the  $LR_{\max}$  were measured at periods outside of the 18-22 s range, again being mostly shorter but with a larger share of  $T > 22$  s, up to 62 s (see Fig. 3.40).

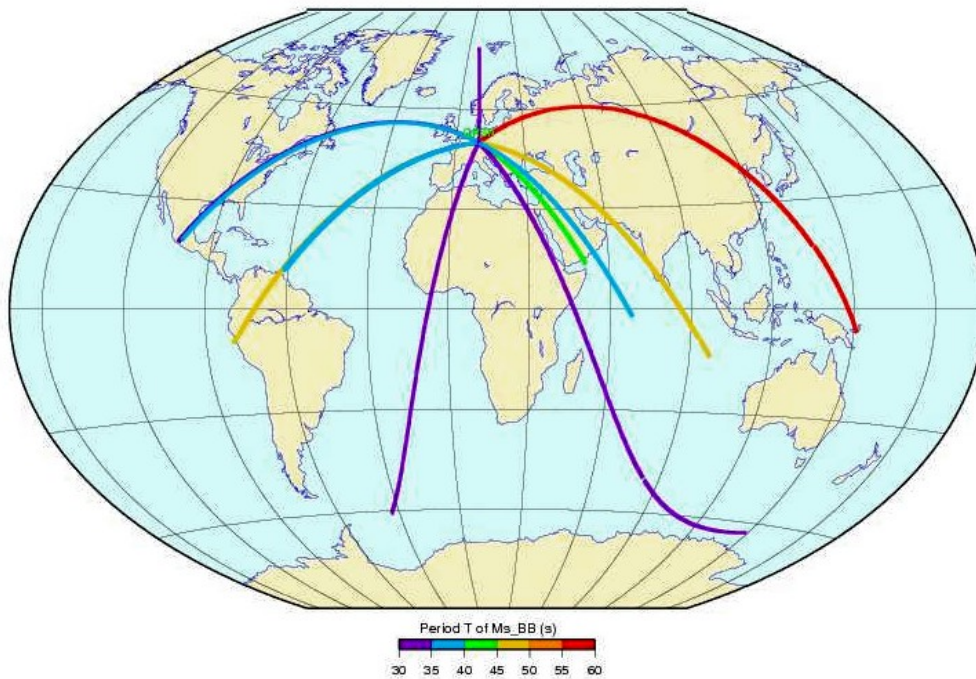


**Fig. 3.40** Difference between  $M_s\_BB$  and  $M_s\_20$  in relation to the period at which  $V_{max}$  was measured on velocity broadband records at GRSN stations in Germany.

Figs. 3.41 shows for different source locations, magnitude ranges and paths the periods at which LR  $V_{max}$  has been measured with  $T < 35$  s at station CLL of the GRSN and Fig. 3.42 for some average regional paths LR  $V_{max}$  measured at even longer periods up to 60 s.



**Fig. 3.41** Periods at which  $V_{max}$  of LR waves has been measured on velocity broadband records at station CLL, Germany, depending on source location, event  $M_s$  and travel path.





**Fig. 3.42** Periods between 30 s and 60 s at which  $V_{\max}$  of LR waves has been measured on velocity broadband records at station CLL, Germany, along some prevailing average regional paths.

From Fig. 3.41 it is obvious that the periods at which  $V_{\max}$  is measured are:

- Generally less than 18 s for pure Eurasian path up to about 80° long, and mostly even shorter (between some 6 to 14 s) for distances less than 30-50°, also when travelling through the African continent via the Mediterranean Sea to Central Europe;
- Mixed Rayleigh-wave paths arriving from comparable backazimuth, e.g., New Zealand, travelling through about half oceanic and half Eurasian continental crust, have already periods between 22 and 26 s.
- Mixed path intercontinental LR travelling from the western coasts of South America, Central America or North America, respectively, to Central Europe have typically periods between 18-22s, 22-26 s (more oceanic) and 14-18 s, respectively;
- Periods longer than 26 s have been measured for some individual earthquakes with either dominating oceanic paths or even for mixed paths, e.g. for the slow 17 July 2006 Mw7.7 Java tsunami earthquake (red dot in Fig. 3.41), whereas from other, faster rupturing earthquakes in the same region (green dots)  $LR_{\max}$  is often recorded at periods between 14 and 18 s.

Interesting in Fig. 3.42 is the rather different and somehow unexpected behavior of the even more long-period LR and mantle Rayleigh waves with periods between 30 s and 60 s. Practically all paths from the relevant source regions are mixed paths to CLL station. The shortest periods between 30 s and 35 s may occur along both the shortest as well as the longest paths. And the longest periods, up to about 60 s, have been measured for mantle Rayleigh wave maxima arriving on the long teleseismic mixed path from SE Asia (e.g., off-coast Papua New Guinea earthquake).

#### ***Question 7: Why does the IASPEI Ms formula not include a period dependence?***

Since Gutenberg restricted the use of his empirically determined  $M_s$  formula to amplitude readings around 20 s, for which according to the later finding by Marshall and Basham (1972) there is also the best agreement between the maximum LR amplitudes for different continental and mixed paths, there was no need for introducing a period-dependent attenuation term.

The situation, however, is different for  $M_{s\_BB}$  which is measured at periods  $3\text{ s} < T < 60\text{ s}$ . Although attenuation losses for periods  $> 20\text{ s}$  may become increasingly negligible, this is not the case for much shorter period. Yet, interestingly, the empirical data collected by Soloviev and Shebalin (1956) and Karnik (1957) revealed a much higher stability of the read velocity amplitudes - or ratio  $(A/T)_{\max}$  - as compared to the respective displacement amplitudes, in a wide period range and still a high overall compatibility of their data with Gutenberg's  $M_s$ . And since the authors of the Moscow-Prague formula, in the interest of best possible data continuity, scaled their data to the Gutenberg relationship they neither investigated nor introduced a period-dependence in their formula.

Ambraseys (1985) confirmed the stability and broad applicability of Moscow-Prague formula (3.35) when applied to European recordings of medium-period instruments at distances  $\geq 4^\circ$ . Quote: *The data available from Northwestern European earthquakes of the last 70 years recorded mainly by light and heavy Wiecherts, Galitzin, Mainka, Quervain-Picard, Milne-Shaw, Grenet and modern medium-period instruments, suggest that on the average for  $D \geq 4^\circ$ ,  $M$  values calculated from equation (1) [here Eq. (3.35)] do not exhibit an unequivocal distance dependence. ... The fit does not improve if we include distance and path corrections following, say, Marshall and Basham. ... Consequently, equation (1) was used over any distance and without any restriction for measurements to a fixed period other than the restrictions imposed on  $D$  and  $T$  by the data proper. ... neither a period constraint nor distance effects for the regional conditions considered seem to play an important role.*"

Also Okal (1989) came to the conclusion that the use of  $(A/T)$  in the Moscow-Prague calibration formula is a **partial and ad hoc compensation** for a large number of frequency-dependent terms ignored by it. The simplest explanation is that attenuation terms which linearly increase with frequency are compensated by multiplying  $A$  with  $f = 1/T$ . When looking at the tremendous range of period variations in Fig. 3.35 at which surface wave  $V_{\max}$  has been measured by seismic stations at distances between  $2^\circ$  and  $100^\circ$  one can hardly believe that they yielded the stable and with  $M_{S\_20}$  generally compatible  $M_{S\_BB}$  data plotted in Fig. 3.37, and this with even slightly smaller average standard deviation than for  $M_{S\_20}$  data (see Table 5 in IS 3.3).

How much  $(A/T)_{\max} = V\pi/2\pi$  readings may stabilized the  $M_S$  estimate has been illustrated by Bormann et al. (2009) with an example observed at the German Regional Seismic Network:

*05 Feb. 2006, Alaska, average network  $\Delta \approx 61^\circ$ , 15-station network  $M_S(BB) = 4.91 \pm 0.08$  at average network  $T = 15.4 \pm 8.3$  sec, station BFO  $M_S(BB) = 5.0$  at  $T = 7.1$  sec, station CLZ  $M_S(BB) = 4.9$  at  $T = 28.6$  sec, i.e., within 0.1 unit the same  $M_S(BB)$  despite  $T_{CLZ} \approx 4 \times T_{BFO}$ .*

Moreover, Ambraseys (1985) also found, that medium-period  $(A/T)_{\max}$   $M_S$  data correlate well with local magnitudes estimated by a number of European stations. This has been confirmed by S. Wendt for  $M_{S\_BB}$  data of station CLL too.

Yet, the phrasing "partial" compensation by Okal (1989) may be reason enough to look closer into the need – or not – of introducing for different seismotectonic environments, especially in the regional distance range, frequency-, path- and/or distance-dependent corrections into the current standard  $M_{S\_BB}$  formula (3.37). Mass data of standard  $M_{S\_BB}$ , expected to become available in future together with the periods at which  $V_{\max}$  has been measured, would be ideal for such studies.

#### **Question 8: Exist $M_S$ scales with a frequency-dependent calibration term?**

Yes, they do. E.g., Yacoub (1998) presented a method for accurate estimation of Rayleigh-wave spectral magnitudes  $M_R$  by velocity and frequency window analysis of digital records. He applied it to records of underground nuclear explosions in the distance range  $5^\circ$  to  $110^\circ$  and compared  $M_R$  with the classical time-window magnitude estimates,  $M_S$ , according to Eq. (3.35). While **both agreed well**, in general  $M_R$  had smaller standard deviations. Another



advantage is that the procedure for  $M_R$  determination can easily be implemented for on-line automated magnitude measurements.

More recently, Bonner et al. (2006) reported about the application of their time-domain variable-period surface-wave magnitude measurement procedure at regional and teleseismic distances. It is based on multi-bandpass filtering in the period range from 8 to 24 s in increments of 4 s, selecting the maximum bandpass amplitude and correcting it with a period-dependent correction term. Yet, also these authors found a generally good agreement of their  $M_s(VMAX)$ , on average better than 0.1 m.u., with the formulas of Rezapour and Pearce (1998) and Vaněk et al. (1962) when applied at teleseismic distances, with the exception of a small distance-dependent trend of  $-0.002$  m.u./° and a standard deviation of 0.21 m.u. of carefully measured station magnitudes for a subset of 33 events from the Mediterranean region.

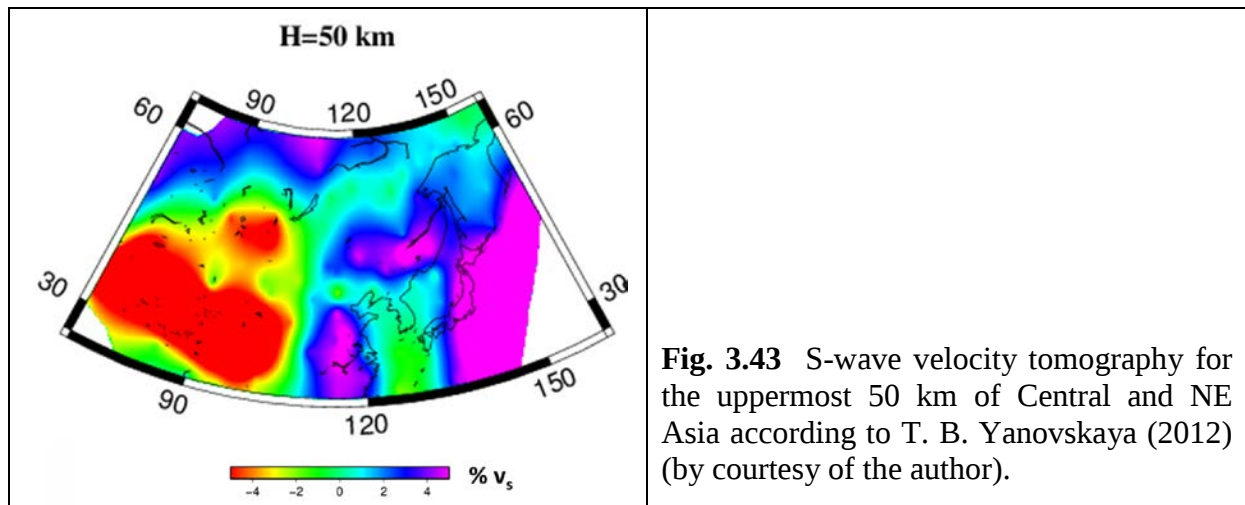
Yet, Bormann et al. (2009) got similar results for  $M_s\_BB$  with almost negligible differences when plotting for a much larger global event set and routine station readings the difference between  $M_s(BB)$  and  $M_s(NEIC)$  for earthquakes at depth  $h < 60$  km over the distance range  $2^\circ$  to  $103^\circ$ . When considering all measured periods of  $V_{max}$  in the range 3 s to 30 s (as compared to the 8-24 s data of Bonner et al.) the standard deviation of 0.33 m.u. and distance-dependent trend of  $-0.0029$  m.u./° were only somewhat larger. However, when using only periods  $< 15$  s in the whole distance range, for which larger attenuation effects would be expected, then the trend even reduced to merely  $-0.0009$  m.u./° and the standard deviation to 0.29 m.u.

Therefore, for the time being, all these comparisons still question the practical relevance of introducing frequency-dependent corrections in the  $M_s$  formulas in view of the generally large data scatter due to station site effects depending also on epicentral distance and backazimuth (see Question 9 on focusing and defocusing of LR waves). Probably the largest positive effect of the multi-bandpass filtering prior to  $M_s(VMAX)$  measurement for periods  $> 8$  s is, when compared with readings on unfiltered broadband records, that on stormy days ocean microseisms are strongly reduced and thus the SNR improved. This surely will reduce the standard deviation of  $M_s$  measurements. Since NEIC is now routinely calculating both  $M_s\_BB$  and  $M_s(VMAX)$  there may in future be more and better data available to assess the relative performance of these two kinds of variable frequency  $M_s$  procedures in different ranges of distance, backazimuth and source depth. This may help to decide about the necessity of introducing frequency-dependent attenuation terms also in the  $M_s\_BB$  formula. Maybe, the frequency-dependence of the quality factor  $Q$ , which we discuss below in conjunction with P-wave periods below 5 s, is another reason, why also surface-wave amplitudes measured at periods larger than 3 s for  $M_s\_BB$  or 8 s for  $M_s(VMAX)$  are less affected by attenuation towards shorter periods than expected from frequency-independent  $Q$  models. If this can be proved then no urgent need may be felt for introducing frequency-dependent attenuation corrections in the standard formula for  $M_s\_BB$  as long as the dominating station site and non-attenuative path effects due to unconsidered lateral velocity anomalies remain uncorrected.

#### ***Question 9: How do lateral velocity inhomogeneities affect the $M_s$ estimates?***

There may be significant regional and also station residual biases due to surface-wave path effects. Lateral velocity variations in the crust and upper mantle as well as refraction at plate

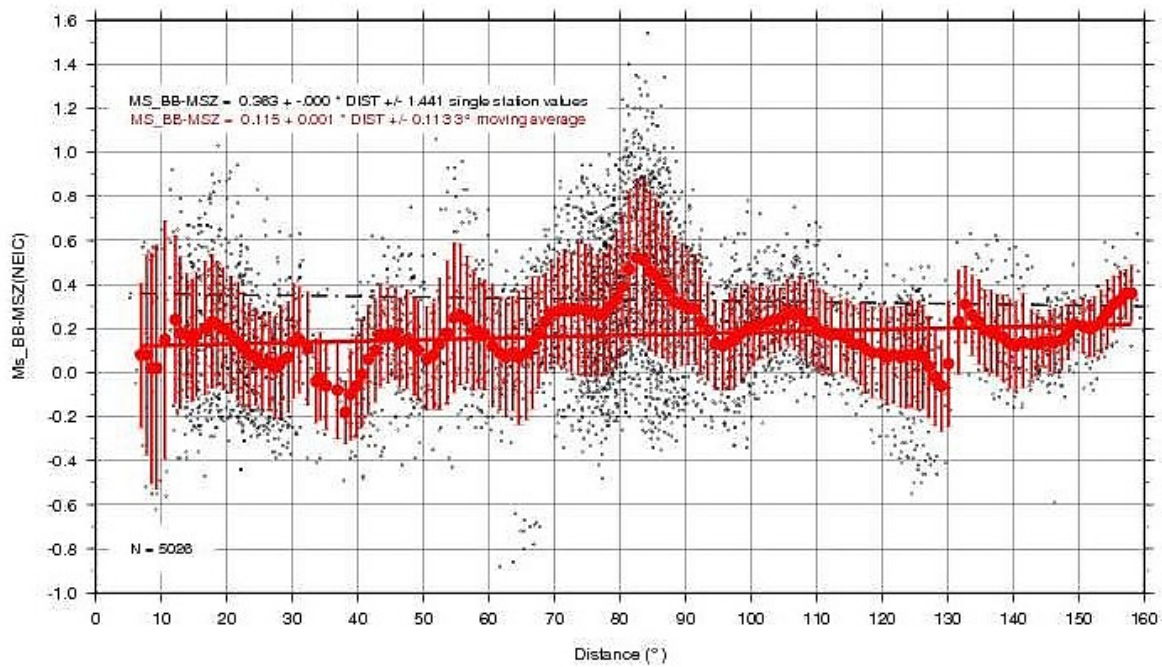
boundaries may result in significant focusing and de-focusing effects and thus regional over- or underestimation of station  $M_s$  or regional  $M_s$  averages as compared to global  $M_s$  event averages (Lazareva and Yanovskaya, 1975; Yanovskaya, 2012; Figs. 3.43, 3.44 a and b).



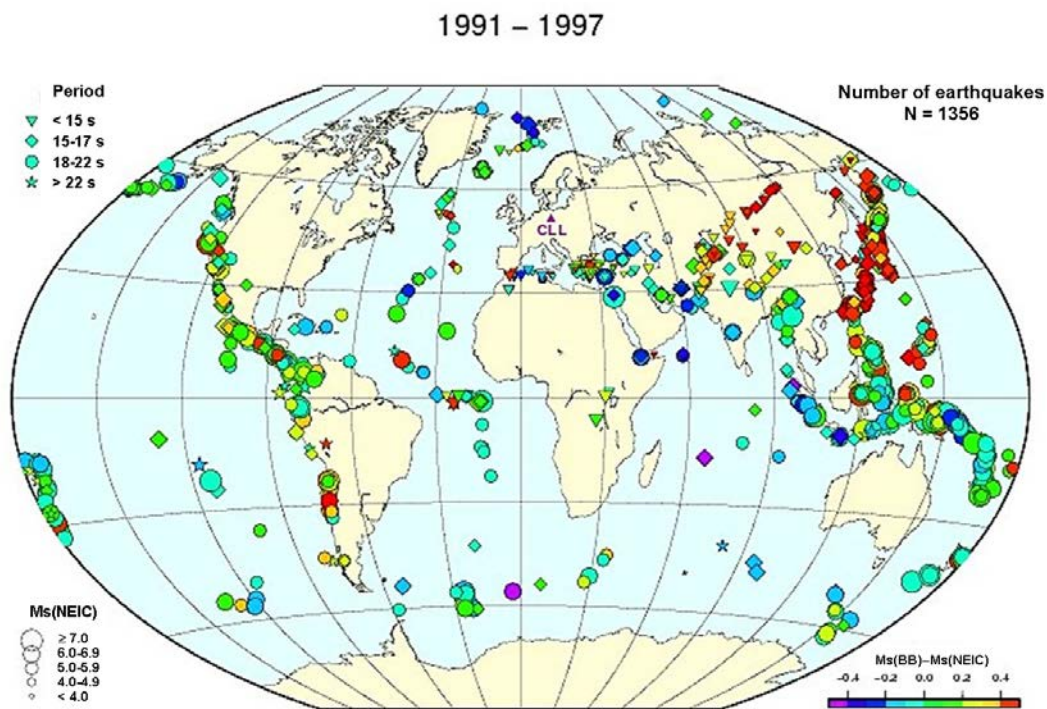
This is confirmed by comparing  $M_s$ \_BB measured at Collm (CLL) station in Germany with global  $M_s$ (NEIC) event averages (Fig. 3.44a). The  $M_s$ -residuals of CLL have been plotted over epicenter distance  $D$ . They show both a positive trend with growing distance as well as oscillations with average amplitudes up to +0.55 m.u. and individual data scatter up to more than one m.u. The large scatter also hints to another dependence on the backazimuth BAZ of the source region. This is confirmed by Fig. 3.44b when comparing the different average  $M_s$  residuals from the Kamchatka, Kuriles and Japan source region with those from SW North America, Central America and NW South America, all being about the same epicentral distance  $D \approx 85^\circ$  away from CLL but mostly with smaller and even negative residuals.

Relating  $M_s$  anomalies as a function of  $D$  and BAZ to major lithosphere velocity anomalies or plate boundaries along the way of travel would be a very important result. For enabling such investigations primary station residual data should be preserved and not be manipulated by trying to reduce them by applying station corrections before reporting the data to national or international seismological data or network centers.

According to Abercrombie (1994) surface-wave focusing seems to be the main cause for the anomalously high  $M_s$  magnitudes of continental earthquakes relative to their seismic moments, rather than differences in the source process. Therefore, she suggests that in order to obtain reliable, unbiased estimates of regional seismic strain rate and hazard, local/regional moment-magnitude relationships should be preferred to global ones.



**Fig. 3.44a** Difference between  $M_s_{BB}(CLL)$ – $M_s(NEIC)$ , based on earthquakes recorded between September 1992 and December 2007. Black dots - individual measurements, bold red dots – average values over increments of  $1^\circ$  in distance together with their standard deviation (vertical red bars). Red line - moving average trend. On average CLL  $M_s_{BB}$  is 0.36 m.u. larger than  $M_s(NEIC)$ .



**Fig. 3.44b** Differences between station  $M_s_{BB}(CLL)$  and global  $M_s(NEIC)$  event averages for different source regions.

**Question 10: How do deviations from the standard instrument responses affect  $M_s$ \_BB and  $M_s$ \_20?**

There is no adverse effect on the measured values of  $V_{\max}$  for  $M_s$ \_BB if the passband range of the velocity-proportional BB response covers the range of periods within which  $V_{\max}$  for  $M_s$ \_BB should be measured, i.e., between about 3 s and 60 s. At least periods up to 40 s, which make for the vast majority of data, should be covered (see Fig. 3.40).

With respect to  $M_s$ \_20 we show in IS 3.3 that even large deviations in the LP response from the recommended WWSSN-LP standard may not bias  $M_s$ \_20 magnitudes as long as it is assured, that the passband range of the alternative response covers with its maximum magnification the period range of 18 to 22 s. This could be demonstrated by filtering the BB records so that they simulate SRO-LP records. The SRO-LP response differs very much from the WWSSN-LP response (see Fig. 3.20) and has the advantage of strongly reducing ocean microseisms with periods between about 3 to 8 s, thus improving the SNR and increasing the number of events for which  $M_s$ \_20 can be measured. None the less, the derived  $M_s$ \_20 values agree within 0.1 m.u. with standard  $M_s$ \_20 (see Figure 14 in IS 3.3) and thus can be reported as such and also their amplitude readings with the nomenclature IAMs\_20.

### 3.2.5.2 Body-wave magnitude scales and the new IASPEI standards

Gutenberg (1945b and c) developed a magnitude relationship for teleseismic body waves such as P, PP and S, measured on broadband or long-period instruments at periods of 2-20 s (Abe, 1981a and 1984; Abe and Kanamori, 1979 and 1980). It is based on theoretical amplitude calculations from a point source, corrected for geometric spreading and only distance-dependent attenuation and then adjusted to empirical observations from shallow and deep-focus earthquakes, mostly in intermediate-period records:

$$mB = \log (A/T)_{\max} + Q(\Delta, h). \quad (3.43)$$

Gutenberg and Richter (1956a) published a table with  $Q(\Delta)$  values for P-, PP- and S-wave observations in vertical ( $V=Z$ ) and horizontal ( $H$ ) components for shallow shocks (see Tab. 7 in DS 3.1), complemented by diagrams  $Q(\Delta, h)$  for PV, PPV and SH (Figures 1a-c in DS 3.1) which enable also compatible magnitude determinations for both shallow and intermediate to deep earthquakes (see Abe and Kanamori, 1979).

In order to use the Gutenberg and Richter (1956a)  $Q(\Delta, h)$  PV diagram in computer processing, the NEIC manually digitized the diagram, creating a table with increments of source depth (see Table 2 in IS 3.3) and a computer program for interpolation between these values (PD 3.1). Both are now used by the NEIC and the ISC as the common basis for calibrating P-wave magnitudes, thus eliminating earlier slight discrepancies in their magnitude calculations due to somewhat different scanning and interpolation. Therefore, these calibration values and the interpolation program should be used in future by other seismological agencies and observatories as well. The  $Q(\Delta, h)$  values are correct when amplitudes are given in micrometers ( $10^{-6}$  m).

Gutenberg and Richter (1956a) proposed to calculate mB values, originally termed *unified magnitude*  $m$ , for all three types of body waves. Because of their different nature and/or

propagation paths they also differ in their frequency content. In addition, these body waves phases leave the source at different take-off angles and/or have different radiation pattern coefficients. Using them jointly for the computation of event magnitudes significantly reduces the effect of the source mechanism on the magnitude estimate, especially, if only data from a few station with insufficient azimuthal coverage are available. Gutenberg and Richter (1956a) also scaled  $m = m_B$  to the earlier magnitude scales  $M_L$  and  $M_s$  so as to match them at magnitudes between about 6 to 7.

**Note:** Up to the present there is still widespread confusion within the seismological community what the **unified magnitude  $m$**  really meant and aimed at. Even in recent publications one finds wrong equating  $m = m_b$  (e.g., in Scordilis, 2006). Many seem not to have read or have had access to the relevant original publications nor understood up to now the principal difference between the short-period narrowband  $m_b$  and the medium-period broadband  $m_B$ . Thus there is an obvious need for a clarifying summary.

Originally, Gutenberg and Richter aimed at calculating a “unified magnitude”  $m$  by converting body-wave magnitude estimates into equivalent  $M_s$  values, which they had agreed upon as being a magnitude standard. This has been practiced, with the exception for intermediate and deep earthquakes, more or less, when calculating the  $M$  values in Gutenberg and Richter’s (1954) “Seismicity of the Earth”. However, in Gutenberg and Richter (1956b) they express already:

*“It is believed that magnitudes determined from body waves of teleseisms are more coherent with the original scale (meaning  $M_L$ , P.B.) than those from surface waves, which have been in use as the general standard. If further investigation confirms this, it will be established that the magnitudes above 7 determined from surface waves have been overestimated, while those below 7 are underestimated.... A complete revision of the magnitude scale... is in preparation. This will probably be based on  $A/T$  rather than amplitudes.”*

Gutenberg (1956) added:

*“In the present paper the “unified magnitude”  $m$  is used throughout for energy calculations. It is assumed to be connected with **the old magnitude  $M_s$**  by*

$m = 0.63M_s + 2.5$  (1) *and with **the newly determined values of  $M_B$**  ... by  **$m = M_B$** . (2).*

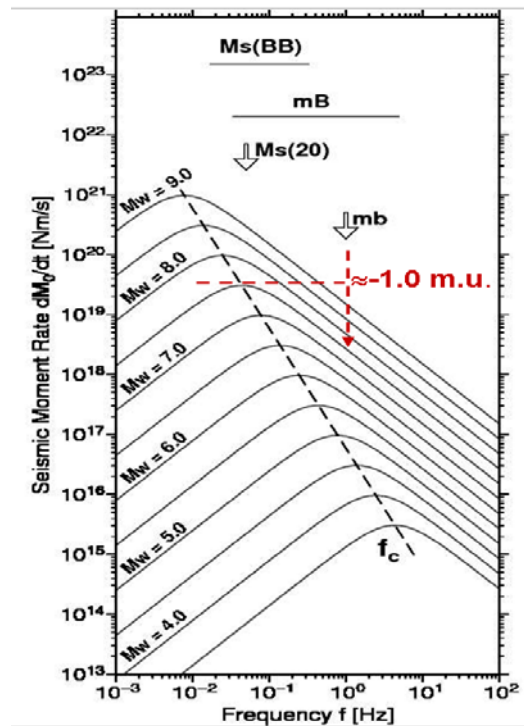
*The energy  $E$  of earthquakes is calculated from  $\log E = 5.8$  (in erg; P.B.) +  $2.4m$ . (3)*

*It should be emphasized that equations (1), (2) and (3) are tentative and should be revised if more data become available.”* (This we have done with Eqs. (3.98-3.100) for (1) and Eq. (3.137) for (3) for  $M_B = m_B = m_{B\_BB}$ )

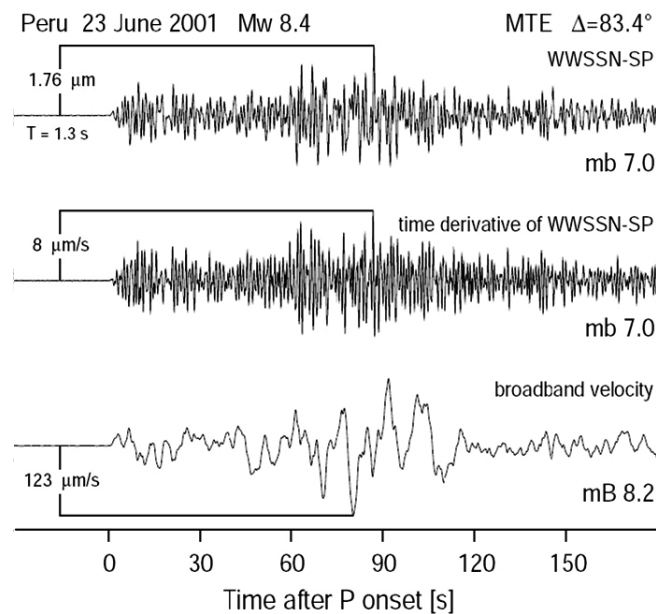
Finally, when publishing the revised body-wave magnitude calibration tables and charts, Gutenberg and Richter (1956) write consistently  $m_B$  instead of  $M_B$  and summarize:

*“Pending further research it is recommended that  $M_L$  continue to be used as heretofore, but  $M_s$  (and ultimately  $M_L$ ) should be referred to  **$m_B$  as a general standard, called the unified magnitude** and denoted by  $m$ . Tentatively.”*

Later, with the introduction on a global scale of the WWSSN short-period 1s-seismometers (for response see Fig. 3.20), it became common practice at the NEIC to use the calibration function  $Q(\Delta, h)$ , developed for medium-period amplitude readings on relatively broadband recording, for band-limited short-period PV measurements only. According to Fig. 3.45, this results in the case of strong earthquakes and when compared with their  $M_{s\_20}$  or  $m_B$  estimates, in an about 1 m.u. lower  $m_b$  value (termed *spectral component of saturation*). In addition, it was recommended in the early 1960s that the largest amplitude be taken within the first few cycles (Engdahl and Gunst, 1966; Willmore, 1979) instead of measuring the maximum amplitude in the whole P-wave train. To measure the largest amplitude within the first 5.5 s after the P onset is still the standard practice at the IDC of the CTBTO (see Chapter 15). Since the average rupture duration of earthquakes of magnitude 6, according to relationship (3.3), is about 6 s, such a fixed measurement time-window will add for larger earthquakes additionally a *time-window component of saturation* (see Fig. 3.46).



**Fig. 3.45** Illustration of estimating approximately the average amount of spectral magnitude saturation of mb with respect to Ms<sub>20</sub> or mB\_BB on the basis of the scaling law in Fig. 3.5.



**Fig. 3.46** Short-period WWSSN and broadband (BB) records of a strong Mw8.4 Peru earthquake. They illustrate the measurement of mb and mB at the largest amplitudes in the P-wave train near 90 s after the P-wave onset. The difference in velocity amplitude of 123 μm/s in the BB record and 8 μm/s in the time-differentiated WWSSN-SP record is due to spectral saturation according to Fig. 3.31, explaining the difference mb-mB = - 1.2. If one would measure for mb  $A_{\max}$  within the first 5.5 s, then this difference would increase by -0.5 m.u. (time-window component of saturation). (Copy of Figure 1 in Bormann and Saul, 2008, Seism. Res. Lett. 79, No. 5, p. 699; © Seismological Society of America).

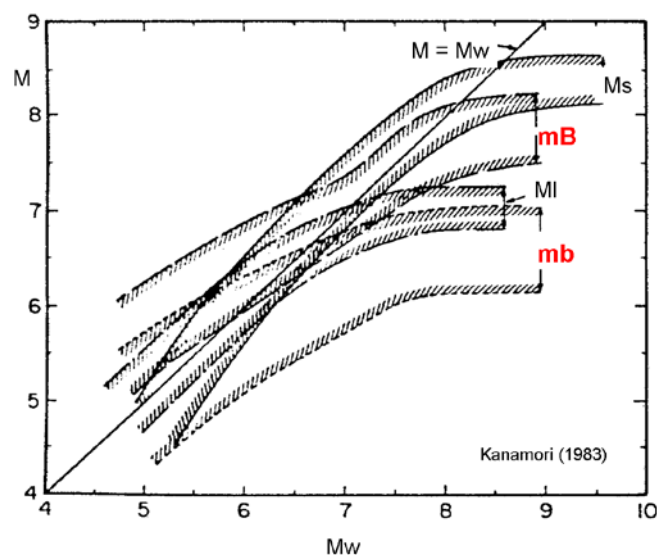


One should be aware that the practice of measuring the P-wave amplitude so early in so short a window was due to the focused interest in the 1960s on discriminating between earthquakes and underground nuclear explosions (UNE). For the latter, having source durations in the order of milliseconds only, their maximum P-wave amplitudes are always observed within the first few seconds after the P onset. Therefore, such a short time-window even improved the discrimination between earthquakes and UNE based on the  $M_s$ - $m_b$  criterion (see section 11.2.5.2??? of Chapter 11).

Yet, Bormann and Khalturin (1975) argued against it in the more general interests of earthquake seismology and recommended (quote):

*“We are of the opinion that the extension of the time interval for the measurement of  $(A/T)_{\max}$  up to 15 or 25 sec., as proposed ...in the Report of the first meeting of the IASPEI Commission on Practice (1972) ...is not sufficient in all practical cases, especially not for the strongest earthquakes with  $M > 7.5$ ....To reduce the systematic differences between magnitude determinations from body waves and surface waves, short-period or medium-period broad-band records should be preferred to short-period narrow-band ones for the MB-determination and the time interval for measuring the maximum value of  $A/T$  should be extended to about 1 minute for the strongest earthquakes.”*

In response, the IASPEI Commission on Practice revised its earlier 1972 measurement recommendation in 1978, stating that the maximum P-wave amplitude for earthquakes of small to medium size should be measured within 20 s from the time of the first onset and for very large earthquakes up to 60 s (see also Willmore, 1979, p. 85). This somewhat reduced the discrepancy between  $mB$  and  $m_b$  but in any event both are differently scaled to  $M_s$  and  $M_w$  with the short-period  $m_b$  necessarily saturating earlier than medium-period  $mB$  (see Fig. 3.47).



**Fig. 3.47** Approximate relationships and ranges of variability between  $M_l$ ,  $m_b$ ,  $mB$  and  $M_s$  with respect to non-saturating  $M_w$  (redrawn and modified from Fig. 4, p. 193, in Kanamori 1983; based on the kind permission of Elsevier Science Publishers for reprinting in NMSOP 2002).

However, some of the national and international agencies have only much later or not even as of now changed their practice of measuring  $(A/T)_{\max}$  for mb in a very limited time-window. E.g., the International Data Centre for the monitoring of the CTBTO still uses a time window of only 6 s (5.5s after the P onset), regardless of the event size. In contrast, the former Soviet/Russian practice (but also the East German practice, see Tab. 3.1 and Fig. 3.69 in section 3.2.5.2) of analyzing short-period records was always to measure the true P-wave maximum on the entire record. These magnitudes, reaching values between 7.3 to 7.6 for the strongest earthquakes, were denoted as MPV(A) or mSKM, because of using modified, relatively broadband short-period Kirnos type of instruments with a response as in Figure 1 of IS 3.7). Similar “whole P-wave train” mb magnitudes were later determined also by Koyama and Zeng (1985), denoted  $m_b^*$ , and by Houston and Kanamori (1986), denoted  $\hat{m}_b$ , which also reached values up to 7.6. With respect to saturation, mSKM,  $m_b^*$  and  $\hat{m}_b$  behave much like MI, as could be expected from their common frequency band used and considering that MI is determined also from the maximum amplitude in the whole short-period event record. MI saturates around 7.5 as well (see Fig. 3.47).

Taking into account the above facts, the CoSOI WG on Magnitudes recommends two P-wave **IASPEI standard magnitudes, mb and mB\_BB**, with suitably modified measurement procedures (IASPEI 2005, 2011 and 2012). In short they are defined as follows:

$$mb = \log_{10}(A/T) + Q(\Delta, h) - 3.0, \quad (3.44)$$

where  $A$  = P-wave ground amplitude in **nm** calculated from the maximum trace-amplitude in the **entire P-phase train** (time spanned by P, pP, sP, and possibly PcP and their codas, and ending preferably before PP);  $T$  = period in seconds,  $T < 3$  s; of the maximum P-wave trace amplitude.

$Q(\Delta, h)$  = attenuation function for **PZ** (P-waves recorded on vertical component seismographs) established by **Gutenberg and Richter (1956a)** in the tabulated or algorithmic form as used by the U.S. Geological Survey/National Earthquake Information Center (USGS/NEIC) (see Table 2 in IS 3.3 and PD 3.1);  $\Delta$  = **epicentral distance in degrees,  $20^\circ \leq \Delta \leq 100^\circ$** ;  $h$  = **focal depth in km**;

and where both  $T$  and the maximum trace amplitude are measured on output from a **vertical-component** instrument that is filtered so that the frequency response of the seismograph/filter system replicates that of a **WWSSN short-period** seismograph (see Figure 1 and Table 1 in IS 3.3) with  $A$  being determined by dividing the maximum trace amplitude by the magnification of the simulated WWSSN-SP response at period  $T$ .

Note that Fig. 3.7 in section 3.1.2.4 shows how new standard mb data relate to mb measured within the first 6 s after the P onset.

The standard relationship for mB\_BB reads as follows:

$$mB\_BB = \log_{10}(V_{\max}/2\pi) + Q(\Delta, h) - 3.0, \quad (3.45)$$

where  $V_{\max}$  = ground **velocity in nm/s** associated with the maximum trace-amplitude in the **entire P-phase train** (...as for mb...), as recorded on a vertical-component seismogram that is **proportional to velocity**, where the period of the measured phase,  $T$ , should satisfy the



condition  $0.2 \text{ s} < T < 30 \text{ s}$ , and where  $T$  should be preserved together with  $V_{\max}$  in bulletin data-bases; and  $Q(\Delta, h)$ ,  $\Delta$  and  $h$  as for mb. Equation (3.45) differs from the equation for  $m_B$  of Gutenberg and Richter (1956a) by virtue of the  $\log_{10}(V_{\max}/2\pi)$  term, which replaces the classical  $\log_{10}(A/T)_{\max}$  term.

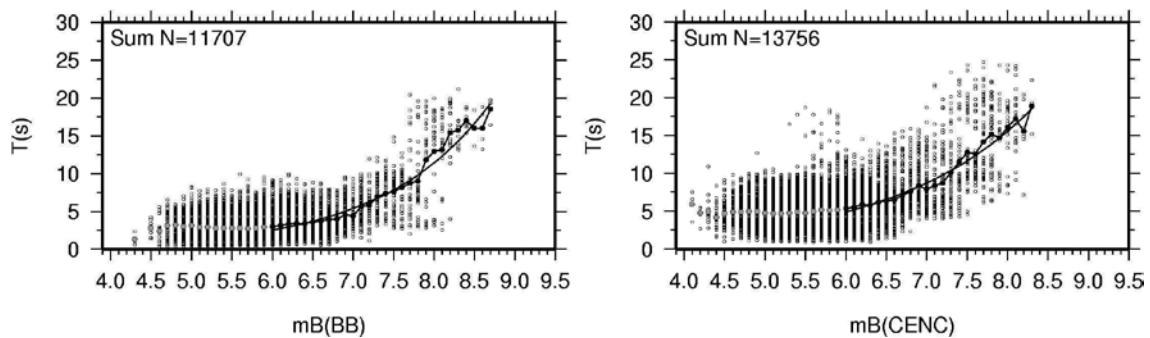
Formula (3.43) can be simplified by including  $-\log_{10}(2\pi) = -0.8$  into the constant, then reading

$$mB\_BB = \log_{10}V_{\max} + Q(\Delta, h) - 3.8. \quad (3.45a)$$

For more details and comments on mb and mB\_BB see IS 3.3.

### Question 1: Is modern mB\_B compatible with the classical Gutenberg mB?

Gutenberg measured himself or used measurements of P-wave amplitudes and periods published in other station bulletins that were commonly based in these years on relatively broadband medium-period instruments with responses that resemble that of the classical Russian Kirnos instruments of type SK and SKD (see the relative response curves 6 and 7 in Figure 1 of IS 3.7; for SKD also the Kirnos response in Fig. 3.20). Most of these response curves were more or less displacement proportional in the period range between  $0.1 \text{ s} < T < 10\text{--}2 \text{ s}$ , such as Mainka, Wiechert, Galitzin or Bosch-Omori (Kanamori, 1988, or Figure 5.6 in Lay and Wallace, 1995). Periods of  $A_{\max}$  measured in displacement records (and thus relating to  $(A_{\max}/T)$ ) tend to be somewhat larger than periods of  $V_{\max}$  in velocity records which correspond to  $(A/T)_{\max}$ . Although Gutenberg aimed at the latter, common practice has been to measure  $(A_{\max}/T)$  instead. This is obvious from Fig. 3.48 too, which compares traditional and period-wise more noisy measurements of P-wave  $A_{\max}$  for classical mB at the CENC with related measurements of  $T$  at  $V_{\max}$  for mB\_BB. Notable in both figures is the exponential increase of  $T$  for  $mB > 6$ . On average this is related to the exponential increase of the corner period of seismic source spectra with magnitude. The great scatter of measured periods at individual stations is both due to the stress drop-dependence of the corner period (see Fig. 3.15) and local station site effects. In records of weaker earthquakes with low SNR, however, the measured values of  $T$  may already be biased by the prevailing periods of ocean storm microseisms with periods between about 3 to 8 s.



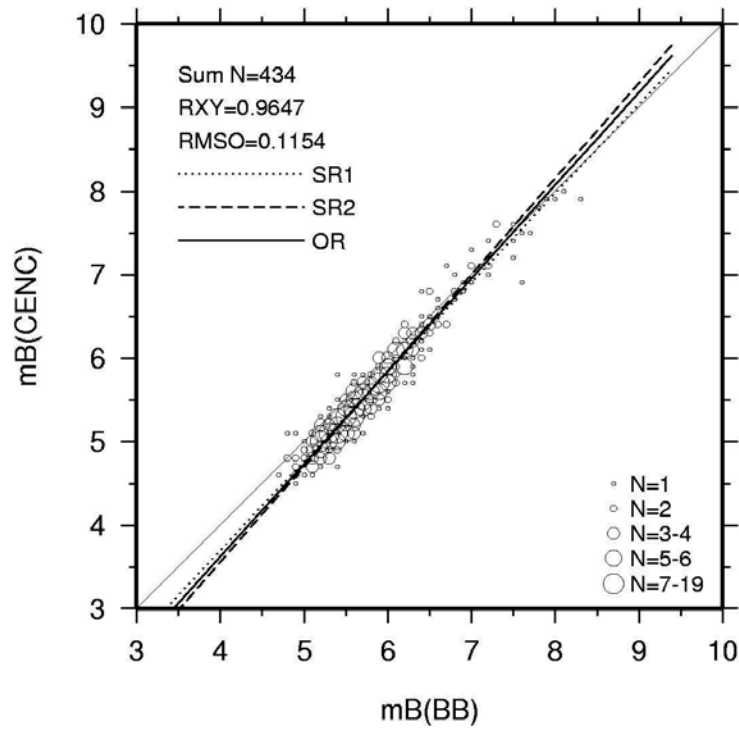
**Fig. 3.48** Dependence on mB\_BB (left-hand panel) and on mB\_CENC (right-hand panel) of the P-wave periods observed at individual stations when measuring  $V_{\max}$  on broadband-velocity records or  $A_{\max}$  on broadband displacement records of SK (Kirnos) type. The polygon line connects the average periods measured at the considered magnitude, the curved line is the exponential fit to it. (Copy of Figure 6 in Bormann et al. (2009), Bull. Seism. Soc. Am. 99, 3, p. 1873; © Seismological Society of America).

A consequence of this procedural difference of measuring directly  $V_{\max}$  (as for  $M_{s\_BB}$  too) may be the not perfect 1:1 trend of the regression relations between  $mB(BB)$  and  $mB(CEN)$  (see Fig. 3.49). For magnitudes  $< 6.5$   $mB(BB)$  yields somewhat larger values than  $mB$ . However, in the range  $mB > 6$ , in which Gutenberg measured this magnitude only, the average difference is  $\leq 0.1$  m.u. and the overall correlation coefficient is very high. Therefore, we can conclude that modern  $mB(BB) = mB\_BB$  is compatible with the  $mB$  intended and practised by Gutenberg within the acceptable error limit of 0.1 m.u. for the standard.

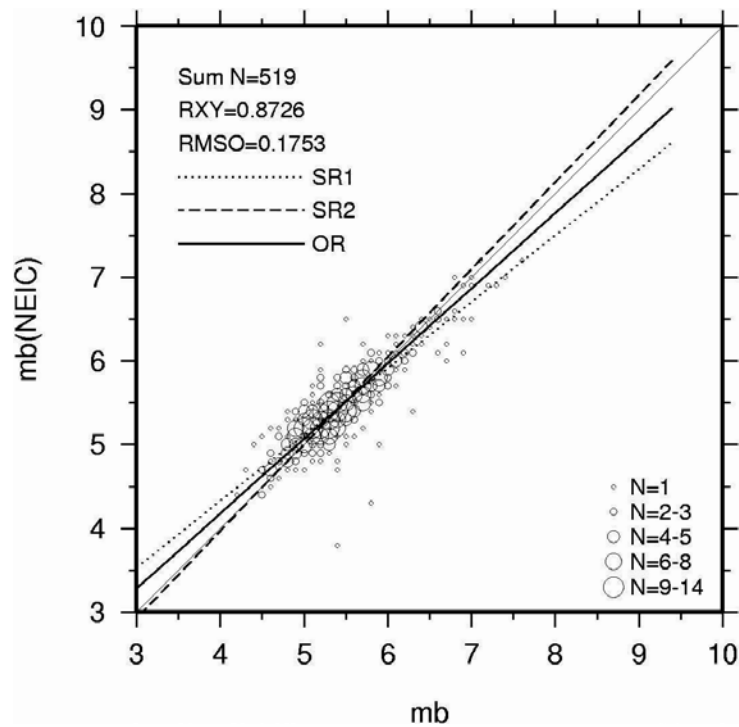
### ***Question 2: How relates standard mb to mb(NEIC)?***

Short-period narrow-band  $mb$  is not a body-wave magnitude recommended by Gutenberg. It has unilaterally been introduced by the concerned U.S. agencies in conjunction with the deployment of the World Wide Standard Seismograph Network (WWSSN) and short-period seismic arrays in the 1960s and 1970s. Since then  $mb$  has become an indispensable de facto standard magnitude, and the vast majority of global magnitude data is now  $mb$ . The reason is that short-period narrowband high-gain records enable magnitudes to be assigned to events that are more than an order of magnitude smaller than can be assigned with  $mB$  which is measured on much more noisy broadband records. However, since its inception, the measurement procedures for  $mb$  have changed several times.

Early U.S.  $mb$  values were measured within the first few P-wave cycles in short-period records. They saturated already near to 6 (see Fig. 3.7). However, following the IASPEI resolution of 1978, the USGS/NEIC changed practice later on and measured  $A_{\max}$  routinely either in the first 20 s or, after automation of the analysis procedure, within the first 10 cycles. Yet, analysts were instructed to extend this window in the final review for the PDE up to 60 s in the case of strong earthquakes. Moreover, the early WWSSN-SP records were later substituted by filtering with the PDE response, which has a steeper roll-off for  $f > 1$  Hz, a slightly larger relative bandwidth and period of maximum gain (see Figure 3 in IS 3.3). Nevertheless, one can expect standard  $mb$  to be close to post-1978  $mb(NEIC)$ , at least for the vast majority of data. According to Fig. 3.50 this is true for  $mb$  between about 4 and 6.5, for larger magnitudes, however, again  $mb(NEIC)$  tends to be somewhat smaller, saturating around 7 instead of around 7.5 as standard  $mb$ . Also, when comparing these two data sets, the data scatter is significantly larger and the correlation coefficient much smaller than in the comparison of standard  $M_{s\_20}$  with  $M_s(NEIC)$  (Fig. 3.37).



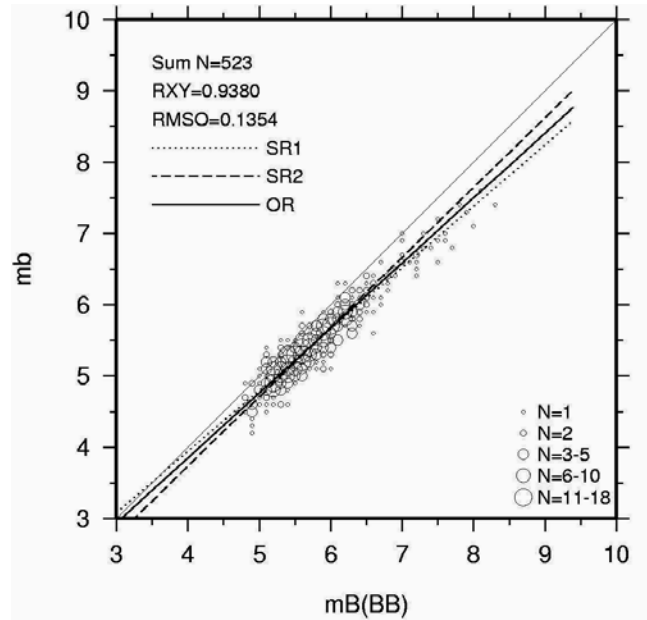
**Fig. 3.49** Regression relationships (standard and orthogonal) between  $mB(BB)$  and classical  $mB(CENC)$ .  $RXY$  = correlation coefficient,  $RMSO$  = orthogonal root mean square error. Figure based on data published in Bormann et al. (2009).



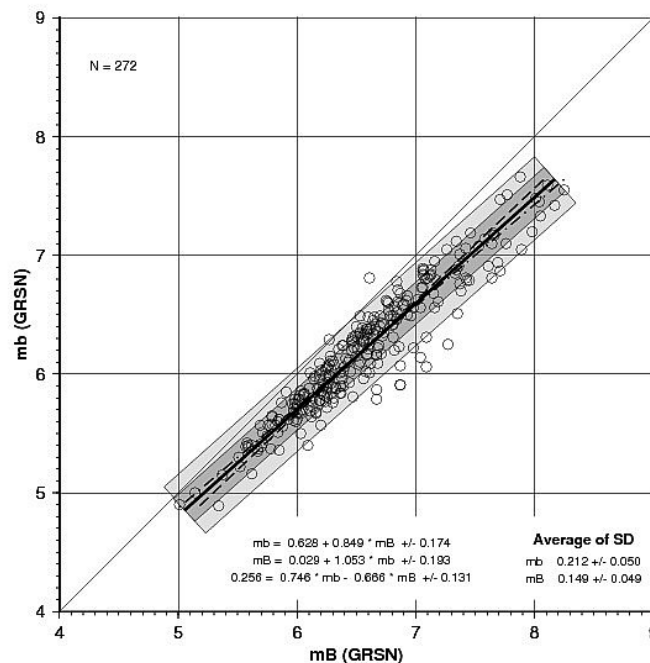
**Fig. 3.50** Regression relationships (standard and orthogonal) between standard  $mb$ , measured at the China Earthquake Network Center (CENC), and  $mb(NEIC)$  for the same events. Legend as in Fig. 3.49. Figure based on data published in Bormann et al. (2009).

**Question 3: How relate standard mb and mB\_BB to each other?**

Fig. 3.51 compares event standard magnitudes mb and mB\_BB, as measured at (CENC) on records of the China National Seismograph Network, and Fig. 3.52 the same but based on records of the German Regional Seismic Network (GRSN), analyzed by S. Wendt.



**Fig. 3.51** Regression relationships (standard and orthogonal) between standard mb and mB(BB), measured at CENC. Legend as in Fig. 3.49. Figure based on data published in Bormann et al. (2009).



**Fig. 3.52** Data and regression relationships (standard and **orthogonal = thick black line**) between the event standard mb and mB\_BB as determined with records of the German Regional Seismic Network (GRSN). The medium and light grey shaded bands limit the 1σ and 2σ standard deviation (SD) range of the OR relationship.

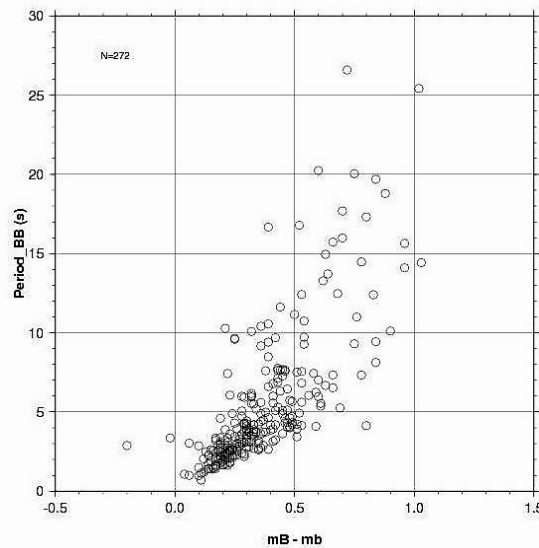
Both Fig. 3.51 and 3.52 show the same tendency, namely, that

- $mB\_BB$  is always larger than  $mb$  and differs the more the stronger the earthquake;
- the difference is smallest (around +0.1 m.u.) at magnitude 5, grows on average to about +0.5 m.u. around  $mB(BB) = 8$ , and may reach in individual cases about one m.u. (see also Fig. 3.31).

Moreover, the average SD values for the calculated event  $mb$  and  $mB$ , given in the lower right corner of Fig. 3.52, indicate that on average the measurement errors for  $mB$  (0.15 m.u.) are significantly smaller than for  $mb$  (0.21 m.u.), less smaller at shorter periods but much smaller for stronger earthquakes with  $V_{max}$  recorded at longer periods.

**Question 4: How does the large scatter of period readings in BB records affect the estimate of  $mB\_BB$ ?**

Fig. 3.48 revealed the large scatter of periods at which  $V_{max}$  is measured in all magnitude ranges. Yet, Fig. 3.53 shows a systematic trend, namely, that the difference  $mB\_BB-mb$  grows on average with the measured period which is related to the corner frequency of the seismic source spectrum and thus to earthquake magnitude itself. The difference between these two magnitudes is less than +0.3 m.u. when the  $mB$  periods fall into the range at which also  $mb$  is measured. The then still positive  $mB$  residuals are likely due to the larger relative bandwidth of the BB records as compared to WWSSN-SP records (see Chapter 4). Yet, the difference between  $mB\_BB$  and  $mb$  may reach 1 m.u. for great earthquakes measured at period around 15 to 25 s, which is in agreement with Fig. 3.45.



**Fig. 3.53** Relationship between the difference of  $mB(BB)-mb$  on the period at which the  $V_{max}$  for  $mB(BB)$  calculation was measured. The data have been measured on velocity-proportional records of the broadband German Regional Seismic Network (GRSN).

The large data scatter in Fig. 3.53 at comparable magnitude differences is most likely due to both stress-drop and thus corner-frequency differences of earthquakes with comparable moment release as well as to local station site effects. This, however, does not essentially bias

the stability and representativeness of the  $V_{\max}$  readings for reliable  $mB\_BB$  estimates, as the following case, observed on teleseismic station records of the GRSN, illustrates (according to Bormann et al., 2009):

- Earthquake off the east coast of Honshu, 18 January 2006, average network distance  $\Delta \approx 82^\circ$ , 16-station network  $mB\_BB = 6.06 \pm 0.23$  with an average  $T = 2.4 \pm 1.4$  s; station IBBN  $mB\_BB = 6.10$  at  $T = 0.7$  s and station BUG  $mB\_BB = 6.10$  at  $T = 5.0$  s; i.e., equal  $mB\_BB$  although  $T_{BUG} \approx 7 \times T_{IBBN}$ .

Thus we can conclude, that the variability of period measurements in BB records on often not very harmonic oscillations does not essentially bias the readings of  $V_{\max}$ . Generally, measurement errors of  $mB$  are significantly smaller than for short-period  $mb$  (compare also Figs. 3.23 and 3.24).

#### ***Question 5: How do deviations from the standard responses affect $mb$ and $mB\_BB$ ?***

$V_{\max}$  and  $T$  readings for  $mB\_BB$  will not be biased if the velocity-proportional passband range of the broadband records covers fully the range of periods within which  $mB\_BB$  is supposed to be measured, i.e., between 0.2 s to 30 s. In fact, however, periods below 1 s are rarely observed, usually for deep earthquakes only.

$mb$  measurements on narrowband short-period records are more strongly affected by the effective relative bandwidth, the steepness of gain roll-off beyond the corner periods of the passband and by the exact period at which the response reaches its maximum gain. Therefore, it is advisable to adhere to the WWSSN-SP simulation filter parameters prescribed for standard  $mb$ . Yet slight differences in the high-frequency roll-off, as they had originally been implemented in some analysis software, may be acceptable within the 0.1 m.u. tolerance range for standard magnitudes (see Fig. 11 and discussion in IS 3.3). For earthquakes with magnitudes  $M_w < 4$  and thus average  $f_c > 2$ -3 Hz of their source spectra (see Fig. 3.5) it would be acceptable to shift the record response to higher frequencies and steepen its slope to lower  $f$  so as to increase the SNR, provided that the spectral displacement plateau and/or velocity peak is still sampled. This would assure good scaling of the  $mb$  for small events to  $M_w$  and/or  $M_e$ . See also related discussions in sections 5.1.1 and 5.1.2 of IS 3.3).

#### ***Question 6: To what extent are differences between $mb$ and $mB\_BB$ due to frequency-dependent attenuation?***

Up to now, both  $mb$  and  $mB$  are calibrated with the Gutenberg and Richter (1956a)  $Q(\Delta, h)_{PZ}$  functions although the latter had been derived from readings with periods between about 2 and 20 s (Abe and Kanamori, 1979) and were intended to be used for calibrating mainly intermediate-period amplitude readings within this range. This raises the question whether it is acceptable at all to use these calibration functions for 1 s P-wave  $mb$  as well or whether the observed systematic differences between  $mb$  and  $mB$  are – at least partially – due to the neglectance of frequency-dependent attenuation. One should note, however, that Gutenberg and Richter obviously did not believe in the applicability of the frequency-dependent attenuation model for their problem of magnitude determination. Because by measuring  $A/T$  instead of  $A$  they compensated in fact largely for the exponential frequency-proportional increase of displacement amplitude attenuation by multiplying  $A$  with  $f = 1/T$ . Thus they

derived a linear model of attenuation proportional to  $\exp(-0.00006 L)$ , where  $L$  is the total length of the ray path from the station to the source. This explains the stability of the G&R  $Q(\Delta, h)$  functions, making them theoretically even equally applicable to 10 s data and 10 Hz data. Although this may not fully be true for frequencies above about 0.5 Hz, various studies have shown (see section 2.5.4.2 on *wave attenuation* in Chapter 2) that there is no clear empirical evidence for frequency-dependent  $Q$  for periods above 2 – 4 s and that for higher frequencies the expected exponential increase in displacement amplitude attenuation is strongly reduced by absorption-band effects. Moreover, it is a matter of fact that the observed differences between  $m_b$  and  $m_B$  depend chiefly on magnitude itself and that they can almost perfectly be explained by the magnitude-dependent shift of the corner frequency of the source spectrum towards lower frequency and the average decay of the velocity amplitudes for  $f > f_c$  with the first order (see Fig. 3.45).

With the same reasoning, the IASPEI standard calibration function for  $M_s$ , measuring  $A/T$  or nowadays directly velocity amplitudes for  $M_s\_BB$  in a wide range of periods is rather stable and reliable as well without accounting for frequency-dependent attenuation. When, however, measuring surface-wave displacement amplitudes in narrow bandpass ranges for  $M_s$  determination, as done in the Bonner et al. (2006)  $M_s(VMAX)$  procedure, then it is indeed indispensable to correct such amplitudes for frequency-dependent attenuation to make them on average compatible with  $M_s\_BB$  or the so-called Prague- $M_s$ . That this is indeed the case has been demonstrated by both Bonner et al. (2006) and Bormann et al. (2009).

Regrettably, reliable experimental data on frequency-dependent attenuation are still rare and model calculations may yield rather different results depending on whether one uses attenuation models with the **quality factor  $Q$**  depending on frequency for shorter periods or not. Therefore, until more reliable empirical spectral attenuation data for ground-motion velocities measurements become available, it is still justified to use the Gutenberg-Richter calibration functions for both  $m_b$  and  $m_B\_BB$  determinations.

### **Question 7: How reliable are the Gutenberg-Richter $Q(\Delta, h)_{PV}$ curves/tables?**

The Gutenberg and Richter (1956a) calibration curves  $Q(\Delta, h)_{PV}$  show many details (see Fig. 1a in DS 3.1), probably too many in view of the interpolation through rather noisy data measured with instruments of different response, often not well known gain and station site effects (see Veith, 1998) and less accurate estimates of epicentral distance and depth they had available in their days. Already efforts by Christoskov et al. (e.g., 1978, 1985 and 1991) for the development of a homogeneous magnitude system (HMS) based on carefully selected first-order stations over Eurasia and accounting for station site corrections showed that the average P-wave calibration curve for both short-period and medium-period amplitude readings from shallow events in the epicentral distance range between 20° and 100° is much smoother than the respective Gutenberg-Richter  $Q(\Delta)$ . The same holds for the distance-depth corrections derived by Veith and Clawson (1972), Lilwall (1987) and Rezapour (2003) for short-period P-waves. For shallow sources the differences between the  $P(\Delta)$  of Veith and Clawson (1972) and  $Q(\Delta)$  reach up to 0.3 m.u. in some limited distance ranges between 20° and 100°. Saul and Bormann (2007) came to the same preliminary result for new broadband  $m_B\_BB$  calibration functions. The latter authors additionally showed that the differences in the regional range between 5° and 20° epicentral distance may be even larger, up to about 0.8 m.u. around 10° (see Figure 16 in IS 3.3). These differences, especially in the regional distance range, may however differ from region to region. Further investigations are required

to clarify this. Standardized mB measurements will ease this task. With respect to the reliability of the medium-period  $Q(\Delta, h)_{PV}$  curves for deep earthquakes see also section 3.2.6.1 and with respect to the alternative Veith and Clawson  $P(\Delta, h)$  curves for 1 Hz P waves section 3.2.6.2. Both Nolet et al. (1998a) for their broadband mb and Rezapour (2003) for his short-period mb distance-depth correction terms use the values of scalar seismic moment  $M_0$  in the Harvard Centroid Moment Tensor catalog to calibrate the P-wave amplitudes.

### ***Still open problems:***

Despite the strong recommendation of the Committee on Magnitudes at the IASPEI General Assembly in Zürich (1967) to report the magnitude for all waves for which calibration functions are available, the ISC and NEIC determine no body-wave magnitudes from PP or S waves despite their merits discussed above and the fact that digital broadband records now allow easier identification and parameter determination of these later phases.

1) At distances beyond  $100^\circ$  the P-wave core shadow prevents to measure mb and mB. Yet, good PP readings from strong earthquakes are possible and calibration functions  $Q(\Delta, h)$  available for PPH and PPV up to  $170^\circ$ . According to Bormann and Khalturin (1975), mB determined from P and PP waves scale perfectly. The orthogonal regression is

$$mB(PP) = mB(P) + 0.05 \pm 0.15 \text{ m.u.} \quad (3.46)$$

If, however, short-period amplitude readings for P and PP are used instead, the orthogonal relationship is magnitude-dependent ( $mb(PP) = 1.25 mb(P) - 1.22$ ) and the standard deviation is much larger ( $\pm 0.26$  m.u.). This again testifies the greater stability of body-wave magnitudes based on medium- to long-period readings.

2) The suitability of large amplitude readings of PKP in the distance range of the core caustic around  $145^\circ$  and beyond for magnitude determinations has also been ignored so far (see 3.2.6.3 and EX 11.3).

3) No proper discrimination had been made in the past at the international data centers between data readings from different kinds of instruments or filters, although respective recommendations were made already at the joint IASPEI/IAVCEI General Assembly in Durham, 1977. However, with the step-wise introduction of the new IASPEI standards since 2010 for some of the most widely used magnitudes both the measurement parameters as well as the related unique nomenclature for measured amplitudes and magnitudes have now been fixed (see Table 4 in IS 3.3). On the other hand, this necessitates that magnitude data resulting from non-standard procedures, which can – on average - not reproduce for the same type of magnitude the standard results within 0.1 m.u., be reported with a specified nomenclature. First proposals have been made in IS 3.2 and are expected to be refined in future.

## **3.2.6 Other amplitude, period, intensity or tsunami based magnitude scales (P. Bormann)**

Below we describe several other complementary procedures for magnitude estimation. They are not (yet) based on IASPEI recommended standards but are (or may become) also useful for applications in seismological practice.



### 3.2.6.1 Broadband and spectral P-wave magnitude scales

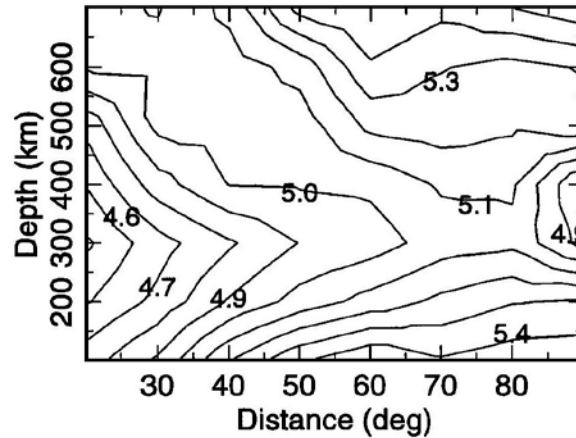
#### *Broadband P-wave magnitude scale for intermediate and deep earthquakes*

A calibration function  $Q_b(\Delta, h)^{NKC}$  has been derived by Nolet et al. (1998a). It is based on  $V_{\max}$  measurement on broadband velocity recordings of P waves (bandpass between 0.01 and 2 Hz) and applicable to intermediate and deep earthquakes in the ranges  $20^\circ \leq \Delta \leq 90^\circ$  and  $100 \text{ km} \leq h < 700 \text{ km}$ . For deriving this calibration function, the measured amplitudes have been normalized to a scalar seismic moment  $M_0 = 10^{18} \text{ Nm}$ , which corresponds to a **short-period**  $mb = 5.7$  according to a relationship  $mb = (\log M_0 - 4.3)/2.4$  for deep earthquakes published by Giardini (1988) and to  $Mw = 5.9$  according to the IASPEI standard  $Mw$ - $M_0$  relationship  $Mw = \log(M_0 - 9.1)/1.5$ .

The  $Q_b(\Delta, h)^{NKC}$  plot looks much smoother (Fig. 3.54) than the Gutenberg and Richter (1956a) plot of  $Q(\Delta, h)_{PZ}$  (see Figure 1a in DS 3.1). Even some of the larger details in the Gutenberg-Richter plot are missing, yet the isolines follow the main feature of  $Q(\Delta, h)_{PZ}$  (around 300 to 400 km depth) and the trend with distance, have small standard deviations and no bias with depth with respect to  $\log M_0$  when normalized to 0 for shallow events (Nolet et al., 1998b).

Quantitatively, the values of  $Q_b(\Delta, h)^{NKC}$  are generally lower than those of  $Q(\Delta, h)_{PZ}$  because Nolet et al. (1998a) wrote their  $mb$  formula in the way as we wrote also (3.37a) for  $Ms\_BB$  and (3.45a) for  $mB\_BB$ , i.e., by including  $-\log 2\pi = -0.8$  into the constant. But it remains an average difference of about -0.3 m.u. with quite some scatter. The scatter may be due to the many local details in the G-R plot that could - partially at least - be related to uncorrected station site effects (Veith, 1998) (see discussion in section 3.2.5.2 related to Question 7). The negative average residual with respect to  $Q(\Delta, h)_{PZ}$ , however, is obviously due the scaling of  $mb(\text{Nolet})$  to  $mb = 5.7$  (or  $Mw = 5.9$  for  $M_0 = 10^{18} \text{ Nm}$ ), which on average corresponds according to Bormann et al. (2009) to an  $mB\_BB = 6.0$  (see also Fig. 3.53 and the average difference of  $mB$ - $mb$  of about 0.3 m.u. for periods of P around 3 s in broadband records. According to Nolet et al. (1998a) the dominant periods of the maximum velocity pulses of deep earthquakes in broadband records fall in the range between 1 – 4 s but are occasionally even larger. Thus they are longer than the dominant periods around 1 s of the short-period P waveforms on which the amplitudes for  $mb$  are commonly read. Therefore, in order to make on average the numbers given on the isolines of Fig. 3.54 comparable with the Gutenberg-Richter (1956a)  $Q$ -values, which were derived for scaling medium-period  $mB$  measurements, one would need to correct for the downscaling to short-period  $mb$ .

No results of a systematic comparison of  $mb(\text{Nolet})$  with standard  $mB\_BB$  for identical  $V_{\max}$  readings have become known to us so far. This, however, would be necessary and the reasons for differences better be understood before one can judge on the potential advantage (or not) of replacing for earthquakes deeper than 100 km the Gutenberg-Richter  $Q(\Delta, h)_{PZ}$  by  $Q_b(\Delta, h)^{NKC}$ . Only after a rigorous comparison of  $mb$  and  $mB\_BB$  values calculated for deep earthquakes by using both the Gutenberg-Richter and the Nolet et al.  $Q$ -values and their confrontation with independent accurate and stable magnitude measurements such as  $Mw$  one might be able to decide which calibration data are better suited for calculating the two standard body-wave magnitudes  $mb$  and  $mB\_BB$  for deep earthquakes. Such an investigation should be encouraged.



**Fig. 3.54**  $Q_b(\Delta, h)^{NKC}$  calibration values, averaged over  $20^\circ$  and 200 km depth range. Copy of Fig. 3 in Nolet et al. (1998), *Geophys. Res. Lett.*, **25**, 9, p. 1453; © American Geophysical Society.

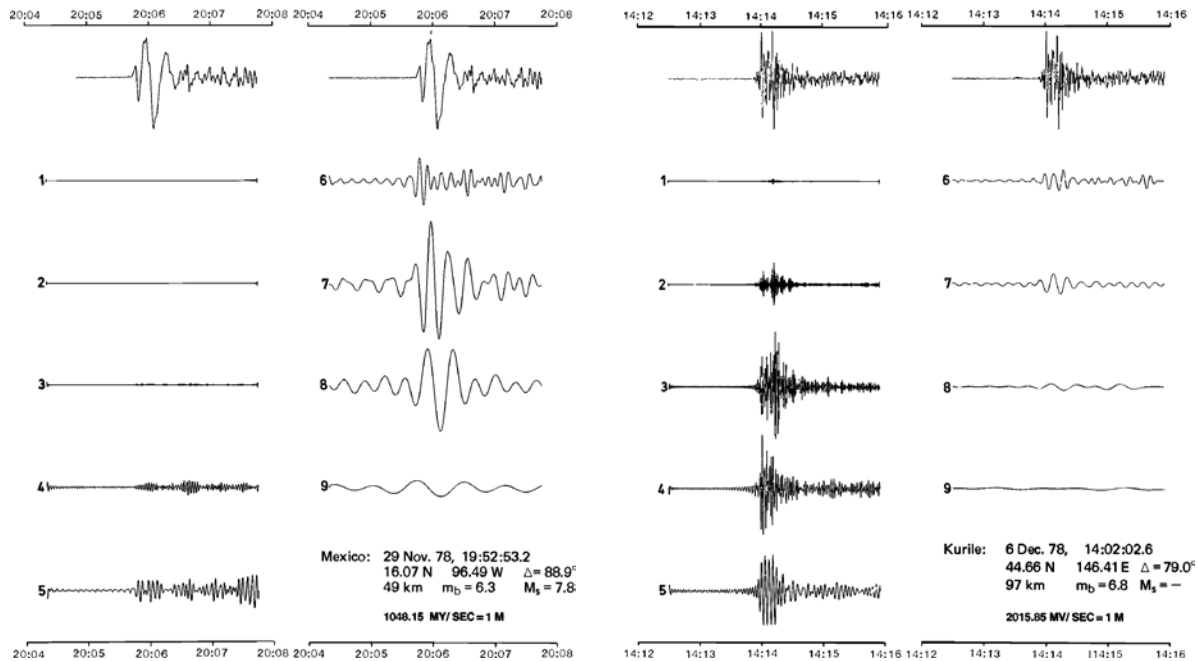
### *Spectral P-wave magnitudes*

Duda and Kaiser (1989) recommended the determination of spectral magnitudes based on measurements of largest amplitudes in one-octave filtered digital broadband records of P waves. Fig. 3.55 show such bandpass records of two earthquakes with comparable magnitude  $m_b$ , recorded at comparable distance, but having different source depth and mechanism. Accordingly, the amplitudes in different spectral ranges differ a lot. This is due to regional differences in ambient stress conditions and related stress drop.

Duda and Yanovskaya (1993) calculated theoretical spectral amplitude-distance curves based on the IASP91 velocity model (Kennett and Engdahl, 1991) using two different attenuation models aiming at magnitude calibration of spectral amplitude measurements (see Fig. 2.39 in Chapter 2). This effort is a response to the “saturation” problems discussed above. Combining the spectral magnitude estimates yields smoothed average values of the radiated seismic spectrum, its spectral plateau, corner frequency and high-frequency decay and thus of  $M_0$  and stress drop of the given event. Thus one may draw inferences on systematic differences in the prevailing source processes (e.g., low, normal or high stress drop), related ambient stress conditions in different source regions but also on the amount and relative share of radiated high and low frequencies and thus of the potential of the earthquake to cause either strong shaking damage or generate a tsunami.

This, however, has so far not yet been the main concern of common seismological routine practice, which aimed at providing just a simple one (or two) parameter size-scaling of seismic events for general earthquake statistics and hazard assessment. Rather, spectral magnitudes were more considered a research issue, which could be tackled even more directly by calculating both  $M_0$  and the radiated seismic energy  $E_s$  proper, by estimating the corner frequency and shape of the overall source spectra from the corrected broadband record spectra and by determining and interpreting jointly both energy magnitude  $M_e$  and moment magnitude  $M_w$  and their difference (see IS 3.5). Good  $M_w$  and  $M_e$  estimates can nowadays already be done fully automatic within about 10-20 min after origin time and allow essentially the same inferences to be drawn as from the multi-bandpass filtering of broadband records (see, e.g., Choy and Kirby, 2004; Di Giacomo et al., 2010a; Bormann and Di Giacomo, 2011).

Therefore, this educationally very appealing concept of spectral magnitudes has not yet found broad application in observatory practice. A fine example is presented in Fig. 3.55.



**Fig. 3.55** Examples of broadband digital records proportional to ground velocity of the P-wave group from two earthquakes with  $m_b = 6.3$  and  $h = 49$  km (left), and  $m_b = 6.8$  and  $h = 97$  km (right), respectively. The earthquakes occurred in different source regions (Mexico and Kuriles). Below their BB records (uppermost traces) their one-octave bandpass-filtered outputs are plotted. The numbers 1 to 9 on the filtered traces relate to the different center periods between 0.25 s (1) and 64 s (9) in one-octave distance. Note that the somewhat “weaker” event on the left has its maximum recorded ground velocity in trace 7, corresponding to a center period of 16 s. In contrast, the  $m_b$ -wise stronger event on the right, for which one would on average expect an even longer corner period, has its  $V_{\max}$  at 1 s (trace 3). Thus, the 1-Hz spectral magnitude in the case of the Kurile earthquake would be much larger than for the Mexico earthquake, hinting to strong felt shaking and potential damage of exposed objects, not, however in the case of the Mexico earthquake with a very small amount of 1-Hz energy released. (Copied from Duda, 1986; with permission of Schweizerbart [www.schweizerbart.de](http://www.schweizerbart.de)).

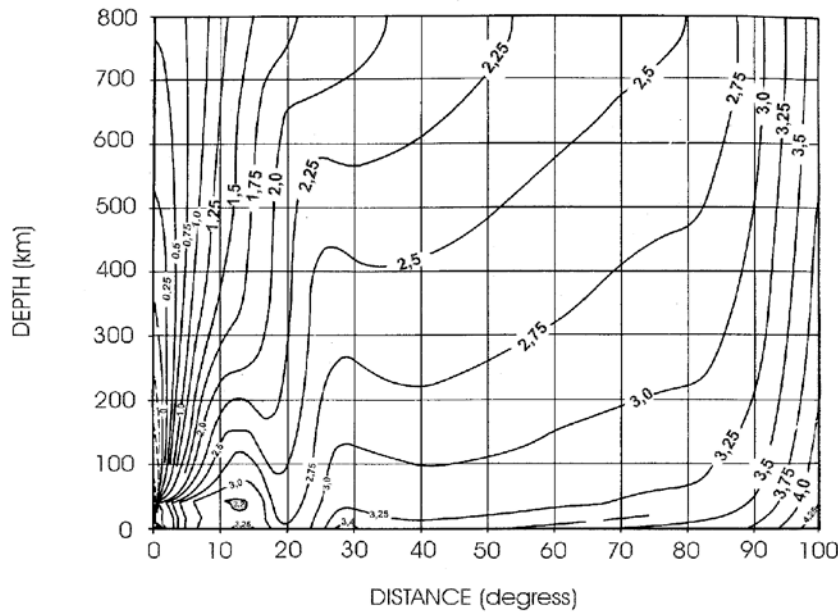
### 3.2.6.2 1-Hz P-wave magnitude $m(P)$ and $m_b(IDC)$

Veith and Clawson (1972) developed a calibration function, termed  $P(\Delta, h)_{SP}$  (see Fig. 3.56), using short-period WWSSN vertical-component P-wave records of underground nuclear explosions, thus avoiding data scatter due to uncorrected source radiation pattern effects. The  $P(\Delta, h)_{SP}$  plot looks much smoother than the  $Q(\Delta, h)_{PZ}$  curves published by Gutenberg and Richter (1956a) and resemble, also at depth, much more an inverse  $A-\Delta$  relationship for short-period P waves as depicted in Figs 2.39 or 3.34. However, using only very shallow-depth sources, the calibration curves at depth are empirically not well constrained. They have been model-derived using still debatable attenuation models and therefore differ often significantly from both the Gutenberg and Richter (1956a) and the Nolet et al. (1998a) calibration curves

for vertical component P waves (compare with Figure 1a in DS 3.1 and Fig. 3.54 above, respectively). Yet, for shallow events mb(P) values agree well with mb(Q) (according to Veith, 2001, with an average difference of -0.03 m.u.) but have less scatter. Therefore, no wonder that the International Data Center (IDC) of the Comprehensive Test-Ban Treaty Organization (CTBTO) uses  $P(\Delta, h)_{SP}$  as standard calibration curve for calculating its mb(IDC) values (see Chapter 15).

For deeper events, however, mb(P) is systematically lower than mb(Q) (up to about 0.4 magnitude units at 450 to 640 km depth) due to a different attenuation law assumed in the upper mantle and transition zone (Veith, 2001). For more details on pros and cons see Nolet et al. (1998b) and Veith (1998). Also note, that deviating from the use of the Gutenberg-Richter  $Q(\Delta, h)_{PZ}$  functions, the calibration values given in Fig. 3.56 have to be applied to maximum P-wave peak-to-trough (2A) displacement amplitudes in units of  $\text{nm} = 10^{-9} \text{ m}$  (instead of peak-to-trough/2 amplitudes in  $\mu\text{m} = 10^{-6} \text{ m}$ ). Accordingly, the Veith and Clawson formula reads:

$$\text{mb(P)} = \log (A_{\text{max;peak-to-trough}}/T) + P(\Delta, h) \quad (3.47)$$



**Fig. 3.56** Calibration functions  $P(\Delta, h)$  for mb determination from narrow-band vertical-component short-period records with peak displacement magnification near 1 Hz (WWSSN-SP characteristic). (Modified according to Veith and Clawson (1972), Magnitude from short-period P-wave data, BSSA, **62**, 2, p. 446, © Seismological Society of America).

The Veith-Clawson magnitude calibration functions are officially used by the IDC in Vienna for mb determination although the IDC filter applied to the digital velocity-proportional broadband data prior to the amplitude measurements for mb results in a displacement response peaked around 3.4 Hz (Granville et al., 2005) instead of around 1 Hz, as required for the use of  $P(\Delta, h)$  (Veith, 2001). According to the spectral  $\log A$ -D curves calculated by Duda and Yanovskaya (1993) for the modified PREM attenuation model (see Fig. 2.39 in Chapter 2),  $\log A$  is in the distance range between  $10^\circ$  and  $100^\circ$  at 4 Hz about 0.15 to 0.5 units smaller

than at 1 Hz. Thus, the use of  $P(\Delta, h)$  in conjunction with the IDC filter response is physically not correct and tends to systematically underestimate  $m_b$ . This is further aggravated by the fact that the IDC determines  $A_{\max}$  within a time window of only 5.5 s after the P onset (Granville et al., 2005). This heuristic procedure, although very suitable for a best possible discrimination between earthquakes and underground explosions on the basis of the  $m_b$ - $M_s$  criterion (see 11.2.5.2??), is not appropriate for proper earthquake scaling, at least for events with magnitudes larger than 6 and thus average rupture durations of more than 6 s [ see relationship (3.3)].

According to the above, the difference between  $m_b(\text{IDC})$  and other  $m_b$  values calculated with a more low-frequent response and longer measurement-time window will grow with magnitude. According to Bormann et al. (2007), based on data kindly provided by S. Wendt (2005), the difference between NEIC  $m_b(\text{PDE}) - m_b(\text{IDC})$  is on average:

- 0.06 m.u. for  $m_b(\text{PDE}) < 4.0$
- 0.37 m.u. for  $m_b(\text{PDE}) = 4.0-4.9$ ,
- 0.48 m.u. for  $m_b(\text{PDE}) = 5.0-5.9$ ,
- 0.61 for  $m_b(\text{PDE}) \geq 6$ ,

and reached 1.5 m.u. for the great 2004 Mw9.0 Sumatra-Andaman earthquake.

Granville et al. (2002) analyzed 10 medium-size earthquakes in the depth range 0 km to 530 km and with magnitudes  $m_b$  between 6.4 and 6.8 according to the PDE (Preliminary Determination of Epicenters) reports of the United States Geological Survey (USGS) and 13 underground nuclear tests (UNTs) with PDE magnitudes  $m_b$  between 4.6 and 6.1. They compared these data, which were derived from simulated WWSSN-SP records, by using the traditional procedure of  $m_b$  determination based on the Gutenberg-Richter Q-functions, with a) the  $m_b$  for the same WWSSN-SP data but calibrated with the Veith-Clawson relationship; b) the body-wave magnitudes reported in the REB (Reviewed Event Bulletin) of the (preliminary) PIDC based on the IDC filter with peak magnification around 3.4 Hz. For comparison of the responses see Figure 4 in Granville et al., 2005, or Figure 3 in IS 3.3).

From this study the following conclusions were drawn:

- The agreement between  $m_b(Q)$  (Gutenberg-Richter) and  $m_b(P)$  (Veith-Clawson) based on WWSSN-SP data was still reasonably good for shallow earthquakes (average difference  $m_b(Q) - m_b(P) = 0.2$ );
- For underground explosions (only shallow-depth events!) the agreement was even better (average  $m_b(P) - m_b(Q) = 0.09$ );
- Yet, the average discrepancy between  $m_b(P)$  and  $m_b(\text{PIDC/REB})$  for earthquakes is much larger (0.5 magnitude units), although the latter are also scaled with the Veith-Clawson calibration functions. For 63% of the earthquake observations the difference was at least 0.4  $m_b$  units, and several of them had even an  $m_b$  offset greater than 1 magnitude unit!;
- In contrast, the average discrepancy between  $m_b(P)$  and  $m_b(\text{PIDC/REB})$  for explosions is 0.0 and 75% of the observations fall between  $-0.1$  and  $+0.1$  m.u.;
- The PIDC (now IDC in Vienna) procedure is adequate for  $m_b$  determination of underground nuclear explosions (UNE) and for the discrimination between UNE and

earthquakes on the basis of the mb-Ms criterion but it is not compatible with mb(PDE) or IASPEI standard mb.

### 3.2.6.3 Short-period PKP-wave magnitude

Calibration functions  $Q(\Delta, h)_{\text{PKP}}$  for short-period amplitude and period readings from all three types of direct core phases (PKPab, PKPbc and PKPdf) have been developed by Wendt (see Bormann and Wendt, 1999; also explanations and Figure 3 in DS 3.1). These phases appear in the distance range  $\Delta = 145^\circ - 164^\circ$  (see Fig. 3.34, Figs. 11.62-63 ??? in Chapter 11 and Figure 1 in EX 11.3) with amplitude levels comparable to those of P waves in the distance range  $25^\circ < \Delta < 80^\circ$ . Many earthquakes, especially in the Pacific (e.g., Tonga-Fiji-Kermadec Islands) occur in areas with no good local or regional seismic networks. Often these events, especially the weaker ones, are also not well recorded by more remote stations in the P-wave range but often excellently observed in the PKP distance range, e.g., in Central Europe. This also applies to several other event-station configurations. Available seismic information from PKP wave recordings could, therefore, improve magnitude estimates of events not well covered by P-wave observations at distances below  $100^\circ$ .

Earlier investigations on the use of PKP phases for magnitude determination have been published by Miyamura (1974), Mizone, (1977), Tittel (1977), Janský et al. (1977), Kowalle et al. (1983) and Janský and Kvasnička (1992).

### 3.2.6.4 High-frequency moments and magnitudes

Koyama and Zheng (1985) developed a kind of short-period seismic moment  $M_1$  which is related to the source excitation of short-period seismic waves and scaled to mb according to

$$\log M_1 = 1.24 \text{ mb} + 10.9 \quad (\text{with } M_1 \text{ in } J = \text{Nm}). \quad (3.48)$$

They determined  $M_1$  from WWSSN short-period analog recordings by applying an innovative approximation of spectral amplitudes

$$Y(f) = 1.07 A_{\text{max}} (\tau/f_0)^{1/2} \quad (3.49)$$

with  $A_{\text{max}}$  - maximum amplitude,  $f_0$  - dominant frequency and  $\tau$  - a characteristic duration of the complicated wave-packets. They analyzed more than 900 short-period recordings from 79 large earthquakes throughout the world in the moment range  $7.5 \times 10^{17} \leq M_0 \leq 7.5 \times 10^{22} \text{ Nm}$ .  $M_1$  did not saturate in this range!

More recently, Atkinson and Hanks (1995) proposed a high-frequency magnitude scale

$$\mathbf{m} = 2 \log a_{\text{hf}} + 3 \quad (3.50)$$

with  $a_{\text{hf}}$  as the high-frequency level of the Fourier amplitude spectrum of acceleration in cm/s, i.e., for  $f \gg f_c$ . They use average or random horizontal component accelerometer amplitudes at a distance of 10 km from the hypocenter or from the closest fault segment.  $\mathbf{m}$  has been scaled to the moment magnitude  $\mathbf{M} = M_w$  for events of average stress drop in eastern North America and California. When  $\mathbf{M}$  is known,  $\mathbf{m}$  is a measure of stress drop. For large pre-

instrumental earthquakes  $m$  can more reliably be estimated than  $M$  from the felt area of earthquake shaking (see 3.2.6.6). When used together,  $m$  and  $M$  provide a good index of ground motion over the entire engineering frequency band, allow better estimates of response spectra and peak ground motions and thus of seismic hazard.

### 3.2.6.5 Earthquake early-warning magnitudes

In recent years great attention has been paid to the development of earthquake early warning systems (EEWS) which allow – besides very quick event location - near real-time magnitude estimates from the very first few seconds of acceleration, velocity or displacement records (e.g., Nakamura, 1988. Nakamura and Saita, 2007; Espinoza-Aranda et al., 1995; Teng et al., 1997; Wu et al., 1997; Allan and Kanamori, 2003; Kanamori, 2005; Simons et al., 2006; Zollo et al., 2006; Gasparini et al., 2007). Based on such data rapid public alarms may be issued and/or automatically triggered risk mitigation actions taken between the detection of a strong earthquake at near stations and the arrival of strong ground motion at locations slightly farther away. EEWS aim at minimizing the area of so-called “blind zones” which are left without advanced warning before the arrival of the S waves which have usually the largest strong-motion amplitudes. This necessitates very dense and robust seismic sensor networks within a few tens of km from potentially strong earthquake sources. Such dense networks are at present available or proposed for only a very regions, e.g. in California, Japan, Taiwan, Turkey (Istanbul) and Italy. Their principles of rapid magnitude estimates are rather different from those mentioned above and below and the data analysis from such systems is largely based on much debated concepts such as the hypothesis of the deterministic nature of earthquake rupturing. Olson and Allen (2005), e.g., claim to be able to estimate with an average absolute deviation of 0.54 magnitude units the size of earthquakes up to magnitude 8 from the maximum period of the first arriving P waves within the first 4 s from many low-pass filtered velocity records within 100 km from the epicenter. However, Rydelek and Horiuchi (2006), who analyzed waveform data recorded by the Japanese Hi-net seismic network (see Chapter 8, section 8.7.3), could not confirm that such a dominant frequency scaling with magnitude exists. Also Kanamori et al. (1997), together with Nakamura (1988) one of the fathers of this idea, expressed much more caution about the prospects of this method after they had run together with Wu an extensive experiment with the Taiwan Early Warning System which has presently a 22-s reporting time for first location and magnitude estimates with magnitude uncertainty of  $\pm 0.25$  m.u. (Wu and Kanamori, 2005). To reduce this reporting time even further down to about 8 s and thus the “blind zone”, they used only the records of the first 8 station within less than 21 km epicentral distance to estimated the magnitude via the magnitude dependence of the average period  $\tau_c$  of the P-wave, as well as the peak displacement amplitudes, if  $> 0.1$  cm, within the first 3 s of the low-pass filtered displacement signal after the P-wave onset. Wu and Kanamori (2005) summarized their experience as follows: “... the slip motion is in general complex and even a large event often begins with a small short-period motion, followed by a long-period motion. Consequently, it is important to define the average period ...during the first motion.” However, after applying the  $\tau_c$  concept to the Taiwan EWS they concluded: “...For EWS applications, if  $\tau_c < 1$  sec, the event has already ended or is not likely to grow beyond  $M > 6$ . If  $\tau_c > 1$  sec, it is likely to grow, but how large it will eventually become, cannot be determined. In this sense, the method provides a **threshold warning**”. Moreover: “If  $\tau_c > 1$  sec, the event is potentially damaging. If it is larger than 2 sec, the event is almost certainly damaging. Combining this information with other data, such as the initial velocity and displacement amplitudes, would allow the damage potential of the event to be assessed more accurately.” More can indeed not

be expected to be said within the first 3 s of a record, especially not for strong and great earthquakes. Only for earthquakes with  $M \leq 6$  and thus total rupture durations that are according to formula (3.5) on average not more than about twice the  $\tau_c$  measurement time windows of Kanamori (3 s) or Olson and Allen (4 s), these new concepts seem to yield reasonably good magnitude estimates, comparable with those provided by tsunami early warning systems (TEWS) in the case of strong to great tsunamigenic earthquakes within some 5 to 30 min (e.g., Lomax and Michelini, 2012). This is illustrated by short reference to and discussion of several representative EEWS.

The Nakamura UrEDAS (Urgent Earthquake Detection and Alarm System) predicts the magnitude from the dominating period  $T_d$  within a few seconds of the initial P-wave motion. According to Nakamura and Saita (2007) these dominating periods, when observed at hypocentral distances between about 50 to 850 km, increase for earthquakes with  $M_{JMA}$  3 to 7 from an average of about 0.2 s to 4 s. However, due to the large scatter of  $T_d$  by a factor of about 3 measured at individual stations for the same event, the event magnitude  $M_{JMA}$  can only be estimated with a possible error of about 0.5 m.u.

The ElarmS system (Allen and Kanamori, 2003), an adaptation of UrEDAS for Southern California, uses TriNet and California Integrated Seismic Network stations (see IS 8.4, and Simons et al., 2006). For estimating the magnitude it measures the **predominant period**  $T_{max}^P$  (between 0.1 and 3 s) in broadband velocity records within the first few seconds (1 to 4 s, depending on event size) after the P-wave onset. Yet, again due to the large variability of the individual station  $T_{max}^P$  (up to a factor of about 10) the average absolute error in the estimated magnitude is  $\pm 0.7$  m.u. when measured only at the closest station to the epicenter. However, when the closest 10 stations within 100 km of the epicenter are used, this error drops to  $\pm 0.35$  m.u. Allen and Kanamori (2003) give two regression relationships for estimating the magnitude via  $T_{max}^P$ , one for events with magnitudes between 3 and 5, the other one for magnitudes between 4.5 and 7.

Grecksch and Kämpel (1997) estimated the magnitudes using the rise time of the first complete peak amplitude, the **predominant period**, and the related Fourier amplitude of the initial part of strong-motion signals. Yet, with this approach the magnitude uncertainty may even be as large as  $\pm 1.35$  m.u. with a single accelerogram and  $\pm 0.5$  m.u. with more than eight accelerograms.

A different approach had been taken by Wu et al. (1998). Using strong-motion records from moderate size earthquakes with  $5.0 \leq M_l \leq 6.5$  and 10 s waveforms after the P-wave onset they derived an  $M_{l10}$  estimator from the strong-motion signal duration which allows to estimate local magnitude  **$M_l = 1.28M_{l10} - 0.85 \pm 0.13$**  with a correlation coefficient of 0.96. The slope of 1.28 indicates that small earthquakes require smaller and bigger earthquakes larger adjustments at 10 s to be made. With this procedure the Taiwan Central Weather Bureau determines epicenters and rather reliable magnitudes in about 20 to 30 s after the occurrence of moderate size earthquakes.

For larger crustal earthquakes Wu and Teng (2004) developed another method of estimating the moment magnitude with a standard deviation of about  $\pm 0.30$  m.u. up to magnitudes of about 7.5 by integration of near source strong-motion amplitudes over the **whole rupture duration**. Yet, such magnitude estimates require more time, and are available only within some 30 to 60 s after origin time. This time lag is not suitable for advanced early warning anymore, however quite adequate for rapid response services (RRS).



In order to reduce the time lag Wu and Zhao (2006) again dealt with estimating the magnitude by using the **P-wave amplitudes** within the first three seconds of the record. Applied to earthquakes in Southern California they found out that their “*Pd magnitudes*” agree with catalog magnitudes with a **standard deviation of  $\pm 0.18$  m.u. for events with magnitudes  $< 6.5$**  but that they saturate for larger ones and can not be used to estimate magnitudes  $\geq 6.7$ . The same problem of saturation of *Pd* amplitudes they had observed already for large earthquakes in Taiwan (Wu et al., 2006), thus providing evidence that the deterministic earthquake initiation model of Olson and Allen (2005) is not applicable to large earthquakes with magnitudes up to 8 and more, with rupture durations up to several 100 s.

In Italy, with very dense accelerometer networks near to major active faults, a procedure has been developed and tested by Zollo et al. (2006) which determines for both P and S waves the distance corrected logarithm of peak displacement ( $\log(\text{PGD}_t^{10\text{km}})$ ) within the first 2 s after the onset of P and S for events in the magnitude range between 4 and 7.3 (93% in the range 4 to 6). Most of the records have been made at distances less than 20 km from the source. The authors give weighted standard for their  $\log\text{PGD}$  estimates of  $\pm 0.13$  (for S) to  $\pm 0.22$  (for P).

A very different approach has been taken by Simons et al. (2006). They determine the earthquake magnitude via a multiscale wavelet analysis from the first 3-4 s of the incoming P wave and analyzed 2272 seismograms recorded by the Southern California TriNet array at epicentral distances up to 150 km in the magnitude ranges from 3 to 5 and  $> 5$  to 7.3, respectively, as in Allen and Kanamori (2003). Yet also they could predict the true magnitude only within approximately one unit. These authors correctly state that “...the nature of wave propagation (which includes the effects on the P-wave amplitude due to focusing, attenuation and site amplification) as well as the complexities of the source mechanism, and its orientation with respect to the measurement stations, are all such that any predictive relationships between the observables gleaned from the P wave and earthquake magnitude are statistical at best.”

This notwithstanding, much of the average potential as well as the principal shortcomings of the above EEW magnitude procedures could have been predicted and wrong claims as those by Olson and Allen (2005) be avoided by taking into account the approximate relationship (3.5) as well the average seismic source spectra in Fig. 3.5 of this Chapter.

According to relationship (3.5) the average rupture duration of magnitude  $M = 3$  events is  $\approx 0.1$  s, for  $M = 5 \approx 1.6$  s, for  $M = 6 \approx 6$  s, and for  $M \geq 7 \approx 25$  s up to about 400 s for  $M = 9$ . Moreover, the large and great earthquakes are usually very individual multiple rupture processes with complicated and often non-symmetric source-time functions (see, e.g., Figs. 3.5, 3.9 3.10, 3.13 in this Chapter). This may even require to approximate the moment-rate release function by multiple point sources to get correct  $M_w$  estimates (e.g., as Tsai et al., 2005, for the great 2004 Sumatra-Andaman earthquake). In view of these facts, to speak of a deterministic nature of earthquake ruptures which allows to predict their size up to magnitude 8 from data within the first 4 s of a seismic records is daring. At best one can speak of a deterministic component in the rupture process. Suitable measurement parameters related to this deterministic component are the average **rupture duration**, uncertain by a factor of about 2-3 for a given event and magnitude (see, e.g., Fig. 3.5), and the approximate average corner frequency  $f_c$  of the source spectrum, also with a variability range at a given magnitude due to variations in stress drop and/or rupture velocity.

Moreover, magnitude is supposed to be an estimator of released seismic energy. The largest amount of energy is released around  $f_c$ . According to Fig. 3.5, right-hand panel,  $f_c$  is on average for  $M_w = 3 \approx 10$  Hz, for  $M = 5 \approx 1$  Hz, for  $M = 6 \approx 0.3$  Hz, and the respective corner periods for  $M = 7, 8$  and  $9 \approx 25$  s, 100s and 400 s, respectively. But this means that up to  $M = 6$  the dominating radiated periods are already contained in the first few second of the P-wave record. And since the rupture duration for  $M = 6$  is on average only about 6 seconds, the related amplitude maximum will also appear within a time window of 3-4 s after the P-wave onset, a symmetric moment rate release function provided. Thus there is no wonder, why the EEW magnitude procedures based on  $P_d$  and/or  $\tau_c$  work reasonably well up to magnitudes 6-6.5, begin to saturate, however, for stronger earthquakes. And then it is understandable, why the strong-motion duration-based procedure by Wu and Teng (1998) works fine up to 7.5 when integrating over time windows between 30 s and 60 s.

And beyond, this principle discussion explains also why procedures as proposed, e.g., by Hara (2007 a and b), Bormann and Saul (2009), and Lomax and Michelini (2009 a and b) do not saturate at all. They relate maximum, summed-up or integrated amplitudes or, as Lomax and Michelini (2011 and 2012) the dominant period  $T_d$  (up to 30 s and more) to estimates of the rupture duration. Lomax and Michelini (2012) even measure  $T_d$  by the peak of the  $\tau_c$  algorithm of Nakamura (1988) and Wu and Kanamori (2005), applied with a 5 s sliding time-window from 0 to 55 s after the P arrival on velocity seismograms. I.e., essentially the same measurement parameters are looked for as by the EEW community, but they avoid saturation by adopting the measurement time window after P to the growing rupture duration and the bandwidth of the record so as to surely cover  $f_c$  and thus  $P_d$ . And additionally, as by Wu et al. (1998) for magnitudes up to 7.5, the rupture duration itself is taken into account. Yet such principle considerations can not be found in any of the other above referenced EEW papers.

### 3.2.6.6 Macroseismic magnitude

In order to apply the useful concept of instrumentally determined magnitudes for earthquake size classification, statistics and seismic hazard assessment also to earthquakes that occurred in the pre-instrumental era, relationships between earthquake magnitudes and macroseismic parameters determined in the instrumental era had to be developed. Some of these efforts aimed at developing specifically magnitude scales which are best suited for earthquake engineering assessment of potential damage and thus seismic risk. According to Grünthal (2011) “intensity data give a surprisingly robust measure of earthquake magnitudes.” These efforts go in two directions: by relating  $M$  to macroseismic intensity  $I$  and/or shaking area  $A_I$  or by focusing on the high-frequency content of seismic records.

*Macroseismic magnitudes*,  $M_{ms}$  are particularly important for the analysis and statistical treatment of historical earthquakes. Gutenberg and Richter (1942) published already a few years after the introduction of the instrumental earthquake magnitude by Richter (1935) a first empirical relationship with the epicentral intensity  $I_0$  followed by Kawasumi (1951) who related his magnitude scale  $M_K$  to the **intensity  $I$  in the JMA scale** (see Chapter 12) **at the 100 km distance**, following Richter’s definition of  $M_I$  as closely as possible. This approach is physically quite reasonable because for most earthquakes a distance of 100 km is already the far field and source finiteness can be ignored. The Kawasumi intensity-magnitude conversion formula reads:

$$M_K = 0.5 I_{100} + 4.85. \quad (3.51)$$

This approach was further developed by Rautian et al. (1989).

Other approaches related either the maximum observed intensity  $I_{\max}$  (e.g., Topozada, 1975) or the epicentral intensity  $I_0$  to  $M$  (e.g., Karnik, 1969). Differences between  $I_0$  and  $I_{\max}$  are expected to be small, but there have been instances where  $I_{\max}$ , due to extreme local site effects exceeded  $I_0$  as much as two degrees (Tinti et al., 1987). Therefore, relationships based on  $I_0$  are nowadays generally preferred to  $I_{\max}$ -based relationships. On the other hand,  $I_0$ -based definitions implicitly assume the point source model and must be often misleading. Of course, with historic catalogs, there is usually no other way. There are three main approaches to compute macroseismic magnitudes:

- 1)  $M_{ms}$  is derived from the epicentral intensity  $I_0$  (or the maximum reported intensity,  $I_{\max}$ ) assuming that the earthquake effects in the epicentral area are more or less representative of the strength of the event (e.g., Karnik, 1969);
- 2)  $M_{ms}$  is derived from taking into consideration the whole macroseismic field, i.e., the size of the shaking is related to different degrees of intensity or the total area of perceptibility,  $A$  (e.g., Topozada, 1975);
- 3)  $M_{ms}$  is related to the product  $P = I_0 \times A$  which is nearly independent of the focal depth,  $h$ , which is often not reliably known (Toperczer, 1953).

Accordingly, formulae for  $M_{ms}$  have the general form of

$$M_{ms} = a I_0 + b, \quad (3.52)$$

or, whenever the focal depth  $h$  (in km) is known, which strongly controls  $I_0$ ,

$$M_{ms} = c I_0 + \log h + d, \quad (3.53)$$

or, when using the shaking area  $A_{li}$  (in  $\text{km}^2$ ) instead,

$$M_{ms} = e \log A_{li} + f \quad (3.54)$$

with  $A_{li}$  in  $\text{km}^2$  shaken by intensities  $I_i$  with  $i \geq \text{III}$ , ..., VIII, respectively.

Since dominating source depth and intensity-attenuation conditions may vary significantly from region to region, the relevant coefficients have to be determined regionally. E.g., in order to derive suitable relationships according to (3.52) for Italy, Tinti et al. (1987) had to subdivide the territory into eight distinct regions for which the constant  $a$  varies between 0.27 and 0.62 and the correlation coefficients between 0.79 and 0.92. Karnik (1969) published a widely used relationship according to (3.53) for Europe:

$$M_{ms} = 0.5 I_0 + \log h + 0.35. \quad (3.55a)$$

Gutdeutsch et al. (2002), however, derived another **orthogonal** regression relationship for Central and Southern Europe, which differs considerably from Karnik's formula:

$$M_{ms} = 0.654 I_0 + 1.868 \log h - 1.682 \quad \text{with SD} = \pm 0.284. \quad (3.55b)$$

For a given magnitude,  $I_0$  depend on source depth  $h$ . Two extreme examples from Illionois earthquakes have been referred to by Nuttli et al. (1979): The same  $I_0 = VII$  has been observed both by an earthquake with  $mb = 3.8$  at  $h = 1$  km source depth and by another one with  $mb = 5.5$  at 20 km depth.

In the above equations,  $M_{ms}$  is used for a generic macroseismic magnitude. Musson and Ceci  (2002) emphasize, however, that for any particular equation it is important to specify what type of magnitude the equation is compatible with ( $M_w$ ,  $M_l$ ,  $M_s$ , or others). Moreover, it is also important to determine the standard error as a measure of uncertainty attached to the estimated magnitude values, as for the relationship (3.55b).

Relationships of the form of equation (3.54), with the area of intensities of shaking from IV to VIII scaled to local magnitude  $M_l$ , have been developed by Topozada (1975) for California and Western Nevada.

Sometimes the mean radius  $R_{ii}$  of the shaking area related to a given isoseismal intensity is used instead of the area  $A_i$  and (3.52) is then written:

$$M_{ms} = g \log R_{ii}^2 + h \log R_{ii} + j. \quad (3.56)$$

Greenhalgh et al. (1989) give such relationships for the radii of Modified Mercalli (MM) intensities III and IV as observed for Australian earthquakes and with  $M_{ms}$  scaled to  $M_l$ .

Thus, in the above relationships, the constants **a** through **j** have to be determined independently for different regions and accordingly,  $M_{ms}$  is mostly scaled to local/regional  $M_l$ . Also  $M_l$  requires local/regional scaling and has proven to be best related to earthquake damage and engineering applications because of  $A_{max}$  being measured in the most relevant frequency range for earthquake engineers between about 0.5 to 10 Hz. Yet, some  $M_{ms}$  relationships in the United States have been scaled instead to the near 1-Hz  $mb(Lg)$ , e.g. by Street and Turcotte (1977) for the Northeastern North America and by Nuttli et al. (1979).

In recent years, however, it has become increasingly fashionable to unify magnitudes in earthquake catalogues to the moment magnitude  $M_w$  (for criticism see 3.2.10), although  $M_w$  is a static measure of earthquake size and derived from the long-period asymptotic plateau of the radiated source displacement spectrum instead of being measured around its corner frequency at which the maximum of seismic energy is radiated. Yet, as long as the considered magnitudes  $M_w < 6.5$  and thus the corner frequencies according to Fig 3.5 on average  $> 0.3$  Hz such an approach may still be acceptable, not, however, when much larger or slow earthquakes play a significant role in the area under consideration.

Gr nthal et al. (2009) published such a relationship based on well constrained moment magnitude  $M_w$ ,  $I_0$ , and  $h$  of European earthquakes. It reads

$$M_w = 0.667 I_0 + 0.30 \log(h) - 0.10, \quad (3.57)$$

or, when no reliable information about  $h$  is available

$$M_w = 0.682 I_0 + 0.16. \quad (3.58)$$

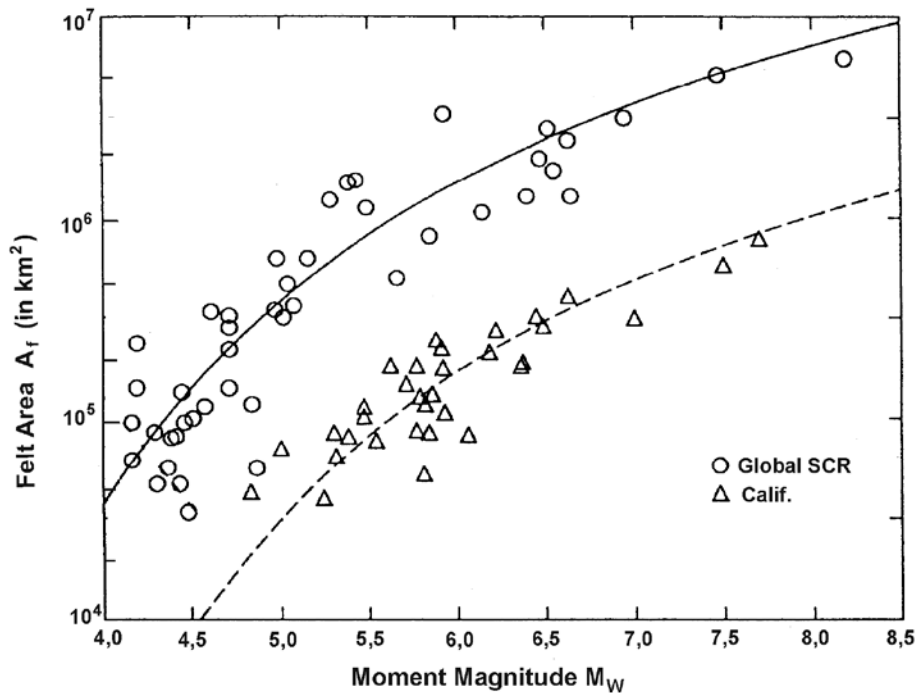
These two relations are valid in the ranges  $5 < I_0 < 9.5$ ,  $3.0 < M_w < 6.4$ , and  $5 \text{ km} \leq h < 22 \text{ km}$ .

According to Musson and Cčić (2002), differences in spectral content for larger earthquakes may affect the way in which earthquake vibration is perceived, and a different scaling with area appears then to apply. Therefore, Frankel (1994) uses for representing the full magnitude range the form

$$M_{ms} = n \log(A/\pi) + 2m/(2.3\sqrt{\pi})\sqrt{A} + a \quad (3.59)$$

where  $n$  is the exponent of geometrical spreading,  $m = (\pi f)/(Q\beta)$  with  $f$  = predominant frequency of earthquake motion at the limit of the felt area (probably 2-4 Hz),  $Q$  the shear-wave attenuation, and  $\beta$  = shear-wave velocity, taken as 3.5 km/s).

Frankel (1994) compared felt area and moment magnitudes for California with its young mountain ranges with a global data set of earthquakes in stable continental regions (SCRs) such as central USA (Fig. 3.57). The main reason for the large difference in felt area for equal moment magnitude  $M_w$  is that the average attenuation is at frequencies around 2-4 Hz, which is the range of best human perceptibility to ground shaking, much higher in California than in SCRs, with corresponding  $Q$  values of about 490 in California and 1600 in SCRs, respectively.



**Fig. 3.57** Felt area  $A_f$  (in  $\text{km}^2$ ) plotted against moment magnitude,  $M_w$ , for global data from stable continental regions (SCR) (open circles; from Johnston, 1993) and California data (triangles, from Hanks et al., 1975; Hanks and Johnston, 1992). Solid and dashed lines are fits according to an equation given by Frankel (1994) (modified from Frankel, Implications of felt area-magnitude relations for earthquake scaling and the average frequency of perceptible

ground motion, Bull. Seism. Soc. Am., Vol. 84, No. 2, Fig. 1, p. 463, 1994; © Seismological Society of America).

Johnston (1996) derived a regression relation for the Modified Mercalli Intensity scale (MMI; see Chapter 12) for stable continental regions to predict seismic moments  $M_0$  according to the functional regression form (3.59) proposed by Frankel (1994) with an uncertainty in the range of  $\pm 0.15$  to  $0.23$   $M_w$  units, and another  $M_{ms}$  scale based on  $P = I_0 \times A$  (in  $\text{km}^2$ ) had been published by Galanopoulos (1961):

$$M_{ms} = \log P + 0.2 (\log P - 6), \quad (3.60)$$

A macroseismic magnitude scaled to the body-wave magnitude of Central United States earthquakes in the range  $2.7 \leq mb \leq 5.5$  was developed by Nuttli and Zollweg (1974):

$$mb = 2.65 + 0.098 \log A_f + 0.054 (\log A_f)^2. \quad (3.61)$$

It is applicable for magnitude estimates of central United States earthquakes with felt areas of shaking  $A_f \leq 10^6 \text{ km}^2$  for which there are intensity maps but no instrumental data available.

Other forms of these equations combine  $I_0$  and the radius  $r$  of the total macroseismic field (Albarelo et al., 1995), either in the way

$$M_{ms} = a I_0 + b \ln r + c, \quad (3.62a)$$

or, when using values for each available isoseismal  $I$ ,

$$M_{ms} = a I_0 + \sum b_i \ln r_i + c. \quad (3.62b)$$

However, because of the subjectivity in isoseismal construction, modern opinion favours the direct use of the distribution of intensity  $I$  data points (IDPs) at epicentral distance  $D$ , without any isoseismals. This is particularly suitable in the case of sparse intensity data. An example are the two formulas given by Braunmiller et al. (2005) for earthquakes in Switzerland:

$$M_{ms} = [I(D) - 0.096 + 0.043D]/1.27 \quad \text{for shallow earthquakes (h} \approx 5 \text{ km)} \quad (3.63a)$$

and

$$M_{ms} = [I(D) + 1.173 + 0.030D]/1.4 \quad \text{for deeper earthquakes (h} \approx 12 \text{ km)}. \quad (3.63b)$$

Currently three systems are available for the joint computation of the main earthquake parameters epicentre/hypocenter and magnitude from IDP sets: The Bakun-Wentworth (B-W) method, originally developed for California by Bakun and Wentworth (1997), the BOXER method for Italian earthquakes with  $M_s > 5.5$  by Gasperini et al. (1999) and MEEP (Macroseismic Estimation of Earthquake Parameters) by Musson and Jiménez (2008). In Chapter 12 (Fig. 12.8) of this Manual, Musson and CeciĆ compare the results of all three methods, applied to the same IDP set of an earthquake in the UK, with the instrumentally calculated epicentre and magnitude. All these methods require a well-calibrated intensity-attenuation model valid for the region under consideration.

A related problem is the determination of magnitudes of prehistoric and historic (pre-instrumental) earthquakes from dimensions (length  $L$ , width  $W$  and/or dislocation  $D$ ) of

observed seismo-dislocations (e.g., Khromovskikh, 1989; Wells and Coppersmith, 1994; Mason, 1996) based on correlation relationships between magnitudes and respective field data from recent events (see 3.3.4).

### 3.2.6.7 Tsunami magnitude

A different kind of magnitude is the tsunami magnitude scale  $M_t$  as defined by Abe (1979 and 1981b). The scale is so constructed that  $M_t$  agrees with the moment magnitude  $M_w$ . According to Abe (1989) holds

$$M_t = \log H_{\max} + a \log \Delta + C \quad (3.64)$$

where  $H_{\max}$  is the maximum single (crest or trough) amplitude of tsunami waves in m as measured by tide-gage records and /or as derived from maximum inundation height,  $\Delta$  - epicentral distance in km to the respective tide station and  $a$  and  $C$  - constants ( $a$  was found to be almost 1). In case of the long-wave approximation, i.e., with tsunami wavelengths being much larger than the bathymetric depths, the maximum tsunami height is strictly related to the maximum *vertical* deformation of the ocean bottom,  $D_{\perp \max}$ , and thus to the seismic moment  $M_0$ .  $M_t$  was calibrated, therefore, with the average condition  $M_t = M_w$  for the calibration data set. This resulted in:

$$M_t = \log H_{\max} + \log \Delta + 5.8. \quad (3.65)$$

(3.65) shows no saturation. For the Chile earthquake 1960  $M_t = 9.4$  while  $M_w = 9.5$ . Sometimes, very slow but large ruptures with a large seismic moment cause yet much stronger tsunami than would have been expected from their moment, surface wave, energy or body-wave magnitudes  $M_w$ ,  $M_s$ ,  $M_e$  or  $m_b$ , respectively. Such events are called "tsunami earthquakes" (Kanamori, 1972; Kanamori and Kikuchi, 1993; Polet and Kanamori, 2009). A striking example is the April 1, 1946 Aleutian earthquake with  $M_s = 7.3$ ,  $M_w = 8.6$  and  $M_t = 9.3$ .  $M_t$  avoids this underestimation of possible tsunami effects of such earthquakes by the common magnitude scales. Hara (2007b) gives two more examples for the 1992 Nicaragua earthquake ( $M_w = 7.6$ ,  $M_t = 8.0$ ) and the 1996 Peru earthquake ( $M_w = 7.5$ ,  $M_t = 7.8$ ). Such earthquakes have negligibly small energy in the high-frequency range and cause little or no shaking damage (see discussions in section 3.2.7.2 on energy magnitude  $M_e$ ).

## 3.2.7 Non-saturating magnitude scales $M_w$ and $M_e$

### 3.2.7.1 Moment magnitude $M_w$ (P. Bormann)

The history of the derivation of the moment magnitude scale and the essential assumptions on which it is based have been outlined in detail by Bormann and Di Giacomo (2011) and in 3.1.2.5. Here we summarize only the essentials.

According to Eq. (3.2) and Fig. 3.5 the scalar seismic moment  $M_0 = \mu \bar{D} A$  is determined from the asymptote of the displacement amplitude spectrum as frequency  $f \rightarrow 0$  Hz. If this condition is fulfilled  $M_0$  does not saturate. Kanamori (1977) proposed, therefore, a moment magnitude,  $M_w$ , which is tied to  $M_s$  but which would not saturate either. He reasoned as follows: According to Kostrov (1974) the radiated seismic strain energy is proportional to the

stress drop  $\Delta\sigma$ , namely  $E_s \approx \Delta\sigma \bar{D} A/2$ . With Eq. (3.2) one can write  $E_s \approx (\Delta\sigma/2\mu) M_0$ . (For definition and determination of  $M_0$  and  $\Delta\sigma$  see also IS 3.1 and EX 3.4). Assuming a reasonable value for the shear modulus  $\mu$  in the crust and upper mantle (about  $3\text{--}6 \times 10^4$  MPa) and assuming that, according to Kanamori and Anderson (1975) and Abe (1975), the stress drop of large earthquakes is remarkably constant (ranging between about 2 and 6 MPa; see Fig. 3.37), one gets as an average  $E_s \approx M_0/2 \times 10^4$  (see Fig. 3.88). Inserting this into the relationship proposed by Richter (1958) between the released seismic strain energy  $E_s$  and  $M_s$ , namely

$$\log E_s = 4.8 + 1.5 M_s \text{ (in SI units Joule J = Newton meter Nm)} \quad (3.66)$$

it follows:

$$\log M_0 = 1.5 M_s + 9.1, \quad (3.67)$$

which holds, however, according to empirical data by Ekström and Dziewonski (1988), for  $M_s > 6.8$  only.

Solving (3.67) for the magnitude and replacing  $M_s$  with  $M_w$  one gets with  $M_0$  in Nm

$$M_w = (\log M_0 - 9.1)/1.5 = (2/3) (\log M_0 - 9.1) \quad (3.68)$$

and  $M_w = M_s$  for  $M_s > 6.8$ . This has been proved to be essentially correct by more recent empirical  $M_w$ - $M_s$  regression relationships with slopes of 0.94 between  $6.5 < M_s < 8.8$  (Bormann et al., 2009), 0.99 between  $6.2 \leq M_s \leq 8.2$  (Scordilis, 2006), 1.06 between  $6.2 < M_s \leq 8.4$  (Das et al., 2011) and 1.10 for  $M_s$  between 6.5 and 8.7 (Di Giacomo et al., 2013b). Formula (3.68) is now the IASPEI (2005 and 2013) accepted standard form of writing the moment magnitude formula. It avoids rounding errors and an ambiguity that arises if multiplication is performed prior to subtraction, which in a certain percentage of cases leads to  $M_w$  being different according to whether the moment is expressed in CGS or SI units (Utsu, 2002, and IS 3.3).

$M_0$  and  $M_w$  scales well with the logarithm of the rupture area [see Eqs. (3.165-3.167) and Figs. 3.94 and 3.95]. The determination of  $M_0$  on the basis of digital broadband records is becoming increasingly standard at modern observatories and network centers. Yet, the source spectra of small to moderate earthquakes with magnitudes  $< 5$  have usually corner frequencies higher than 0.5 Hz (see Fig. 3.5) and are often recorded at local to near regional distances with short-period instruments only.  $M_0$  is then usually obtained in the spectral domain by determining the spectral plateau amplitude  $u_0$  and using formula (3.2) (e.g., EX 3.4). For stronger up to great earthquakes ( $M > 8$ ) however, recorded at teleseismic distances and much longer wavelengths, least-square waveform fitting of synthetic to long-period filtered seismograms is the preferred procedure. Dziewonski et al. (1981) developed such an algorithm for the full inversion of moment tensors from long-period body- and surface-waveforms with periods up to 45 s and 135 s, respectively. The automated version, which has become known as the Harvard Centroid Moment Tensor project and since 2006 as the Global CMT (<http://www.globalcmt.org/>), also yields the scalar seismic moment  $M_0$  and  $M_w$  as spin-off parameters. This method is still used, with some upgrading, e.g. those aiming at faster and more reliable solutions for very great complex mega-earthquakes. For such events the Harvard CMT moments turned out to be systematically too low because the longest periods considered were below the corner period of the single point source spectrum and the algorithm did not provide for fitting multiple source moment release functions (Stein and



Okal, 2005; Tsai et al., 2005). Provisions have now been made to use longer periods for megathrust earthquakes. GCMT catalog data for events with  $M_w > 5.5$  are considered to be available rather complete since 1977 but in more recent times also available down to  $M_w \approx 4.5$ , although incomplete (see Figs. 3.81 and 3.82). For a most recent review about this project see Ekström et al. (2012).

Besides the GCMT also other agencies offer CMT solutions routinely for global and/or regional seismic events, such as the USGS NEIC, the GEOFON Big Alert Service of the GFZ German Research Centre for Geosciences (<http://geofon.gfz-potsdam.de/eqinfo/gevn>), the Swiss Earthquake Service (SED), the Japan Meteorological Survey (JMA), the Geophysical Survey of the Russian Academy of Sciences in Obninsk, to name just a few. In general, however,  $M_0$ , published by different authors or agencies, depends on details of the individual inversion methodologies and thus related event  $M_w$  may differ. Several alternative algorithms sharing the waveform inversion approach have been developed, e.g., by Aki and Patton (1978) and Kanamori and Given (1981) using surface waves and by Langston (1981), Sipkin (1982: 1986); and Nabelek (1984) using body waves. The USGS moment tensor solutions according to Sipkin (1986) use 15-55 s passband filtered waveform data. Therefore, this procedure also underestimates  $M_w$  for really great earthquakes. The code of Ammon (2001), developed for use on smaller magnitude earthquakes, works well with bandpass filtered waveform data between some 20 and 50 s. Thus,  $M_w$  methodologies and results have also to be critically checked for the range of their non-saturating applicability. For a detailed summary on the theory and methodologies of seismic moment tensor solutions and available services see IS 3.8, IS 3.9, and related exercises EX 3.5 and 3.6.

### 3.2.7.2 Energy magnitude $M_e$ (D. Di Giacomo and P. Bormann)

Also the energy magnitude  $M_e$  is in principle a non-saturating magnitude, provided that it can be measured by covering a sufficiently wide range of spectral velocity amplitudes. Bormann and Di Giacomo (2011) outlined in detail the common roots of  $M_e$  and  $M_w$  as well as their differences both in theory and practical application (see also 3.1.2.5). In the following we recapitulate the essentials.

According to Kanamori (1977)  $M_w$  agrees very well with  $M_s$  for many earthquakes with a rupture length of about 100 km. Furthermore, he suggested that the  $\log E_s$ - $M_s$  relationship Eq. (3.66) also gives a correct value of the radiated seismic-wave energy for earthquakes with rupture dimensions larger than about 100 km. Therefore, Kanamori considered the  $M_w$  scale to be a natural continuation of the  $M_s$  scale for larger events. Thus, when inserting into Eq. (3.66) the value of  $M_w = 9.5$  for the Chile 1960 earthquake instead of the saturated value  $M_s = 8.5$  one gets a seismic energy release that is 30 times larger.

When substituting in Eq. (3.66) the surface-wave magnitude  $M_s$  by an energy magnitude  $M_e$ , it follows

$$M_e = (\log E_s - 4.8)/1.5 = (2/3) (\log E_s - 4.8). \quad (3.69)$$

In the case that Kanamori's condition  $E_s/M_0 \approx 5 \cdot 10^{-5}$  holds, Eq. (3.69) reduces to  $M_e = (2/3) (\log M_0 - 9.1) = M_w$ . The latter has been published earlier by Purcaru and Berckhemer (1978), yet is valid only for the average apparent stresses (and related stress drop) on which the Kanamori condition is based. However, Choy and Boatwright (1995) showed that apparent stress, which is related to the ratio of  $E_s/M_0$ , may vary even for shallow events over a

wide range between about 0.03 and 20.7 MPa. For deep earthquakes, the range is between about 0.03-7.4 MPa. They also found systematic variations in apparent stress as a function of focal mechanism, tectonic environment, and seismic setting. Oceanic intraplate and ridge-ridge transform earthquakes with strike-slip mechanisms tend to have higher stress drop and  $E_s/M_0$  ratio than interplate thrust earthquakes, with fault maturity obviously playing a major role (see IS 3.5). Accordingly,  $M_e$  for the former will often be significantly larger than  $M_w$ . The opposite will be true for the majority of thrust earthquakes in subduction zones, for which  $M_w$  will be larger than  $M_e$ . Therefore, Choy (2012) proposed  $\Delta M = M_e - M_w$ , termed “differential magnitude”, as a useful diagnostic parameter to discriminate between earthquakes with unusually high, intermediate and anomalously low energy radiation (see Figure 5 in IS 3.5). This point had repeatedly been made already in earlier publications (see next paragraph).

Riznichenko (1992) gave a correlation on the basis of data from various authors. It predicts (despite rather large scatter) an average increase of  $\Delta\sigma$  with source depth  $h$  according to  $\Delta\sigma = 1.7 + 0.2 h$ , i.e., stress drops even larger than 100 MPa can be expected for very deep earthquakes. On the other hand, Kikuchi and Fukao (1988) found from analyzing 35 large earthquakes in all depth ranges that  $E_s/M_0 \approx 5 \cdot 10^{-6}$ , i.e., a ratio that is one order of magnitude less than the condition used by Kanamori for deriving  $M_w$ . Therefore,  $M_e$  is not uniquely determined by  $M_w$ . Rather,  $M_e$  and  $M_w$  are complementary descriptions of earthquake size and can be considerably different, up to 1 m.u. and even more. Striking examples have been published by Choy and Kirby (2004), Di Giacomo et al. (2010a), Bormann and Di Giacomo (2011) and by G. Choy in IS 3.5 of this Manual.

The difference between  $M_w$  and  $M_e$  reflects that  $M_w$  is nothing but a static measure of earthquake size [see relationship (3.1)] and thus of its overall tectonic effect. Thus,  $M_w$  does not contain any information about the dynamics of the source process. In contrast,  $M_e$  is mainly controlled by the latter, namely by differences in stress drop and rupture velocity, which have a direct bearing also on the dominating frequency content of the radiated seismic energy (see also section 3.1.2.5).

Nowadays, with digital broadband recordings and fast computer programs, it is feasible to determine directly the seismic energy,  $E_s$ , by integrating the radiated energy flux in velocity-squared seismograms over the duration of the source process and correcting it for the effects of geometric spreading, attenuation and radiation pattern. According to Di Giacomo and Bormann (2011), via Venkatamaran and Kanamori (2004), it holds

$$\begin{aligned}
 E_s &= \left( \frac{1}{15\pi\rho\alpha^5} + \frac{1}{10\pi\rho\beta^5} \right) \int_{-\infty}^{\infty} \left| \hat{\dot{M}}(f) \right|^2 df \\
 &= \left( \frac{2}{15\pi\rho\alpha^5} + \frac{1}{5\pi\rho\beta^5} \right) \int_0^{\infty} \left| \hat{\dot{M}}(f) \right|^2 df \\
 &\approx \left( \frac{2}{15\pi\rho\alpha^5} + \frac{1}{5\pi\rho\beta^5} \right) \int_{f_1}^{f_2} \left| \hat{\dot{M}}(f) \right|^2 df
 \end{aligned} \tag{3.70}$$

where  $\rho$  = density,  $\alpha$  and  $\beta$  =  $P$ - and  $S$ -wave velocities of the medium,  $\hat{\dot{M}}(f)$  = derivative of the seismic moment rate,  $f$  = frequency and  $f_1$  and  $f_2$  the lower and upper bounds of the integration, respectively.

A method, developed by Boatwright and Choy (1986) is now routinely applied at NEIC to compute radiated energies for shallow earthquakes of  $m_b > 5.8$  (see IS 3.6). Its application is not so trivial and not for use with single stations. Using almost 400 events, Choy and Boatwright (1995) derived the relationship for  $E_S$ - $M_s$  as

$$\log E_S = 1.5 M_s + 4.4. \quad (3.71)$$

It indicates that (3.66) slightly overestimates  $E_S$ . On the basis of such direct energy estimates these authors developed the energy magnitude relationship

$$M_e = (2/3) (\log E_S - 4.4) \quad (3.72)$$

For earthquakes satisfying Kanamori's condition  $E_S/M_0 \approx 5 \cdot 10^{-5}$

$$M_e = 2/3 \log M_0 - 5.80 = M_w + 0.27 \quad (3.73)$$

i.e., an  $M_e$  that is even larger than  $M_w$ . The Kanamori condition, however, is a narrow case.

To the contrary, according to the global average ratio  $E_S/M_0$  in Choy et al. (2006) the average difference is  $\Delta M = M_e - M_w = -0.16$  and with respect to the newest data compilation in Figure 5 of IS 3.5 even less ( $\Delta M = -0.36$ ), which corresponds to  $E_S/M_0 = 5.75 \cdot 10^{-6}$  instead of  $E_S/M_0 \approx 5 \cdot 10^{-5}$  assumed by Kanamori (1977) for deriving the  $M_w$  formula and in very good agreement with the  $E_S/M_0 \approx 5 \cdot 10^{-6}$  by Kikuchi and Fukao (1988).

$\Delta M$  may become even much more negative, down to almost -1.5, for slow “tsunami earthquakes”, corresponding to  $E_S/M_0 \approx 1 \cdot 10^{-7}$ . The latter may generate a strong (namely long-period) tsunami but only weak short-period ground motion, which may cause no shaking-damage and might not even be felt by people in the near field of the rupture. This has repeatedly been the case, e.g., on September 2<sup>nd</sup>, 1992 with the  $M_w 7.6$  Nicaragua earthquake, which had only an  $m_b = 5.3$  and an  $M_e = 6.7$ , on June 2<sup>nd</sup>, 1994 with the  $M_w 7.8$  Java tsunami earthquake with reported values for  $M_e$  between 6.5 and 6.8, and on July 17<sup>th</sup>, 2006 with the  $M_w 7.7$  Java earthquake with  $M_e = 6.8$ . Moment-wise, all three earthquakes were comparably large, yet their rupture durations  $T_R$  ranged between about 100 s and 220 s (Kanamori and Kikuchi, 1993; Hara, 2007b; Lomax and Michelini, 2009a; Bormann and Saul, 2008 and 2009b). This is, according to the relationship  $T_R \approx 0.6M - 2.8$  by Bormann et al. (2009), about 2-4 times longer than expected on average for earthquakes with  $M_w$  around 7.7.

This strongly contrasts with earthquakes  $M_e \gg M_w$ , e.g., for the 12 January 2010  $M_w 7.1$  Haiti earthquake with  $M_e$  values of 7.5 and 7.6, respectively, which caused major devastations and a large number of casualties due to strong ground shaking in conjunction with inferior building stocks. Other examples are given by Di Giacomo et al. (2010a and b). Thus, the difference  $\Delta M = M_e - M_w$  is an important diagnostic aid to quickly assess differences in the kind of seismic hazard and damages to be expected from strong earthquakes.

Another argument in favor of  $M_e$  is that it follows more closely the original intent of the Gutenberg-Richter magnitude formula with  $(A/T)_{\max}$  as the input measurement parameter which relates magnitude to the velocity power spectrum and, thus, to energy. In contrast,  $M_w$  is related to the scalar seismic moment  $M_0$  that is derived from the low-frequency asymptote

of the displacement spectrum. Consequently,  $M_e$  is more closely related to the seismic potential for damage while  $M_w$  is related to the final static displacement and the rupture area and thus related more to the overall tectonic effect of an earthquake.

However, one major advantage of non-saturating  $M_0$  and thus  $M_w$  is that they have to be determined only from very long-period seismic waves with  $T \gg T_c$  (= corner period of the radiated seismic spectrum). This is more easily done and less affected by small-scale complexities in the source process and inhomogeneities along the wave propagation path. Therefore, measurement errors in  $M_w$  are rather low, probably even less than 0.1 m.u. when based on good data.

In contrast, to assure a non-saturating  $M_e$  the squared velocity spectrum would need to be measured widely enough towards both sides of  $f_c$  (respectively  $T_c$ ) in Fig. 3.5 in order to assure that at least some 80% of the total  $E_s$  is obtained from the integration over the source spectrum.  $M_e$  would then be underestimated by less than 0.06 m.u., which is negligible. When assuming that the low frequency part of the source spectrum is completely available, then - according to Singh and Ordaz (1994) - the highest frequency  $f_{\max}$  still to be included should be 6 times  $f_c$  for an  $\omega^{-2}$  model as in Fig. 3.5. However, the frequency range that can in practice be covered within the limits  $f_{\min}$  and  $f_{\max}$  may cut off significant amounts of seismic energy contained both in the low- and the high-frequency part of the source spectrum, especially, if the bandwidth of integration does not cover well frequencies around  $f_c$ . In the latter case  $M_e$  may be underestimated up to a few tenths of magnitude units.

Current routine procedures operate in the period ranges between 0.2 and 100 s (procedure according to Choy and Boatwright, 1995). It is outlined in IS 3.6 and applied at the US Geological Survey (USGS), or 1 and 80 s (automatic near real-time procedure according to Di Giacomo et al., 2010a and b), applied by the GFZ German Research Centre for Geosciences (GFZ). The inclusion of frequencies above 1 Hz is generally difficult because of the then usually low SNR in teleseismic recordings and the difficulties to correct for frequency-dependent attenuation (see Chapter 2, section 2.5.4.2) and local site effects. But then, the condition  $f_{\max} = 6 \times f_c$  for the inclusion of higher frequencies is already difficult to meet for  $M_w < 5.5$  in the case of an average stress drop  $\omega^{-2}$  source model as assumed in Fig. 3.5. And since the average corner periods for magnitude 8.5 earthquakes is expected to be already around 100 s the current procedures will also not be able to sample the spectral amplitudes towards  $T > T_c$ . According to Bormann and Di Giacomo (2011) Eq. (3.72) can be written as

$$M_e = M_w + [\log(\Delta\sigma/2\mu) + 4.7]/1.5. \quad (3.74)$$

This allows to roughly assess the possible influence of variations in stress drop on estimates of  $M_e$ , using the scaling model in Fig. 3.5 but for varying in increments of one order for  $\Delta\sigma$  between 0.1 and 100 MPa (see Tab. 3.3). This range of variations in  $\Delta\sigma$  encompasses most of the published data (e.g., Abercrombie, 1995; Kanamori and Brodsky, 2004; Parolai et al., 2007; Venkataraman and Kanamori, 2004a). The related shift of  $f_c$  for these different  $\Delta\sigma$  values has been plotted in Fig. 3.16 for an earthquake with  $M_w = 6.5$ , and in the inset for a wider magnitude range between  $M_w = 5.5$  and 8.5. According to Fig. 3.16a, variations in  $\Delta\sigma$  do not influence at all the displacement plateau of the source spectrum and thus estimates of  $M_0$  as long as the basic condition of analysing only frequencies  $f \ll f_c$  is observed. In contrast, according to Eq. (3.74) and in agreement with Fig. 3.16b, variations in stress drop by one order are - for a given seismic moment - expected to change the released energy by one

order as well and thus the estimates of  $M_e$  by almost 0.7 m.u.. This complicates matters even more.

In order to sample nearly 100% of the energy radiated by earthquakes with magnitudes between  $5.5 \leq M_w \leq 8.5$  and stress drops between  $0.1 \leq \Delta\sigma \leq 100$  MPa one would need to consider a much wider range of frequencies between 0.001 Hz and 16 Hz instead. This is, however, not realistic, neither with respect to handling the very high-frequency part of the radiated energy nor for fast routine  $M_e$  estimations using teleseismic P-wave records, taking also into account that the longest periods that can be used have to be shorter than the time difference between the onset of the P- and S-wave groups which can not be handled together. For a rough orientation Tab. 3.3 summarizes the magnitude-dependent underestimation of  $M_e$ ,  $-\Delta M_e$ , based on the source model of Fig. 3.5 but assuming variable stress drop between 0.1 and 100 MPa and integration over frequencies between 12.4 mHz and 1 Hz only, as in the GFZ procedure for quick  $M_e$  estimation (see D. Di Giacomo et al., 2010 a and b).

**Tab. 3.3** Estimates of  $-\Delta M_e$  for the GFZ procedure of  $M_e$  determination according to Bormann and Di Giacomo, 2011).

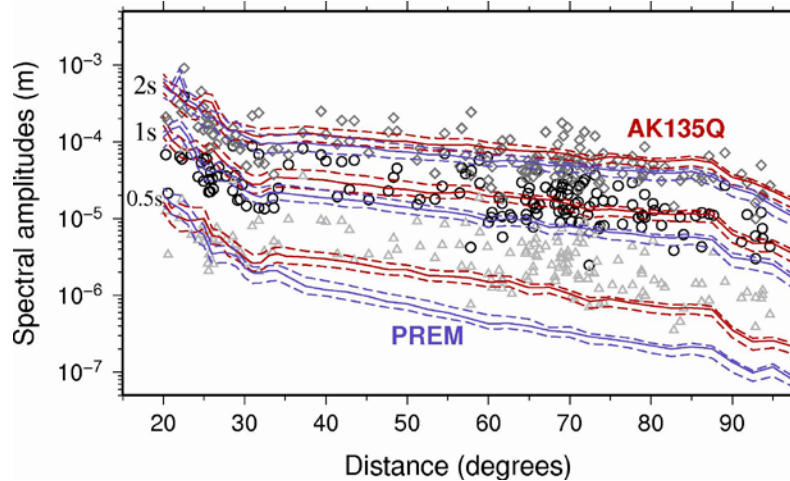
| $\Delta\sigma$ (MPa) | $\Delta M_e$ for $M_w = 5.5$ | $\Delta M_e$ for $M_w = 6.5$ | $\Delta M_e$ for $M_w = 7.5$ | $\Delta M_e$ for $M_w = 8.5$ |
|----------------------|------------------------------|------------------------------|------------------------------|------------------------------|
| 0.1                  | 0.05                         | 0.02                         | 0.05                         | 0.25                         |
| 1                    | 0.13                         | 0.03                         | 0.02                         | 0.10                         |
| 10                   | 0.30                         | 0.08                         | 0.02                         | 0.03                         |
| 100                  | 0.66                         | 0.19                         | 0.05                         | 0.02                         |

According to Tab. 3.3, an  $E_s$  procedure operating in the bandwidth range from 1 to 80 s may underestimate  $M_e$  up to  $\approx 0.66$  m.u. for moderate ( $M_w \approx 5.5$ ) earthquakes with very high stress drop and thus  $f_c > f_{\max}$ . But for most earthquakes with  $M_w$  between about 6.5 and 8.5 and intermediate  $\Delta\sigma$  between about 1 to 10 MPa the underestimation is expected to be  $< 0.1$  m.u. Yet, for great earthquakes ( $M_w > 8$ ) with  $\Delta\sigma < 1$  MPa (possibly  $f_c < f_{\min}$ ),  $-\Delta M_e$  may also reach 0.2-0.3 m.u., or even more for extreme events. However, such estimates with respect to  $M_e$  that are based on so simple a single point source rupture model must be used with caution. No models can account for all real source, propagation path and station site complexities, which luckily do not much affect teleseismic  $M_w$  estimates. Therefore, Tab. 3.3 can only for give a rough orientation of the possible range of  $M_e$  uncertainties.

Underestimations for extreme events should be somewhat reduced by the more elaborate off-line USGS procedure that uses a slightly larger bandwidth and a residual integral for frequencies above  $f_c$  (see Boatwright and Choy, 1986). In summary, according to McGarr and Fletcher (2002),  $E_s$  released by earthquakes can probably be estimated only within a factor of 2-3 if multi-station high-quality seismic broadband data are available. This corresponds to an accuracy within 0.2 to 0.3 m.u. in  $M_e$ . In the same range source directivity may affect related single station  $M_e$  estimates (Venkataraman and Kanamori, 2004b)

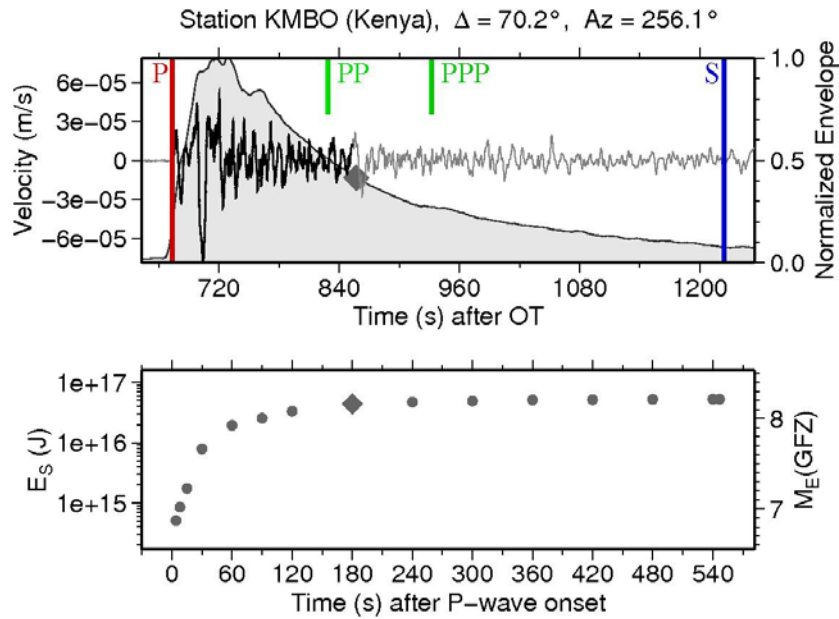
Correcting for high-frequency attenuation is one of the most challenging tasks in calculating  $M_e$ . It requires a detailed knowledge of the velocity and attenuation structure along the propagation paths, and also especially below the seismic stations. Available global attenuation models differ still significantly and are controversially discussed and empirical data show significant scatter. Fig. 3.58 compares empirical data of short-period spectral velocity

amplitudes, measured on records of many world-wide distributed broadband stations. The respective spectral amplitude-distance curves have been calculated for two different Earth models, assuming frequency-independent quality factors  $Q$ : a) the AK135Q according to Montagner and Kennett (1996) and b) the Preliminary Earth Reference Model (PREM) according to Dziewonski and Anderson (1981). The inconsistencies for shorter periods are obvious. The usually poor signal-to-noise-ratio at frequencies higher than 1 Hz in the teleseismic range represents another big limitation, especially for moderate and small earthquakes.



**Fig. 3.58** Velocity spectral amplitudes at 0.5 s (2 Hz, light gray triangles), 1 s (1 Hz, black circles) and 2 s (0.5 Hz, gray diamonds) for the  $M_w = 7.9$  Wenchuan, China, earthquake of 2008, May 12, superimposed to the theoretical spectral amplitude decay functions for the same frequencies for the AK135 model (median, 25<sup>th</sup> and 75<sup>th</sup> percentile in solid and dashed red lines, respectively) and the PREM model (median, 25<sup>th</sup> and 75<sup>th</sup> percentile in solid and dashed violet lines, respectively). An arbitrary offset has been added to the theoretical curves to make easier the comparison with the real data. (Courtesy of D. Di Giacomo, 2011).

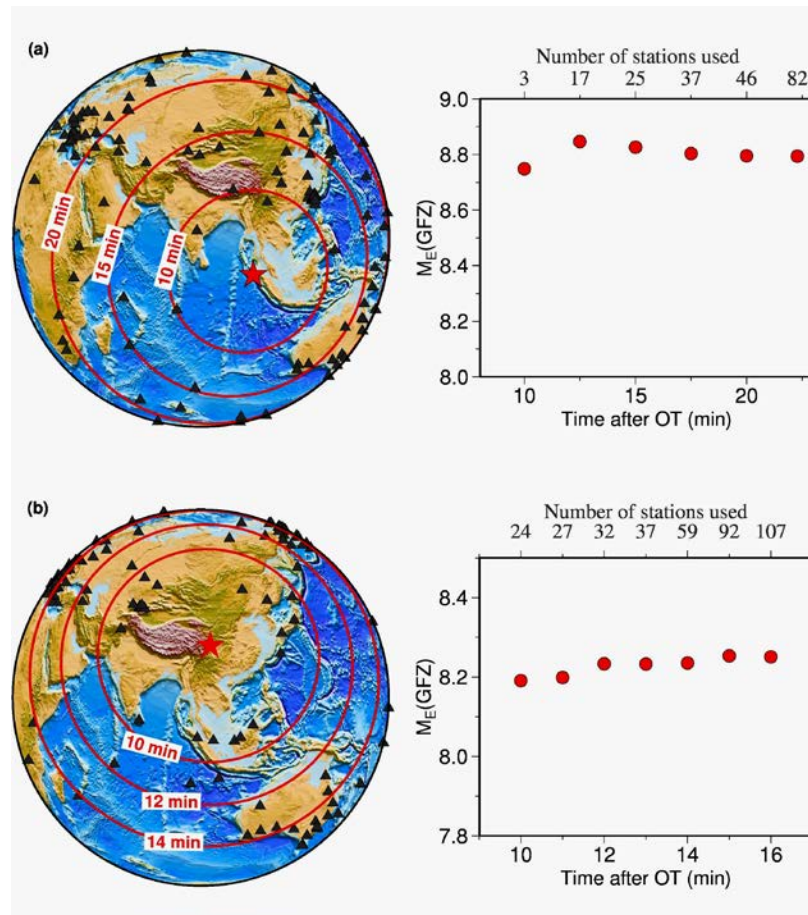
These combined difficulties and thus - as compared to  $M_0$  and  $M_w$  determinations - related larger possible errors are reasons that  $E_s$  and  $M_e$ , despite their importance in assessing the damage potential of earthquakes, have rarely been used for this purpose so far. This, however, also means that one should not over-interpret differences between  $M_w$  and  $M_e$  less than 0.2 m.u., yet larger ones may already be alarming because they may hint to different kinds of disaster consequences. Since  $M_e$  for the Wenchuan earthquake was almost 0.4 m.u. larger than  $M_w$  (see Fig. 3.59) this may at least partially explain the extra ordinarily large shaking damages. In contrast, the great tsunamigenic Sumatra-Andaman  $M_w = 9.3$  earthquake of 2004 was relatively slow with rupture velocities dropping down from 2.8 km/s at the beginning to 2.1 km/s in the Nicobar rupture segment. Accordingly, the related energy radiation efficiencies (see 3.1.2.5) were only 0.2 to 0.1 as compared to values between some 0.4 to 0.6 for other great earthquakes (Kanamori, 2006). Accordingly, energy magnitude  $M_e(\text{GFZ}) = 8.8$  is 0.5 m.u. units less than  $M_w$  for this earthquake (see Fig. 3.60a) which thus ranks between “normal” great earthquakes and the extremely slow tsunami earthquakes.



**Fig. 3.59** Upper panel: vertical component velocity recording of the  $M_w = 7.9$  12 May 2008 Wenchuan earthquake recorded at the station KMBO (Kenya). The theoretical P-, PP-, PPP- and S-wave arrival times have also been marked. The gray shaded area represents the envelope of high-frequency velocity amplitudes used to constrain the overall rupture time duration (Bormann and Saul, 2008), and the diamond at the end of the black record trace marks the end of the time window after the P-wave onset for which the final  $M_E$ (GFZ) single station value is calculated. Thus, the full rupture duration is included in the  $E_S$  calculation. However, the rupture duration may last for several minutes (as for the great 26 December 2004 Sumatra earthquake). Then the entire S-P window should be considered. Lower panel:  $E_S$  (left y-axis) and  $M_E$  (right y-axis) values for different cumulative P-wave windows. The diamond marks the end of the P-wave window that has been used for the single station  $M_E$  estimate. (Courtesy of D. Di Giacomo, 2011).

To evaluate the amount of time needed by this procedure to provide a stable  $M_E$  in a real- or near-real time implementation, Fig. 3.60 shows the  $M_E$  determinations at different times after OT for both the great 26 December 2004 Sumatra earthquake (Fig. 3.60a) and the 12 May 2008 Wenchuan earthquake (Fig. 3.60b). In the exceptional case of the 2004 Sumatra earthquake, for which the rupture duration as estimated from the duration of the P-wave train was about 500 s (Bormann and Wylegalla, 2005; Ni et al., 2005), this procedure could have yielded a stable  $M_E$  already some 15 min after OT (Fig. 3.60a, right panel), since the major energy release occurred within the first 250 s of the rupture process (Choy and Boatwright, 2007). For the case of the Wenchuan earthquake, instead, using P-wave time windows of 180 s, more stations (24) could have been used already 10 min after OT, and a stable  $M_E$  could have been released. Noticeably, in both cases the preliminary (alarm)  $M_E$ (GFZ) available after 10 min are very close to the final values obtained by using all available stations. Of course, the time performance of this approach depends on the station availability with respect to the earthquake location. However, the worldwide station deployment is becoming increasingly dense, especially in areas for which a lack of instrumentation was still common a few years ago. Therefore, this procedure could yield in the near future rapid  $M_E$  estimates within 10 min after OT also for great ( $M_w \geq 8$ ) earthquakes.





**Fig. 3.60** **a) Left:** Map showing the location of the 26 December 2004 Mw = 9.3 Sumatra earthquake (red star), the broadband stations used to calculate  $M_E(\text{GFZ})$  (black triangles), and the red circles represent the S-wave arrival after OT. **Right:**  $M_E(\text{GFZ})$  determination of the Sumatra earthquake at different time after OT. The number of stations used to compute  $M_E(\text{GFZ})$  are also given; **b) As for a)** but for the 12 May 2008 Mw = 7.9 Wenchuan earthquake. Here the red circles on the map mark the time needed to record 180 s of P-waves.

### 3.2.8 Complementary non-saturating magnitude scales (P. Bormann)

#### 3.2.8.1 Cumulative body-wave magnitude mBc

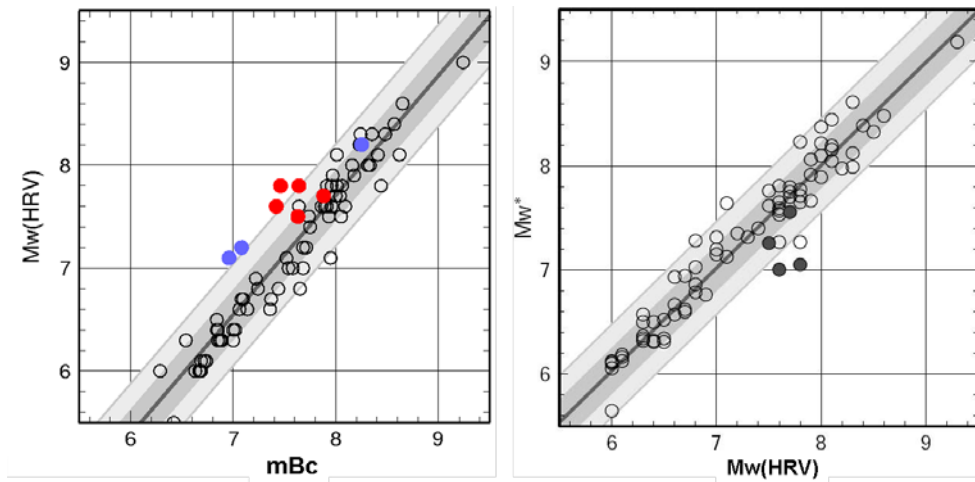
In their fundamental paper about energy and magnitude of earthquakes Gutenberg and Richter (1956a) shortly explain that their mB-log $E_S$  relationship is based on the assumption of a point source which radiates a simple-shaped waveform whose largest amplitude and duration are related to each other. This makes it possible to estimate the seismic energy travelling with this waveform by scaling it to  $A_{\max}$ , respectively  $(A/T)_{\max}$ . They also mention that because of lacking data for stronger and likely more complex earthquakes the mB-log $E_S$  relationship, which in fact is based on only 20 data points in the magnitude range between  $5.3 \leq m \leq 8$  and one data point for  $m = 2$  (see Fig. 3.83), may be less reliable. Indeed, larger earthquakes tend to be multiple source ruptures (see Figs. 3.9-3.13). Single-amplitude mB is then representative only for the largest sub-rupture. This let Bormann and Khalturin (1975) propose:



*“In such cases we should determine the onset times and magnitudes of **all clear successive P-wave onsets separately**, as they give a first rough impression of the temporal and energetic development of the complex rupture process.” And further: „The magnitude  $MP = \log \Sigma n (A_i / T_i) + Q(D, h)$  ( $n$  is the number of successive P-wave onsets) could be considered as a more realistic measure of the P-wave energy released by such a multiple seismic event [...].“*

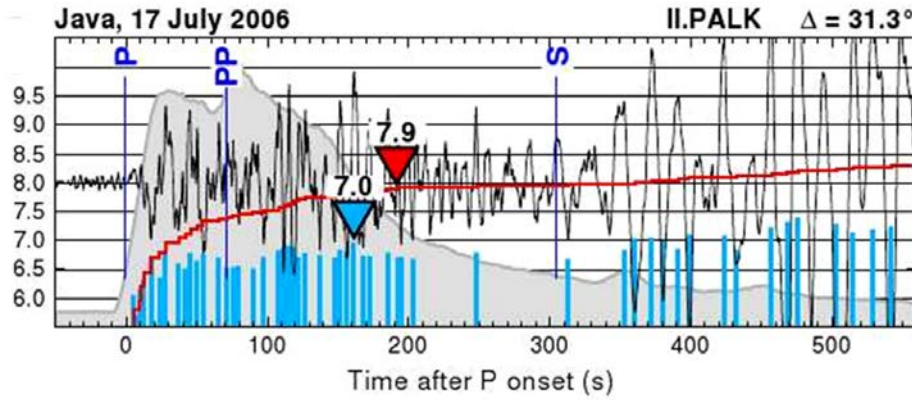
Bormann and Wylegalla (2005) demonstrated this by way of example for the great 2004 Mw9.3 Sumatra-Andaman earthquake. Just a single broadband station record near Berlin yielded a cumulative  $\Sigma mB$ , now termed mBc, of 9.4 and a rupture duration estimate of about 500 s. Meanwhile, this procedure has been fully automated (see Figs. 3.13 and 3.62, and for details Bormann and Saul, 2009b). Fig. 3.61 left shows the data and orthogonal regression relation (3.75) of Harvard CMT Mw(HRV) vs. mBc. Using Eq. (3.75) mBc can be converted into a proxy estimate of moment magnitude,  $Mw(mBc) = Mw^*$ . Fig. 3.61 right compares  $Mw^*$  with proper Mw(HRV). With  $SD_{OR}$  = orthogonal standard deviation and  $SD_Y$  in y direction holds:

$$Mw(HRV) = 1.22 \text{ mBc} - 2.11 \text{ with } SD_{OR} = 0.145 \text{ and } SD_Y = 0.23 \quad (3.75)$$



**Fig. 3.61 Left:** Orthogonal regression Mw(HRV) over mBc with 1-SD and 2-SD limits in dark and light grey; **red dots - tsunami earthquakes, blue dots – deep earthquakes**; **Right:** proxy  $Mw^* = Mw(mBc)$  calculated via Eq. (3.75); **black dots – tsunami earthquakes**. Data according to Bormann and Saul (2009); Figure kindly provided by J. Saul, 2009.

Thus, equation (3.75) allows to compute a purely empirical  $Mw^*$  estimator  $Mw(mBc)$ , which is sufficiently accurate and quick to be used as a preliminary proxy Mw in tsunami early warning applications until Mw(HRV) (now Mw(GCMT)) becomes available. Most events follow the two regressions well. For 80 percent of the events,  $|Mw^* - Mw| < 0.25$  m.u. For only two events  $Mw^*$  is significantly too low. They correspond to the very slow Mw 7.6 1992 Nicaragua and the Mw 7.8 1994 Java tsunami earthquakes. Yet, for two more tsunami earthquakes in our data set ( $M_w$  7.5 Peru 1996;  $M_w$  7.7 Java 2006, see Fig. 3.62)  $Mw^*$  agrees within less than 0.25 m.u. with Mw, also for other major tsunamigenic events like the great 2004 and 2005 Sumatra earthquakes.

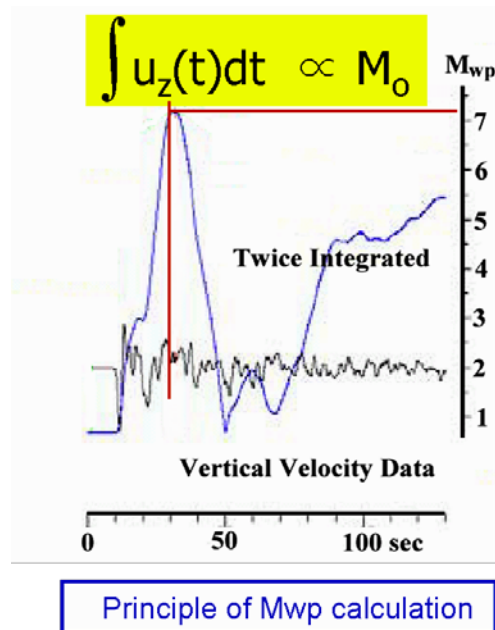


**Fig. 3.62** Velocity-proportional broadband record of the Mw 7.7 July 2006 Java tsunami earthquake at station II.PALK processed with the automated mBc procedure by Bormann and Saul (2009b). Blue bars correspond to mB values of major amplitude onsets according to the condition  $V_i \geq 0.6V_{\max,t}$  with  $V_{\max,t}$  being the maximum velocity amplitude up to the time  $t$ . The largest sub-event single-amplitude mB = 7.0 has been marked by a blue triangle. The red curve shows the development of mBc with time and the red triangle the asymptotically saturated mBc value reached prior to the onset of S. The grey shaded area marks the high-frequency (HF) P-wave envelop. The apparent rupture duration  $d_{app}$  is measured from the P-onset up to the red triangle where the HF envelope amplitudes have dropped to 40% of their maximum value, corresponding to 16% energy wise.

From Fig. 3.61 left one realizes that mBc tends to overestimate increasingly magnitudes for  $M_w < 8$ . This means that for smaller earthquakes Gutenberg and Richter's (1956a) original assumptions of a simple scaling of the maximum far-field P-wave amplitude to the duration of the waveform and its related seismic energy holds on average sufficiently well. Accordingly, summing up multiple P-wave onsets for smaller earthquakes tends to yield  $mBc > M_w$  and  $> mB$ . The apparent rupture duration estimates published by Bormann and Saul (2008 and 2009) agree for 90% of the investigate earthquakes within 10% with those published by Lomax and Michelini (2009a) for the same events, yet with a tendency of mB-mBc durations to be deliberately somewhat longer. This was necessary for assuring that the depth phases of P are included within this time window in order to satisfy IASPEI rules for mB measurements (see explanations to formulas (3.44 and 3.45). For the Java tsunami earthquake in Fig. 3.62  $d_{app} = 190$  s, which is almost three times longer than the rupture duration of 66 s estimated with formula (3.3) for an "average"  $M_w = 7.7$  earthquake.

### 3.2.8.2 Mwp

A simple, fast and robust method of  $M_w$  determination from broadband P waveforms, termed Mwp, has originally been developed by Tsuboi et al. (1995) for quick estimation of the moment magnitude  $M_w$  and the tsunami potential of large near coastal earthquakes offshore of Japan. It has later been extended by Tsuboi et al. (1999) for application also to teleseismic earthquakes in general. Since then Mwp has become the main procedure at the Pacific Tsunami Warning Center (PTWC) for rapid rough  $M_w$  estimation. The procedure is based on the double integration of broadband P-wave velocity records and scaling of the maximum of the resulting source-time function via  $M_0$  to  $M_w$  (Fig. 3.63).



**Fig. 3.63** **Black trace:** Original velocity-proportional vertical component P-wave record; **Blue trace:** Twice integrated record trace. According to Seidl and Berckhemer (1982) the **area** underneath the blue trace is proportional to  $M_0$ . For simplicity Tsuboi et al. (1995) scaled the first maximum of the blue trace to  $M_{wp}$ .

However, the original scaling as well as later introduced corrections were not so straightforward. Scaling to the first peak in the double integrated trace which is in fact the far-field source-time function, and not to the area underneath the total source-time function (see also Fig. 3.65a), assumes a single point source rupture model with a more or less symmetric moment-rate function and a simple relationship between the maximum amplitude and the area underneath the curve. Later, Tsuboi et al. (1999) changed this by scaling  $M_{wp}$  to the largest of the first two maxima, accepting that this could relate either to a stronger depth phase (pP or sP) or to a stronger sub-event of the progressing rupture. Yet still, in the case of very large, more complex or very slow ruptures,  $M_{wp}$  tends to underestimate the true  $M_w$ , because even later sub-ruptures than the second one may have the largest amplitude, and most importantly, the rupture duration is not taken into account. For equal maximum amplitude of the source time function its duration and thus the area underneath the function may vary strongly, being much larger, e.g., in the case of slow earthquakes. Even with magnitude-dependent corrections introduced the  $M_{wp} = 8.2$  for the great 2004 Sumatra earthquake turned out to be much too small (see Hirshorn and Weinstein, 2009). This value was later increased to  $M_{wp} = 8.5$  by assuming a variable apparent P-wave velocity (Kanjo et al., 2006).

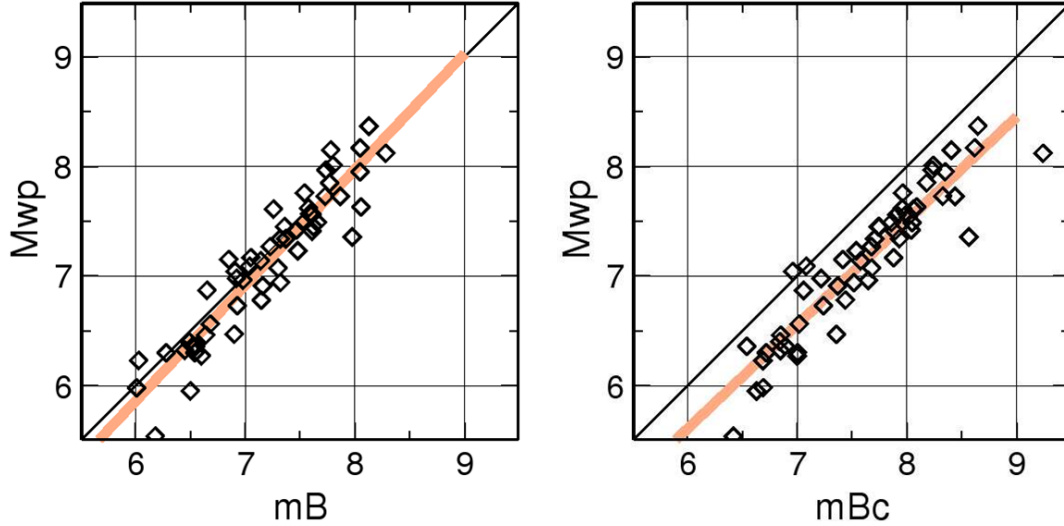
Accordingly,  $M_{wp}$  turned out to be neither more efficient nor faster than the simple automated direct measurement of the traditional Gutenberg  $m_B$  on velocity broadband records, which is now a new IASPEI body-wave standard. The reason is that also  $m_B$  is based on the same simplified assumption, namely that the maximum amplitude of a single point source waveform in the far-field is scaled to the area underneath this waveform.

When plotting  $M_{wp}$  over  $m_B$  (see section 3.2.8.1)) one gets almost a 1:1 orthogonal regression relationships whereas  $m_{Bc}$  is always larger but not saturating as do both the single amplitude  $m_B$  and  $M_{wp}$  (Fig. 3.64). The regression relations are:

$$M_{wp} = 1.06 mB - 0.09$$

and

$$M_{wp} = 0.95 mBc - 0.47.$$



**Fig. 3.64** Data and orthogonal regression relationships between  $M_{wp}$  of the Pacific Tsunami Warning Center and  $mB$  (left), respectively  $mBc$  (right), according to data of Bormann and Saul (2009). Figure courtesy of J. Saul, 2009.

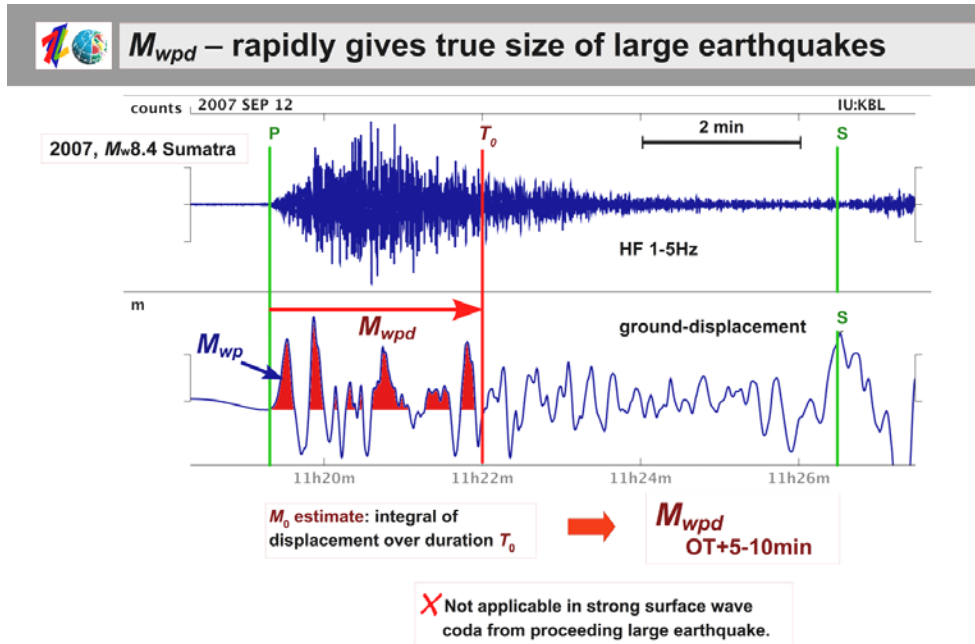
### 3.2.8.3 $M_{ED}$ and $M_{wpd}$

The main drawback of both  $M_{wp}$  and  $mB$  (in contrast to  $mBc$ ) is that rupture duration and related major sub-events in seismic energy and moment release are not taken into account. The lack of these contributions, which may become very significant for great multi-source or extremely long-lasting ruptures (see Fig. 3.65a below), results in the underestimation of the overall event magnitude and thus of the tsunamigenic potential of really great earthquakes and of slow tsunami earthquakes in particular (e.g., Fig. 3.74 in section 3.2.9.4 and Hirshhorn and Weinstein, 2009).

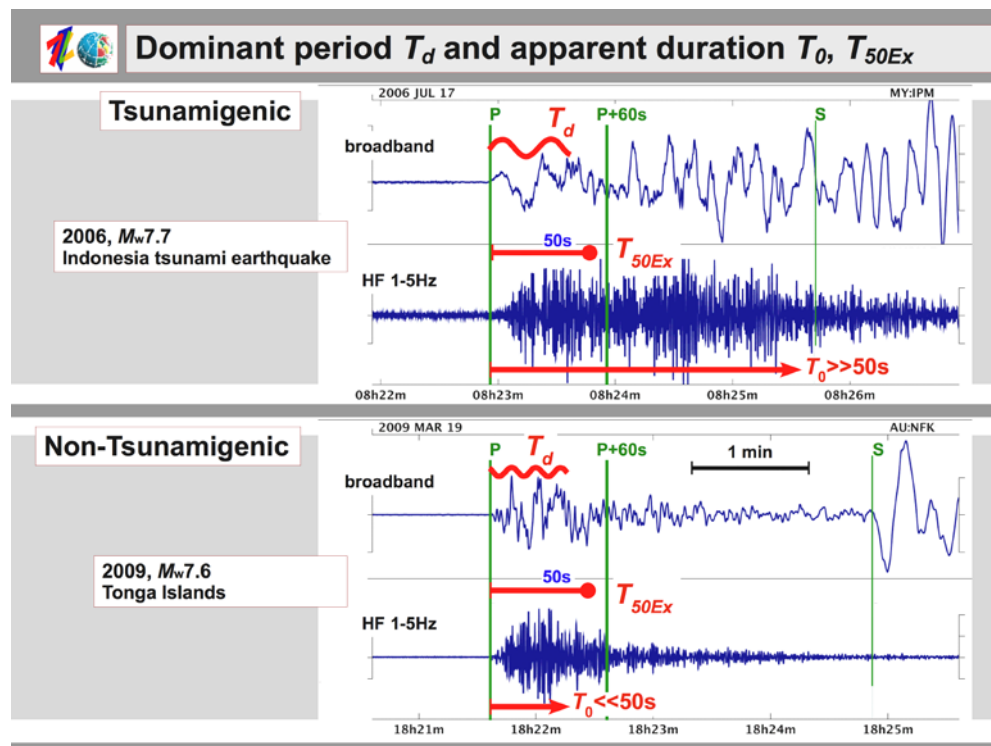
To address this problem, and for rapid and accurate determination of the size and tsunamigenic potential of strong and great earthquakes, Lomax et al. (2007) and Lomax and Michelini (2009a) developed non-saturating, energy,  $M_{ED}$ , and moment,  $M_{wpd}$ , magnitudes, which explicitly take into account rupture duration. Similar to the procedures by Bormann and Saul (2008 and 2009) for measuring  $mB$  and  $mBc$  on velocity broadband records, the  $M_{ED}$  and  $M_{wpd}$  procedures estimate the apparent rupture duration,  $T_0$ , on high-frequency (HF; 1-5 Hz) filtered vertical component broadband seismograms.  $M_{ED}$  is based on an energy-duration procedure which cumulated energy in the time interval  $t_{T_0} - t_p$  on vertical component, broadband velocity seismograms. The developed relationship between the estimated released seismic energy and the measured apparent rupture duration allowed a rapid estimation of  $\Theta = \log(E_S/M_0)$  for identifying slow but dangerous tsunami earthquakes, namely, if  $\Theta$  drops below -5.7.

Mw<sub>pd</sub> is based on a duration-amplitude procedure which is equivalent to cumulative application of the Mw<sub>p</sub> algorithm (i.e., moment determination) in the time interval  $t_{T0} - t_p$  on vertical component, broadband displacement seismograms. In operational application Mw<sub>pd</sub> values can be made available within about 10 min after origin time (OT) (Fig. 3.65a). First results showed, however, that additionally a scaling of the moment estimates was necessary in order to assure a better match with Mw(CMT). With such a scaling the difference  $|Mw_{pd} - Mw(CMT)|$  for events with magnitude  $6.6 \leq Mw \leq 9.3$  was typically within  $\pm 0.2$  m.u., with a standard deviation  $SD = 0.11$  m.u. This moment correction has recently been simplified, along with other modifications, to enable faster, more robust real-time estimates of Mw<sub>pd</sub> (Lomax and Michelini, 2012).

Lomax and Michelini (2009b, 2011) show that neither  $\Theta$  nor Mw alone proved to be good indicators for tsunamigenic earthquakes in general, while apparent rupture durations longer than about 50 s combined with predominant P-wave period  $T_p$  greater than about 10 s are good indicators. Lomax and Michelini (2009b) also show that measures on 1-5Hz HF filtered velocity broadband records alone can show for most earthquakes within 4-6 min after OT if the rupture duration is likely to exceed 50 s, indicative for tsunamigenic and tsunami earthquakes. Although less accurate than  $T_0$ , this rapidly available measure,  $T_{50EX}$ , is an important complement in operational tsunami early warning schemes. By combining  $T_{50EX}$  with a measurement of the predominant period,  $T_d$ , in the early part of the P-wave record, and later by estimating the apparent total rupture duration  $T_0$ , the products  $T_d T_{50EX}$  and  $T_d T_0$  can be calculated (see Fig. 3.65b). These products proved to be reliable indicators for the tsunamigenic potential of earthquakes in general and for slow tsunami earthquakes in particular, superior to Mw(GCMT) (Lomax and Michelini, 2011, 2012). The reason is that the efficiency of tsunami generation by a shallow earthquake depends on the amount of sea floor displacement, which can be related to a finite-faulting model expressed by the *seismic potency*,  $LWD$ , where  $L$  is the length,  $W$  the width, and  $D$  the mean slip of the earthquake rupture (Kanamori 1972; Abe 1973; Polet and Kanamori, 2009). Mw is calculated from the seismic moment  $M_0 = \mu LWD$  by assuming that  $\mu$ , the shear modulus at the source, and thus also  $LWD = M_0/\mu$  are constant and independent on source depth. This, however, is not the case, especially in shallow subduction zones.  $\mu$  may there strongly decrease with source depth and fault rupture may occur close to the sea bottom on non-planar, lystric splay faults in water saturated sediments in the accretionary wedge (Fukao, 1979; Moore et al., 2007; Lay and Bilek, 2007). This will result in strongly reduced rupture velocity and stress drop and thus longer rupture duration with longer dominating periods of the radiated P-waves. In general, CMT algorithms, do not account for this low velocity and shear modulus. Accordingly, the seismic potency and tsunami potential may be underestimated, especially in the case of unusually slow “tsunami earthquakes”, which, by definition, cause larger tsunami waves than one would expect from their Mw (Kanamori, 1972; Satake, 2002; Polet and Kanamori, 2009; Newman et al., 2011). Both  $T_d$  and  $T_0$ , as well as the estimator of critical duration,  $T_{50EX}$ , “sense” these differences in source-depth dependent  $\mu$ , and related reduced rupture velocity, increased rupture duration and potency  $LWD$ , and consequently increased sea surface displacement and tsunami potential.



a)



b)

**Fig. 3.65** Illustration of a) the duration estimation and displacement integration for  $M_{wpd}$  determination and b) the estimation of the predominant period  $T_d$ , the duration exceedence probability parameter  $T_{50Ex}$  and of the total apparent rupture duration  $T_0$  and their use for assessing the tsunamigenic potential of earthquakes. For details see text and Lomax and Michélini (2012). (Figure provided by courtesy of A. Lomax, 2013; with kind permission from Springer Science and Business Media).

The Mw<sub>pd</sub> and TdT<sub>50EX</sub>-Td T<sub>0</sub> procedures have been implemented operationally at the Istituto Nazionale di Geofisica e Vulcanologia (INGV) in Rome, Italy.

#### 3.2.8.4 Hara's maximum amplitude-duration magnitude M

In a slightly different and very straight forward empirical procedure, Hara (2007a and b) combines measuring the high-frequency duration of the P-wave train with just the maximum P-wave displacement amplitude in broadband records to determine an empirical relation for a non-saturating magnitude M scaled to Mw in the Harvard CMT catalog:

$$M = 0.79 \log A + 0.83 \log \Delta + 0.69 \log t + 6.47$$

where A is the maximum displacement (in m) during the estimated duration t (in s) of high-frequency energy radiation since the P-wave first arrival time and  $\Delta$  the epicentral distance (in km).

The Hara M matches with Mw(CMT) typically within  $\pm 0.3$  m.u., with SD = 0.18. It is also applicable to tsunami earthquakes (2007b) that are characterized by relatively longer rupture duration but smaller amplitudes. However, in this case M underestimates, as Mw, the tsunami magnitude  $M_t$  (see 3.2.6.7).

#### 3.2.8.5 Mantle magnitude Mm

Another important teleseismic magnitude is called mantle magnitude Mm. It uses surface waves with periods between about 60 s and 410 s that penetrate into the Earth's mantle (see section 2.3.4 in Chapter 2). The concept has been introduced by Brune and Engen (1969) and further developed by Okal and Talandier (1989 and 1990). Mm is firmly related to the seismic moment  $M_0$  and thus avoids saturation. On the other hand, it is closer to the original philosophy of a magnitude scale by allowing quick, even one-station automated measurements (Hyvernaud et al., 1993), that do not require the knowledge of either the earthquake's focal geometry or its exact depth. The latter parameters would be crucial for refining a moment estimate and require (global) network recordings.

Mm is an estimate of  $(\log M_0 - 13)$  (when  $M_0$  is given in Nm) and defined as:

$$Mm = \log X(\omega) + C_S + C_D - 0.90$$

where  $X(\omega)$  is the spectral amplitude of a Rayleigh wave in  $\mu m-s$ ,  $C_S$  a source correction, and  $C_D$  a frequency-dependent distance correction. For details of the correction terms, see Okal and Talandier (1989 and 1990).

Best results are achieved for  $Mw > 6$  at distances  $> 15-20^\circ$  although the Mm procedure has been tested down to distances of  $1.5^\circ$  (Talandier and Okal, 1992). However, at  $D < 3^\circ$  the seismic sensors may be saturated in the case of big events. Also, at short distances one may not record the very long periods required for saturation-free magnitude estimates of very strong earthquakes, and for  $Mw < 6$  the needed very long-period records may become too noisy. A signal-to-noise ratio larger than 3.0 is recommended for reliable magnitude estimates.



Mm determinations have been automated at the Pacific Tsunami Warning Center (PTWC) and the Centre Polynésien de Prévention des Tsunamis (CPPT) (Weinstein and Okal, 2005; Hyvernaud et al., 1993) so that values are available in near real-time within about 10 min after OT from near stations, however typically within about half an hour, plus a few minutes more for really great earthquakes measured at the longest periods. Mm does not – or only marginally – saturate, even for very great, slow or complex earthquakes.

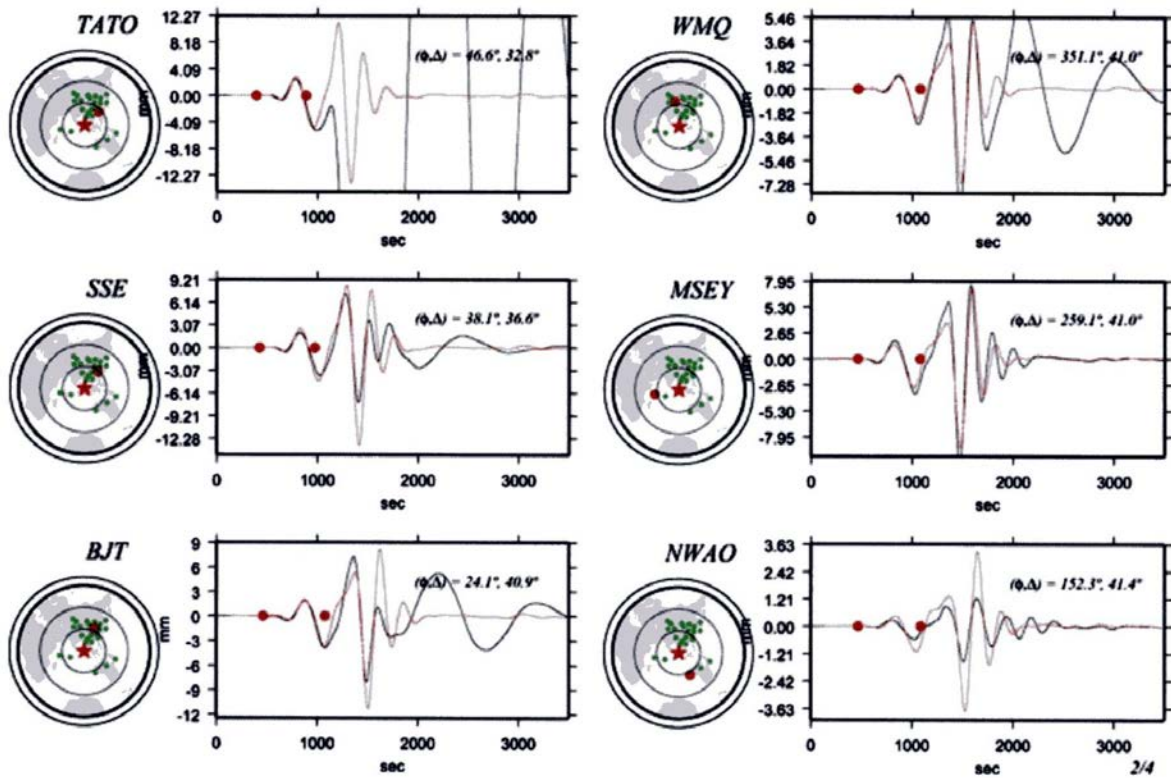
Applications of Mm to the reassessment of the moment of shallow, intermediate and deep historical earthquakes are extensively described by Okal (1992 a and b). For the Chile 1960 earthquake Okal (1992a) calculated values  $M_m \approx 10$  to  $10.3$  and for  $M_0 = 3.2 \cdot 10^{23}$  Nm. Mm determinations were extensively verified and are said to be accurate by about  $\pm 0.2$  magnitude units (Hyvernaud et al., 1993).

### **3.2.8.6 W phase rapid Mw estimates and moment tensor solutions**

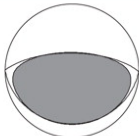
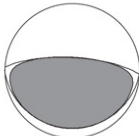
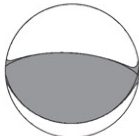
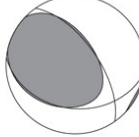
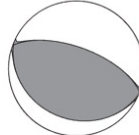
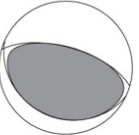
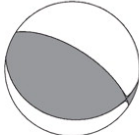
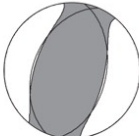
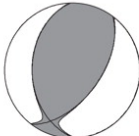
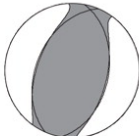
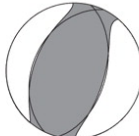
In section 2.6.6 of Chapter 2 the nature and appearance of the very long-period W phase between the P and S-wave onsets have been shortly described (see Figs. 2.74 and 2.75). The inversion of the W phase for sufficiently strong earthquakes by optimizing the fit between W phase synthetics to empirical records (see Fig. 3.66) yields rather rapid (within about 25 minutes after the earthquake occurrence) and reliable first seismic moment tensor solutions and thus of both Mw (within about 0.1 m.u.) and the source mechanism of strong earthquakes (see Fig. 3.67). Therefore, such applications have recently gained immense practical importance for speeding up seismic tsunami warning, as proposed by Kanamori and Rivera (2008). First real-time implementations are operational at the U.S. Geological Survey's NEIC (Hayes et al., 2009) as well as at the GFZ Potsdam and the Indonesian Tsunami Early Warning System (InaTEWS) since 2011.



**Sumatra\_2004 (0.001 Hz - 0.005 Hz,  $n = 4$ )**



**Fig. 3.66** Examples of fitting synthetic W waveforms (thick traces) to very long-period (200 to 1000 s) filtered observed waveforms at several seismic stations (thin traces). The two red dots on each trace indicate the time window over which the W-phase is inverted. From the overall best fitting solution the moment tensor parameters are estimated (Cut-out of Figure 11 of Kanamori and Rivera (2008); © Geophysical Journal International).

|                                                                                                                                   | W-Phase, 50°<br>Optimized Centroid                                                                                                         | W-Phase, 90°<br>Optimized Centroid                                                                                                        | gCMT Solution                                                                                                                        | W-Phase, 90°<br>Optimized Centroid<br>Denser Grid Search                                                                                    |
|-----------------------------------------------------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------|
| <b>2008/11/16</b><br><b>Minahasa, Indonesia</b><br>Dist., (3)-(1) ~28 km<br>Dist., (3)-(2) ~28 km                                 | <br>11 Stations<br>$t_h = t_d = 14.2$ s<br>$M_w = 7.31$   | <br>37 Stations<br>$t_h = t_d = 11.4$ s<br>$M_w = 7.38$  | <br>$t_d = 12.3$ s, $t_h = 11.0$ s<br>$M_w = 7.3$  |                                                                                                                                             |
| <b>2008/11/24</b><br><b>Sea of Okhotsk</b><br>Dist., (3)-(1) ~14 km<br>Dist., (3)-(2) ~14 km                                      | <br>12 Stations<br>$t_h = t_d = 5.2$ s<br>$M_w = 7.30$    | <br>57 Stations<br>$t_h = t_d = 5.2$ s<br>$M_w = 7.28$   | <br>$t_d = 11.9$ s, $t_h = 10.8$ s<br>$M_w = 7.3$  |                                                                                                                                             |
| <b>2009/01/03, 19:43 PM</b><br><b>Papua, Indonesia</b><br>Dist., (3)-(1) ~59 km<br>Dist., (3)-(2) ~50 km<br>Dist., (3)-(4) ~11 km | <br>19 Stations<br>$t_h = t_d = 10.1$ s<br>$M_w = 7.52$   | <br>38 Stations<br>$t_h = t_d = 12.0$ s<br>$M_w = 7.56$  | <br>$t_d = 15.4$ s, $t_h = 15.9$ s<br>$M_w = 7.6$  | <br>38 Stations<br>$t_h = t_d = 20.0$ s<br>$M_w = 7.59$  |
| <b>2009/01/03, 22:33 PM</b><br><b>Papua, Indonesia</b><br>Dist., (3)-(1) ~27 km<br>Dist., (3)-(2) ~27 km                          | <br>19 Stations<br>$t_d = 4.4$ s<br>$M_w = 7.39$         | <br>30 Stations<br>$t_h = t_d = 6.0$ s<br>$M_w = 7.36$  | <br>$t_d = 3.5$ s, $t_h = 12.0$ s<br>$M_w = 7.4$  |                                                                                                                                             |
| <b>2009/01/15</b><br><b>Kuril Islands</b><br>Dist., (3)-(1) ~23 km<br>Dist., (3)-(2) ~45 km<br>Dist., (3)-(4) ~26 km              | <br>11 Stations<br>$t_h = t_d = 10.1$ s<br>$M_w = 7.41$ | <br>43 Stations<br>$t_h = t_d = 8.3$ s<br>$M_w = 7.40$ | <br>$t_d = 9.8$ s, $t_h = 12.0$ s<br>$M_w = 7.4$ | <br>43 Stations<br>$t_h = t_d = 8.3$ s<br>$M_w = 7.39$ |

**Fig. 3.67** Examples of operationally derived source parameters from W phase inversion in comparison to global Centroid Moment Tensor (gCMT) solutions. Solutions derived from station records at up to 50° epicentral distance were on average available 24 min after OT and those from stations up to 90° epicenter distance within about 48 min. Optimized denser grid search takes more time with no significant difference. Copy of Figure 3 of Hayes et al. (2009), Seismological Research Letters, Vol. 80, No. 5, page 821; © Seismological Society of America.

### 3.2.9 Relationships among different magnitude scales

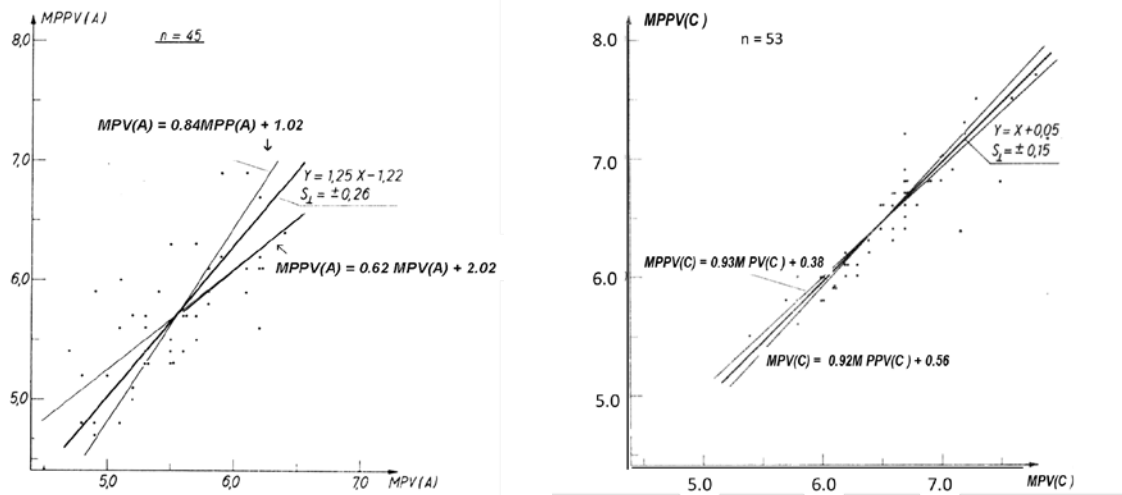
(P. Bormann, D. Di Giacomo and S. Wendt)

#### 3.2.9.1 Remarks about regression and correlation relationships (P. Bormann)

Most of the published classical relationships between different types of magnitudes are simple linear standard regressions of type one (SR1), also termed ordinary least-square (OLS) regression. But often authors have not even specified in an unambiguous way the type of regression or data correlations they have applied. Yet, least-square fits may be done in very different ways. Sometimes, even eye-fits are not the worst, because they intuitively accept that the data scatter relates to both compared variables. The approximate relationship (3.3) is such a best eye-fit through a largely scattering data cloud yielding a very simple and sufficiently reliable formula in the context of the great variability of individual data points. Also the famous Gutenberg-Richter relationship between  $\log E_s$  and  $m$  (respectively  $m_B$ ) shown in Fig. 3.83 is obviously just an eye-fit but never stated explicitly as such. Moreover, some relationships, such as the  $\log E_s$ - $M_s$  relationship by Choy and Boatwright (1995), are just a least-square fit with respect to a prescribed slope, namely to the slope of 1.5 in the famous Richter (1958)  $\log E_s$ - $M_s$  relationship. The reason was to assure compatibility of the new relationships with already since long widely used and well established relationships, at least with respect to some essential parameters such as the slope or general trend. Yet, proper linear regression through the same data set would yield both another slope and another constant (see section 3.3.3).

Standard regressions are least-square fits through the data cloud, which consider the scatter in plots of the  $(x_i, y_i)$  data pairs as being related solely to the dependent ordinate variable  $Y$  while assuming that the errors of the “independent” (or given) abscissa variable  $X$  are zero or negligibly small. In contrast, the inverse standard regression ISR (or standard regression SR2), is based on the opposite assumption that only  $X$  is afflicted with initial errors, but not  $Y$ . Accordingly, the slopes of SR1 and SR2 will differ, the more the larger the scatter of the data points and the smaller their correlation co-efficient.

Since generally all types of measured/calculated magnitude values are afflicted with initial errors, such standard regression relationships are not optimal. They can only be used in the way they have been derived, i.e., for converting a given magnitude, assumed to be error free, into another one by estimating its average value and related uncertainty, such as its standard deviation. In fact, a standard regression projects the actual errors in the given variable into an increased error of the estimated dependent variable. Therefore, resolving an SR1 relationship  $Y$  over  $X$  for the variable  $X$ , as sometimes practiced when converting magnitudes, is not correct. Such a misuse of standard regressions easily results in conversion errors of 0.2 – 0.3 magnitude units (Gutdeutsch et al., 2011) but may become, in the case of rather noisy, especially short-period magnitude data, as large as about 1 m.u. in some magnitude ranges [e.g. in Fig. 3.68 left and conclusion drawn from comparing Eq. (3.78a) and (3.78d)], unless the initial errors of both variables are rather small and the correlation coefficient is close to 1 so that the difference in slope between SR1 and SR2 in the same  $X$ - $Y$  plot becomes negligible (see Fig. 3.68 right).



**Fig. 3.68** Standard regressions SR1 of vertical component PP-wave magnitudes over vertical component P-wave magnitudes, measured on either short-period records of type A (left) or broadband records of type B (right), which correspond to type A4 and C, respectively, in Fig. 3.20 left). SR1 is compared with the respective inverse standard regression SR2 of MPV over MPPV and the common orthogonal standard regression OSR, written in the form Y over X (see thick solid line between SR1 and SR2). Note the different slopes of SR1 and SR2, depending on the data scatter, which is much less for the broadband versions of MPV and MPPV.  $S_{\perp}$  = orthogonal standard deviation. From Figures 3 and 4 in Bormann and Khalturin (1975).

The simplest way to mitigate this problem is to calculate orthogonal standard regressions, OSR, assuming that both variables are afflicted with initial errors of about the same size (ideally with an error ratio 1). This has been practiced, e.g., by Bormann and Khalturin (1975), Bormann and Wylegalla (1975), Ambraseys (1990), Gusev (1991), Gutdeutsch et al. (2002), Grünthal and Wahlström (2003), Bormann et al. (2007 and 2009), Ristau (2009) and Das et al. (2011). In the two papers by Bormann et al. (2007 and 2009), the OSR solutions are deliberately presented together with SR1 and SR2 in order to see the differences and their dependence on the correlation coefficient (e.g., Figs. 3.49 and 3.50). In another publication, Castellaro and Bormann (2009) could show, that as long as the true error ratio, or the square root of the true variance ratio  $\eta_{\text{true}}$ , of the two variables ranges between  $0.7 < \sqrt{\eta_{\text{tru}}} < 1.8$  the orthogonal standard regression under the assumption of  $\eta = 1$  is still superior to SR1 or SR2. This condition is usually fulfilled when comparing magnitudes that have been measured at comparable periods. But outside of this error ratio range, either SR1 or ISR = SR2 represent the data relationship better, depending also on how much their slope deviates from 1.

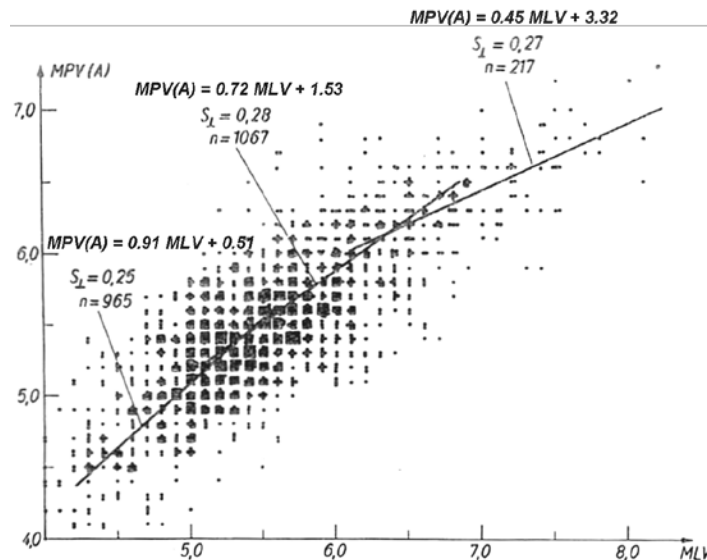
In reality, however,  $\sqrt{\eta}$  may become as small as about 0.3, e.g. when comparing short-period mb with typical initial errors in the range between 0.2 and 0.3 (see also Table 5 in IS 3.3), with Mw(GCMT), for which initial errors are assumed to be not larger, often even less than 0.1 m.u. (personal communication by G. Ekström, 2009; see also Helfrich, 1997; Kagan, 2003 and 2003; Gasperini et al., 2012 and 2013a). Gasperini et al. (2013b) assumed, e.g., for mb determinations an average initial error of 0.25 m.u. and for Mw of 0.1 m.u., corresponding to  $\sqrt{\eta} = 0.4$  or  $\eta = 0.16$ .

Regrettably, in most classical papers the initial measurement errors of used event magnitudes are not given, probably not even known, also because the data reports, catalogs or bulletins of the de facto seismological World Data Centers such as the USGS-NEIC, the Global Centroid Moment Tensor (GCMT) project, or the International Seismological Centre (ISC) did not provide consistently such information. This problem is expected to disappear at the ISC thanks to the implementation of the new ISC locator (Bondár and Storchak, 2011) which computes  $M_s$  and  $m_b$  with uncertainties. Therefore, usually one could only estimate the approximate error ratio by own investigations. These may, however, yield different average measurement errors (standard deviations) even for equal types of magnitudes, depending on differences in instrument calibration, record reading accuracy, aperture and station site distribution of the network, regional variability of station site effects due to variable topographic and geologic conditions, etc. Table 5 in IS 3.3 compares such estimates for single expert readings of event magnitudes from records of the German Regional Seismic Network with those of an analyst team at the China Earthquake Network Center (CENC) based on records of the much larger China National Seismic Network.

If  $\sqrt{\eta}$  differs from 1 and one knows sufficiently well this average error ratio, then a general orthogonal regression (GOR) (for procedures see Fuller, 1987; Castellaro et al., 2006; Das et al., 2011; Lolli and Gasperini, 2012) would be closer to optimal than OSR. Yet, there has been another approach in this direction by Stromeier et al. (2004; see also Appendix 2 in Gasperini et al., 2013b), applying a Chi-squared regression for seismic strengths parameter relations. The *Chi-square* method is based on the theory of independent and normally distributed errors and provides a mathematical explanation for the results derived by Castellaro and Bormann (2009). Gutdeutsch et al. (2011) could show, that the results of these two different approaches agree well as long as the mean initial errors are  $< 0.5$  m.u. Also Lolli and Gasperini (2012) investigated in detail the performance of GOR, the Chi-squared regression and the weighted total least squares approach (Krystek and Anton, 2007). Although the formulations of these three approaches appear quite different, the authors showed that, under appropriate conditions, all compute almost exactly the same regression coefficients and very similar formal uncertainties. Thus, the availability of more general and theoretically correct tools for performing regression analysis and deriving conversion relationships between different types of magnitudes demonstrates the importance of knowing and taking into account their true average initial errors. Regrettably, these were commonly unknown so far. Therefore it would be highly desirable to complement current data parameter calculations and documentation at the most relevant seismological data centers accordingly. And in all publications about regression and conversion relationships the specifics of the applied procedure and its compatibility or incompatibility with other approaches should be unambiguously outlined and documented.

One should also be aware, that GOR procedures developed, proposed or applied by different authors may differ significantly. E.g., a modified GOR methodology proposed by Das et al. (2012) and Wason et al. (2012) and applied by Das et al. (2013) for converting  $m_b$  into  $M_w$  values for regional data from Japan, Mexico, the Indian Himalaya region and the Peninsula Indian region claims to reduce significantly both the average absolute difference as well as the standard deviation between the  $M_w$  proxy estimates and observed  $M_w$ (GCMT). Moreover, it is said to be less dependent on the usually not well known initial error ratio  $\eta$ . However, Gasperini and Lolli (2013) argue that the new method is based on some incorrect assumptions, in particular the use as goodness-of-fit criterion of the simple standard deviation (or of the absolute difference) between computed and observed  $M_w$  which would limit the suitability of such methodology and of the regression relations derived therefrom.

Yet, the issue is even more complicated because of the “saturation” tendency of several magnitude scales (see Fig. 3.47). Saturation is the larger the more narrowband the records are on which the amplitudes and related periods are measured and the more the seismographs peak-response frequencies are shorter than the source corner-frequencies. Therefore, linear regression relations have only a limited value when comparing strongly saturating magnitudes (e.g., short-period mb or ML) with long-period Ms or Mw. Then only stepwise linear fits in limited magnitude ranges may yield still reasonable conversion relationships with not to large conversion errors (e.g., Fig. 3.69), or non-linear regression relationships as in Figs. 3.70a and b, 3.81-82 and 3.91-92. Moreover, short-period magnitudes are generally afflicted with larger errors (see Fig. 3.68) due to the larger influence of local site effects and small-scale heterogeneities along the propagation paths. But the larger the measurement errors of the independent (given; to be converted) magnitude are with respect to those of the dependent (to be estimated) magnitude the greater is the need to avoid standard least-square or orthogonal standard regression, linear or non-linear, but to use GOR or Chi-squared regressions instead with reasonable estimates of the error ratio (see Fig. 3.70b). One cannot regress correctly a pronounced non-linear magnitude relationship without accounting for the difference in the measurement errors, which is usually associated with this non-linearity.

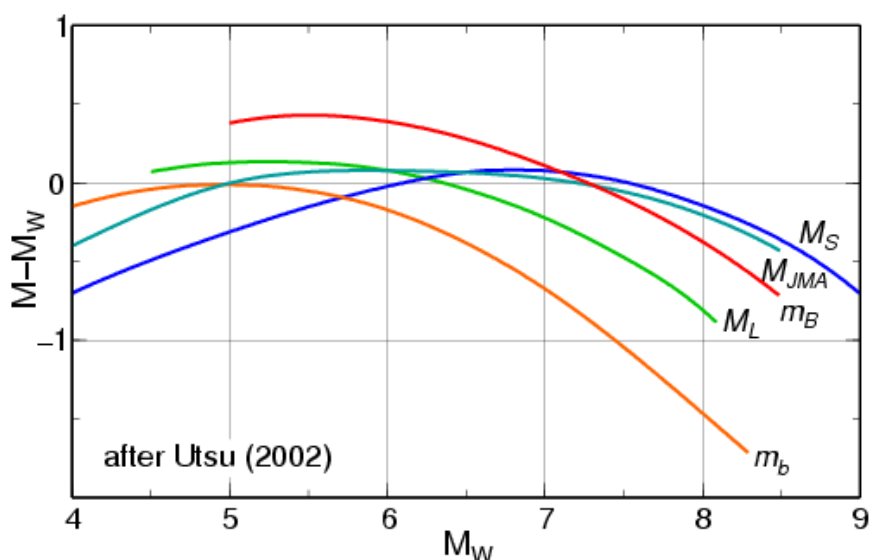


**Fig. 3.69** Step-wise OR regression between vertical-component short-period mb (in a wide measurement time-window up to 1 min) and MLV(C) (of Ms\_BB type) based on magnitude determinations at station MOX, Germany, for global earthquakes. Amended Fig. 5 of Bormann and Khalturin (1975). When extrapolating the uppermost right OR relationship in Fig. 3.69 for  $6.0 < \text{MLV} \leq 8.3$  up to the  $\text{MLV} = \text{Ms\_BB} = 8.9$  measured for the great Sumatra-Andaman Mw9.3 earthquake, one would “predict” an  $\text{mb} = \text{MPV} = 7.3$ , which agrees within 0.1 m.u. with the actual observation.

Non-linear “maximum-likelihood” regressions have been systematically applied for the first time by Gusev (1991) by investigating the relationships between Mw and the magnitudes mb (with  $A_{\text{max}}$  measured within first few seconds only), mSKM (with  $A_{\text{max}}$  measured in the whole P-wave group), mB,  $m_b^*$  (according to Koyama and Zheng, 1985),  $\hat{m}_b$  (according to Houston and Kanamori, 1986), ML, Ms, and M(JMA) in both graphic and tabular form. Some of these

relations are presented in Figure 9 of IS 3.7 in this Manual with respect to  $M_w$  and the Russian  $K = \log E_s$  class system of earthquake classification.

Another form of non-linear average eye-fit relationships of classical magnitudes with respect to  $M_w$  as a reference have also been derived by Utsu (2002) (see Fig. 3.70a), who also discusses extensively the specifics of the types of magnitudes and data, on which these relationships are based.



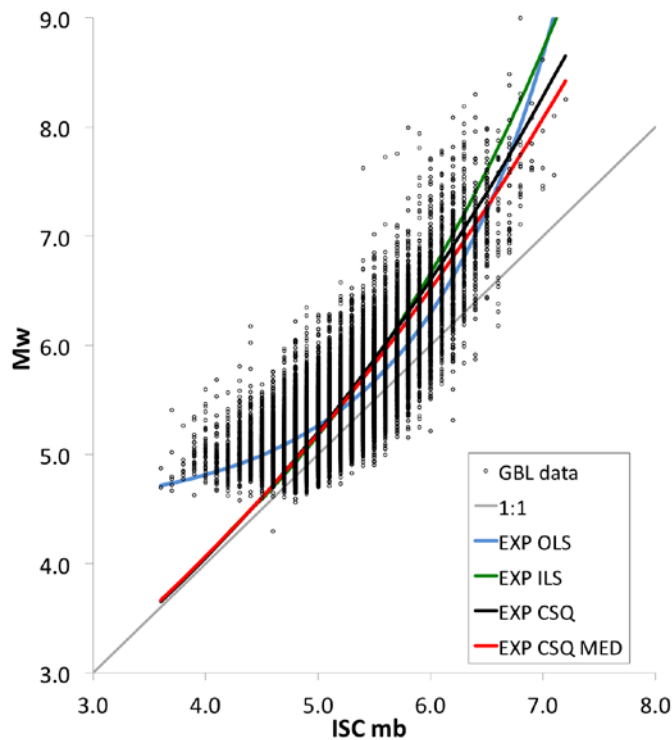
**Fig. 3.70a** Average non-linear relationships between different common types of magnitudes and the moment magnitude  $M_w$ . (Modified from Figure 1 in Utsu (2002): Relationship between magnitude scales. In: Lee W. H. K., Kanamori, H., Jennings, P. C., Kisslinger, C. (Eds) *Earthquake and Engineering Seismology*, Vol. A, 733-746; © Academic Press, London).

In conjunction with the elaboration of the ISC-GEM Global Instrumental Earthquake Catalogue (1900-2009) (see [www.isc.ac.uk/iscgem/citing.php](http://www.isc.ac.uk/iscgem/citing.php)) new empirical relationships were derived by using both linear GOR and exponential non-linear ordinary least-square (OLS) models to obtain  $M_w$  proxies from  $M_s$  and  $m_b$  (Di Giacomo et al., 2013; see Figs. 3.81 and 3.82). The new models were **tested against true values of  $M_w$ (GCMT)**, and the newly derived exponential models were found to be preferable to the linear ones (see section 3.2.9.5 for the data and comparison with GOR regression). Instead of the bi- or tri-linear regression (as in Fig. 3.69), the two data sets were fit with single, continuous regression curves using an exponential model of the form  $M_y = e^{(a+b \times M_x)} + c$ . The regression was performed using the non-linear least square algorithm by Bates and Watts (1988) and Bates and Chambers (1992), which is freely available with the R-language (<http://www.r-project.org/> or <http://stat.ethz.ch/R-manual/R-patched/library/stats/html/nls.html>).

However, Gasperini et al. (2013b) point out that also exponential models, based on OLS fits, i.e., assuming that only the ordinate variable (in this case  $M_w$ ), are not optimal. Moreover, with respect to  $m_b$ - $M_w$  conversion relationships, one has to consider that global centroid moment tensor solutions (GCMT) tend to be incomplete at values below  $M_w(\text{GCMT}) = 5.5$ , but  $m_b(\text{ISC})$  only at about one magnitude unit smaller values. Gasperini

et al. (2013b) suspect therefore that this early truncation or incompleteness of  $M_w$  values results in a biased, too much bended trend relationship with too small slope of  $M_w$  over  $m_b$  towards smaller magnitudes. Moreover, they argue that this bias is particularly strong developed when applying the OLS exponential fit instead of a Chi-squared (CSQ) fit that assumes a realistic error ratio between  $M_w$  and  $m_b$ . Their CSQ fit yields for  $m_b < 4.5-5$  a nearly 1:1 relationship between  $m_b$  and  $M_w$  (see Fig. 3.70b). This is in better agreement with Fig. 3.70a, the relation of  $m_b$  and  $M_l$  with smaller regional  $M_w$  values (see Tab. 3.6 with comment on the  $M_l$ - $M_w$  relationships, Fig. 3.70b for  $M_w$ (MED) data and the related Figure 2 in Gasperini et al., 2013b). A nearly 1:1  $m_b$ - $M_w$  relationship is also supported by theory. According to Fig. 3.5 it is likely that magnitudes based on  $A_{\max}$  measured at frequencies larger than about 0.3 Hz sample spectral plateau amplitudes required for  $M_0$  and thus  $M_w$  determination and therefore do no longer systematically underestimate  $M_w$ .

Fig. 3.70b illustrates the above discussion. Different exponential regression relationships have been applied to an integrated global  $M_w$  data set (GBL) compiled according to Gasperini et al. (2012) and the related  $m_b$  estimates provided by the ISC on-line bulletin. For data sources see Appendix A in Gasperini et al. (2013b). The figure reveals, how different also non-linear regression relations may be when they are based on very different assumptions about the error ratio of the two variables. Note that also the exponential inverse least square (ILS) curve has been plotted. It is even closer to the CSQ fit than the OLS curve (see discussion below). Formulas for the exponential Chi-square regression model are given in Appendix 2 of Gasperini et al. (2013b).



**Fig. 3.70b** Comparison of different exponential (EXP) fits through the Gasperini et al. (2012) global (GBL)  $M_w$  dataset and the related ISC  $m_b$  data. For the Chi-square (CSQ) regression an error ratio  $M_w/m_b$  of 0.4 has been assumed. OLS and ILS stand for the ordinary (SR1) and the inverse (SR2) least-square regressions and MED for a regional (Mediterranean)  $M_w$ - $m_b$  data set down to  $M_w$  3.5 as plotted in Figure 2 by Gasperini et al. (2013b). Figure provided by courtesy of P. Gasparoni and B. Lolli.



Most strikingly in Fig. 3.70b are: a) the large difference between EXP OSL (blue curve) and EXP CSQ solutions for both the GBL (full black curve) and the Mediterranean (MED) Mw-mb data sets (full red curve) when assuming a realistic Mw/mb error ratio and b) the proximity of EXP ISL (full green line) and EXP CSQ.

For understanding these partially surprising results one should note that with respect to a) the low magnitude tail with  $m_b < 4.5$  is constrained by only about 10% of the data and no histogram equalization has been applied as by Di Giacomo et al. (2013) and b) that Castellaro and Bormann (2009) already showed that linear orthogonal standard regressions (OSR), which assume both variables to be afflicted with initial errors of the same size, are superior to linear OLS or ISL, respectively, only as long as  $0.7 < \sqrt{\eta_{\text{tru}}} < 1.8$  holds. Outside of this error ratio range either OLS or ISL are even better approximations to the data than OSR and thus closer to GOR. This surely applies to the GBL data set with Mw errors estimated to be on average only 0.4 times the initial errors for mb. Accordingly, EXP ISL is already very close to EXP CSQ, the more so since the average data slope in Fig. 3.70b is already larger than 1, i.e., into the direction of the ISL trend.

Therefore, in summary, one should be cautious when calculating and applying magnitude conversion relationships, avoid over-interpretation of their representativeness and thus the accuracy of proxy magnitude estimates, being aware of the assumptions on which the respective regression relations are based as well as of our usually insufficient knowledge whether our data really meet these conditions.

Accordingly, a few general rules and facts should be taken into account before deciding on the most appropriate regression to be applied:

- Single linear regression relations should only be calculated if an eye assessment of the cloud of available data does not indicate a significant non-linearity within the range of data scatter (e.g., Fig. 3.68).
- If the degree of non-linearity in the data is significant, as in Figs. 3.69, 3.81 and especially 3.82, and thus obvious already by simple eye check of the data, then there exist two opportunities: a) approximating the data in sub-segments by two or more linear curves of different slope, as in Fig. 3.69 or 3.81 or b) by a non-linear regression.
- In the case of modest non-linearity, as in Fig. 3.81, option a) may still yield reasonably good conversion results for practical applications with acceptable average biases  $< 0.1$  m.u. in most parts of the chosen magnitude ranges. However, conversion results will be inconsistent near the discontinuity points where the linear curves with significantly different slope intersect. Therefore, option b), i.e., calculating continuous non-linear regression relations, is generally preferable.
- Non-linear relations between different types of magnitude have two main reasons: a) different degree of spectral saturation, as discussed in conjunction with Figs. 3.5 and 3.45, i.e., when the two magnitudes are measured in ranges of different period and bandwidth and b) if the regression is extended down to magnitudes in which one type is still completely recorded but not the other one, as, e.g., GBL or GCMT Mw for  $M_w < 5.5$  in the case of Fig. 3.70b or Fig. 3.82 (magnitude truncation effect).

- Before deciding on the most appropriate regression procedure one should have an idea about the approximate average measurement errors of the different types of magnitudes compared. Standard least-square regression, which assumes that the chosen independent variable is error-free and all data scatter is related only to the dependent variable is principally not correct. In the case of large data scatter SR1 and SR2 may differ significantly. Therefore, standard regression relationships should not be resolved for the independent variable but used only in the way as they have been derived. If, however, the data scatter is small and the correlation coefficient very large than the difference between SR1 and SR2 may become negligibly small (e.g., in Figs. 3.68 or 3.71, upper right).
- If one has good reasons to assume that both types of magnitudes are measured with similar errors, which is commonly the case for magnitudes based on amplitudes that are measured at comparable periods in similar bandwidth ranges, then orthogonal standard regression (OSR) yields more reliable regression relations than OLS. Yet in the case that the error ratio is  $> 1.8$ , respectively  $< 0.7$ , then SR1, respectively SR2 = ISR, yield better average estimates of the respective proxy magnitude than OSR (see Castellaro and Bormann, 2007). However, in any event, GOR or Chi-squared regressions, linear or non-linear, are in such cases more appropriate, provided that sufficiently reliable average error ratio estimates for the compared magnitudes are available.
- A very important point is to assure, as far as possible, the homogeneity of the regressed magnitude data with respect to the procedures and accuracy with which they were measured over the considered time span and for the considered source area(s). The latter may range from local to regional and global.
- In order to assure that the regression relationship is not biased by the very different data frequency and thus weight in different magnitude ranges histogram equalization should be applied, if sufficient data are available (see Di Giacomo et al., 2013). This has usually not been done.
- Also usually not done, but of utmost importance, is to test the validity of the obtained regression model with values not used in obtaining the model (see Di Giacomo et al., 2013b). Neglecting this may lead to unjustified claims with respect to the accuracy and representativeness of the derived relationships and bears the risk of significantly biased magnitude proxy estimates.
- The accuracy and representativeness of magnitude relationships largely depends on the amount, accuracy, homogeneity and completeness of available data. However, huge amounts of data in a wide magnitude range may be available only when considering large time spans and source areas (global as in section 3.2.9.5, or for regions with very high seismicity). On the other hand, over large time spans magnitude procedures and the accuracy of measurements have rarely been homogeneous and often differed also regionally. Improving the homogeneity and completeness of such long-term global data sets may require labor and time intensive critical data checks and re-computations as practiced in conjunction with the ISC-GEM project for creating a Global Instrumental Earthquake Catalogue (1900-2009) (Storchak et al., 2013; Di Giacomo et al., 2013 a and b).

- On the other hand, relationships between different types of magnitude may vary significantly for different regions. Examples will be given in the following. Therefore, there is a need to derive magnitude regression and conversion relationships also on a regional basis in conjunction with investigations of regional seismicity and seismic hazard. Yet, for many such areas the limited amount of available data, often only for a limited time span and magnitude range, may reduce both their reliability and representativeness for wider areas, time spans and magnitude ranges. Therefore, these “parameters” should always be specified in conjunction with such relationships.
- Nonetheless, we will show in the following also examples for early magnitude relationships that are based on rather limited but homogeneously and competently measured data, sometimes just at a single station or a regional network but still agree for globally distributed events astonishingly well with much more recent relationships based on many more data from global network stations.

### 3.2.9.2 Summary of classical magnitude relationships (P. Bormann)

Gutenberg and Richter (1956a, b) provided correlation relations between various magnitude scales:

$$m = m_B = 2.5 + 0.63 M_s \quad (3.76a)$$

$$m = m_B = 1.7 + 0.8 M_l - 0.01 M_l^2 \quad \text{and} \quad (3.76b)$$

$$M_s = 1.27 (M_l - 1) - 0.016 M_l^2, \quad (3.76c)$$

where  $m$  is the unified magnitude as the weighted mean of the body-wave magnitudes  $m_B$  determined from medium-period broadband recordings. Relationship (3.76a) has become crucial for deriving the Richter (1958)  $\log E_s$ - $M_s$  relationship and via the latter also the  $M_w$  formula by Kanamori (1977) and Hanks and Kanamori (1979).

Practically the same relationship as (3.76a) but based on some more (yet still less than 200 data points) in a wider range of magnitudes between  $5.5 \leq m_B \leq 8.2$  was derived later by Abe and Kanamori (1980):

$$m_B = 2.5 + 0.65 M_s. \quad (3.77)$$

It has only a slightly larger slope than (3.76a) and satisfies the  $m_B$  and  $M_s$  data well between magnitude 5.2 and 8.7.

All these above regression relations are of the SR1 type. Sometimes they were wrongly applied, e.g. in a paper reviewed by the editor, where the author solved Eq. (3.70) for  $M_s$ , then converting short-period  $m_b$  values as published by international data centers into  $M_s$  and finally calculated with these proxy  $M_s$  values the released seismic energy  $E_s$  via the Richter (1958)  $E_s$ - $M_s$  relationships (3.66). Yet, Eq. (3.76a) is suitable only for converting  $M_s$  into  $m_B$  and then applying the only energy-magnitude relationship  $\log E_s$ - $m_B$  which Gutenberg favored [see Eq. (3.95) in section 3.3.2], because he was aware that  $m_B$  determined via  $(A/T)_{\max}$  in a wide range of periods is a much better estimator of released seismic energy than the long-period spectral magnitude  $M_s$  measured at periods around 20 s only.

When using **medium-to-long-period readings of P and surface waves in displacement broadband records of type C (Kirnos SKD)**; see Fig. 3.20) and **SR1 regression**, practically identical relationships to Eq. (3.76a) were found both by Bune et al. (1970) on the basis of records of the former **USSR-wide station network** and by Bormann and Wylegalla (1975) for the **single station MOX** in Germany (magnitude range 4.7 to 8.5). The latter is

$$MPV(C) = 2.5 + 0.60 MLH(C). \quad (3.78a)$$

Relationship (3.76a) could also be reproduced more than 60 years later with practically identical SR1 relationships based on many more modern IASPEI standard mB\_BB and Ms\_BB measurements on velocity broadband readings of globally distributed earthquakes that were recorded by stations of both the China National Seismographic Network (CNSN) and the German Regional Seismic Network (GRSN) (see Bormann et al., 2009 and section 3.2.9.3).

Yet, the related **orthogonal regression** to Eq. (3.78a), calculated for the same data set, is rather different:

$$MPV(C) = 1.83 + 0.70 MLH(C) \quad (3.78b)$$

and the respective best fitting SR2 regression is

$$MLH(C) = -1.54 + 1.25 MPV(C). \quad (3.78c)$$

The latter is clearly different from

$$MLH(C) = -4.17 + 1.67 MPV(C) \quad (3.78d)$$

when resolving incorrectly Eq. (3.78a) for MLH. This would result in an overestimation of MLH by about 1.2 magnitude units for mB = 8 and an underestimation of 0.8 units for mB = 5.

The single random-parameter regression relationship between short-period mb and Ms based on **global WWSSN station network and earthquake data** analyzed at NOAA (with amplitude measurement for mb within the first few cycles only after the P-wave onset) is very different from Eq. (3.76a), namely, according to Gordon (1971),

$$mb = (0.47 \pm 0.02) Ms + (2.79 \pm 0.09). \quad (3.79)$$

This agrees, with a constant offset of -0.16 m.u. (station correction?), with the **single-station** average formula derived by Karnik (1972) for the Czech station Pruhonice (PRU):

$$mb(sp, PRU) = 0.47 MLH + 2.95.$$

On the other hand, when approximating vertical component short-period P-wave MPV(A) values plotted over the related MLV(C) values, both measured for globally distributed earthquakes at the **single station** MOX, Germany, in the wide magnitude range  $4.0 \leq \text{MLV(C)} \leq 8.3$  by shorter overlapping tri-linear orthogonal regression relationships (see Fig. 3.69), Bormann and Khalturin (1975) derived

$$\text{MPV(A)} = 0.91 \text{ MLV(C)} + 0.51 \quad \text{between } 4.0 \leq \text{MLV(C)} \leq 5.5 \quad (3.80a)$$

$$\text{MPV(A)} = 0.72 \text{ MLV(C)} + 1.53, \quad \text{between } 5.1 < \text{MLV(C)} \leq 6.8 \quad (3.80b)$$

and

$$\text{MPV(A)} = 0.45 \text{ MLV(C)} + 3.32 \quad \text{between } 6.1 < \text{MLV(C)} \leq 8.3. \quad (3.80c)$$

The difference between short-period mb and long-period Ms strongly depends on the specifics of the seismic source process, e.g., on differences source size, source-time function, source depth, stress drop and thus radiated seismic spectra, especially the relative ratio of long-period to short-period spectral amplitudes. Therefore, plotting (20 s) Ms over (1 s) mb has become a powerful discriminator between natural earthquakes and underground nuclear explosions (see, e.g., Chapter 11, section 11.2.5.2???, Fig. 11.22???; Figure 5 in Richards, 2002; Figure 2 in Richards and Wu, 2011). A relationship

$$\text{Ms} = 1.25 \text{ mb} - 2.45 \quad (3.81)$$

separates almost all shallow earthquakes globally from underground nuclear explosions (Richards, 2002). Interestingly, its slope is practically identical with the average slope of 1.23 in the Ms range between 4.0 and 6.8 when resolving the orthogonal relationships (3.80a and b) for  $\text{MLV} = \text{Ms}$ . However, the negative constant in (3.81) is 1.34 m.u. larger in order to account for the large data scatter around the average Ms-mb relationship for earthquakes.

Being aware of the even more general potential of Ms-mb, Prozorov and Hudson (1974) proposed a *creepex parameter*

$$c = \text{Ms} - a \times \text{mb} - b \quad (3.82)$$

with a and b being constants to be determined empirically for different source types, stress-drop conditions and seismotectonic regions. *Creep* stands for very slow rupture motions (see *slow* and *silent earthquakes* in the Glossary) and *ex* for explosions or other small seismic sources with extremely short source-time functions. Thus, *creepex* aims at discriminating between normal, very slow and explosion-like (fast rupture high-stress-drop) earthquakes. World-wide determination of this parameter for earthquakes in different regions revealed interesting relations of c to source-geometry and seismotectonic conditions. Global maps of *creepex* have been published by Prozorov et al. (1983) and Kaverina et al. (1996). Prozorov and Shabina (1984) could show that the regional earthquake catalog for Mexico contains earthquakes with rather different *creepex* values, that are related to three major types of tectonic structures. E.g., *Creepex* is positive in zones of spreading and high heat flow, and negative in areas where relatively cold lithospheric plates are subducted into the mantle. Similar systematic regional differences were also reported for  $\text{Ms} - \text{Mw}$  (Ekström and Dziewonski, 1988; Patton, 1998) and  $\text{Me} - \text{Mw}$  (Choy and Boatwright, 1995; Choy and Kirby, 2004; Di Giacomo et al., 2010a; Bormann and Di Giacomo, 2011). Panza et al. (1993) extended the *creepex* concept to Md-ML and applied it to Italy.

Surface-wave magnitudes determined from vertical and horizontal component recordings using the so-called Prague-Moscow calibration function Eq. (3.35) correlated almost 1:1. According to Bormann and Wylegalla (1975), the orthogonal relationship is

$$MLV - 0.97 MLH = 0.19 \quad (3.83)$$

with a standard deviation of only 0.11 and a correlation coefficient of  $r = 0.98$ . This clearly justifies the use of this calibration function, which was originally derived from horizontal amplitude readings, for vertical component (Rayleigh wave) magnitude determinations, too.

When using medium-period broadband data only, the orthogonal regression relation between magnitude determinations from PV and PPV or SH waves, respectively, are almost ideal. Gutenberg and Richter (1956a) had published Q-functions for all three phases (see Figures 1a-c and Table 6 in DS 3.1). Bormann and Wylegalla (1975) found for a global earthquake data set recorded at station MOX the orthogonal fits:

$$MPPV(C) - MPV(C) = 0.05 \quad (3.84)$$

with a standard deviation of only  $\pm 0.15$  magnitude units (see also Fig. 3.68 right) and

$$MSH(C) - 1.1 MPV(C) = -0.64, \quad (3.85)$$

with a standard deviation of  $\pm 0.19$  and magnitude values for P and S waves, which differ in the whole range of MPV(= mB) between 4 and 8 less than 0.25 units from each other. This confirms the good mutual scaling of these original body-wave calibration functions with each other, provided that they are correctly applied to medium-period data only. Regrettably, international data centers do no longer encourage data producers to report also amplitudes from PPV and SH waves for the determination of mB, which was Gutenberg's original intention.

Kanamori (1983) summarized in graphical form the relationship between the various magnitude scales (see Fig. 3.47 in section 3.2.5.2 above). He also gives the ranges of uncertainty for the various magnitude scales due to observational errors and intrinsic variations in source properties related to differences in stress drop, complexity, fault geometry and size, source depth etc. The range of periods for which these magnitudes are usually determined are for mb:  $\approx 1$  s; for MI:  $\approx 0.1 - 3$  s; for mB:  $\approx 0.5 - 25$  s; for Ms:  $\approx 20$  s and for Mw:  $\approx 10 \rightarrow \infty$  s. Accordingly, these different magnitude scales saturate differently: the shorter the measured periods the earlier "saturation", respectively underestimation of magnitude, occurs, i.e., according to the Kanamori graph for mb on average around 6.5, for MI around 7, for mB around 8 and for Ms around 8.5 while Mw does not saturate. This is in good agreement with the general conclusions drawn on the basis of seismic source spectra (see Fig. 3.5). For mb, however, "saturation" will happen clearly later, when the maximum amplitudes are measured really within the whole P-wave train, as recommended by the new IASPEI standards, and not within an a-priori fixed limited measurement time-window after the first arrival of P.

Ambraseys (1990) and Ambraseys and Bommer (1990), in an effort to arrive at uniform magnitudes for European earthquakes, re-evaluated magnitudes in the range  $3 < M < 8$  and

published in 1990b, with more data than in 1990a, the following updated *orthogonal regression relationships* between the various common magnitude scales:

$$0.76 \text{ mb} - 0.66 \text{ mB} = 0.31 \quad (3.86)$$

$$0.75 \text{ mb} - 0.66 \text{ Ml} = 0.51 \quad (3.87)$$

$$0.87 \text{ mb} - 0.50 \text{ Ms} = 1.91 \quad (3.88)$$

$$0.82 \text{ Ml} - 0.58 \text{ Ms} = 1.20 \quad (3.89)$$

with mb being determined according to the ISC procedure from short-period P-wave recordings and mB using medium-period P-wave records. These relations can be solved for either one of the two variables. Other relationships have been published by Nuttli (1985) which allow estimating Ms for plate-margin earthquakes when mb is known. For mb > 5 their results differ less than 0.2 magnitude units from those of Eq. (3.88) when solved for Ms.

The most recent GOR conversion relationships between more than 70.000 global ISC mb and Ms data during the period 1976-2007 have been published by Wason et al. (2012). This relationship reads, when rounded to the nearest second decimal:

$$\text{Ms} = 1.87 \text{ mb} - 4.44 \quad (3.90)$$

with a correlation coefficient  $R_{XY} = 0.84$  and a root mean square  $\text{RMSO} = 0.24$ . This relationship is rather close to the OSR relationship derived by Bormann et al. (2009) (next section) but differs more from Ambraseys and Bommer's (1990) OSR relationship (3.88) for Europe and adjacent areas, when resolved for Ms:

$$\text{Ms} = 1.74 \text{ mb} - 3.02. \quad (3.91)$$

In the latter relationship Ms is of the Ms\_BB type, measured also at regional distances down to 4° and at much shorter than 20 s periods which results in larger Ms values than the dominatingly 20 s Ms(ISC). At mb = 4 the difference is already 0.9 m.u.

Finally, we refer to a recent work by Ristau (2009), who gives both OSR, SR1 and SR2=ISR relationships between mb and Ml for all depth, shallow ( $h < 30 \text{ km}$ ) and deeper earthquakes in New Zealand. The average OSR relationship for all events reads:

$$\text{mb} = (0.94 \pm 0.07) \text{ Ml} - (0.06 \pm 0.07) \quad (3.92)$$

with a distinct difference in slope (1.11 and 0.79, respectively) as well as scatter for shallow and deeper earthquakes.

### 3.2.9.3 Linear orthogonal and standard regressions between IASPEI standard magnitudes (P. Bormann, D. Di Giacomo and S. Wendt)

Bormann et al. (2009) derived the first OSR relationships between the four IASPEI standard magnitudes mb, mB\_BB, Ms\_20 and Ms\_BB. They are based on data of some 400 to 500 globally distributed earthquakes that have been recorded by the China National Seismographic Network (CNSN) and analyzed at the China Earthquake Network Center (CENC) according to the standard procedures recommended by IASPEI (2005 and 2011) and in IS 3.3. Fig. 3.71 and Tab. 3.4 summarize the results. In the Table, besides the OSR

relations, also the related SR1 and SR2 solutions are given with an  $\leftarrow$  instead of  $=$  sign in order to indicate that these relations can only be solved in this direction but not in the opposite one.

With reference to the Das et al. (2012) GOR relation (3.90) the standard  $M_s$ -20-mb OSR relation looks similar and reads:

$$M_s_{20} = 1.82 \text{ mb} - 4.75 \quad (3.93)$$

with  $R_{XY} = 0.891$  and an  $RMSO = 0.21$ . The somewhat steeper slope with a smaller constant in (3.90) is due to the fact that  $mb(ISC)$  saturates already around 7, standard  $mb$ , however, only around 7.5. Accordingly, an  $mb(ISC) = 7$  relates with (3.90) already to an  $M_s = 8.6$ , a standard  $mb = 7$ , however, according to (3.93), only to an  $M_s = 8.0$  but  $mb = 7.5$  to  $M_s = 8.9$ .

However, of much greater importance than the rather noisy and in fact rather non-linear (see section 3.2.9.5) relationship between short-period  $mb$  and  $M_s$  is the relationship between the new standard broadband  $mB_{BB}$  and  $M_s_{BB}$  with respect to the fundamental relationship between  $mB$  and  $M_s$  published by Gutenberg and Richter (1956a and b):

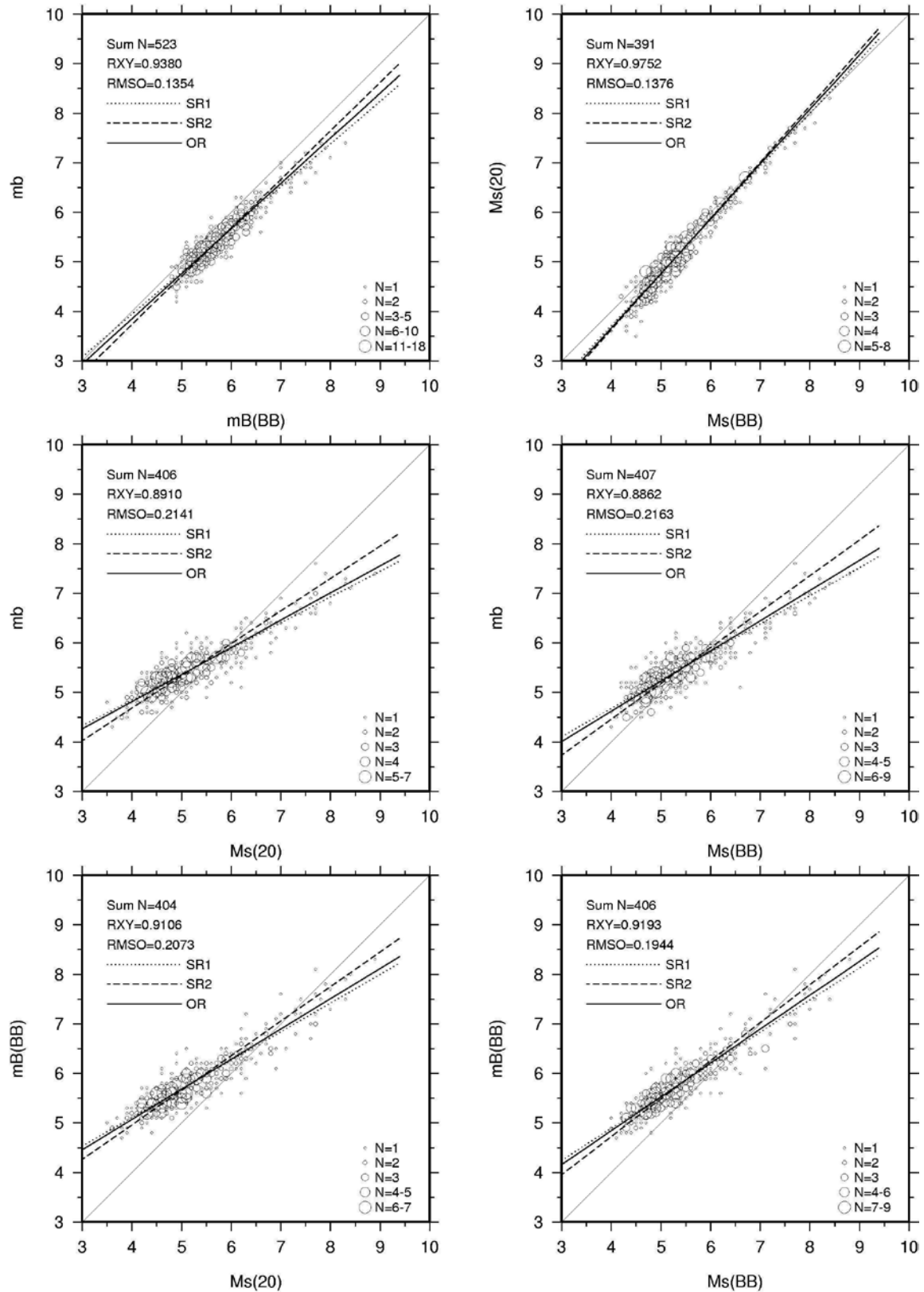
$$mB = 0.63 M_s + 2.5. \quad (3.94)$$

This SR1 relationship is applicable for magnitudes between about 6 and  $<8.5$ . From it follows that  $mB = M_s$  at magnitude 6.75, yet  $mB$  is on average larger at smaller values of  $M_s$  and smaller at larger  $M_s$ . When substituting  $mB$  by (3.94) in the primary Gutenberg-Richter (1956 and b) relationship between body-wave magnitude and released seismic energy (with  $E_s$  in Joule):

$$\log E_s = 2.4 mB - 1.2 \quad (3.95)$$

one arrives at the relationship (3.66) proposed by Richter (1958) (there with a typo) which has been instrumental in the derivation of both the moment magnitude and energy magnitude scales (see section 3.2.7 of this Chapter and Kanamori, 1977; Hanks and Kanamori, 1979; Choy and Boatwright, 1995; Bormann and Di Giacomo, 2011). Accordingly, the whole foundation of modern earthquake size, respectively strength classification essentially rests on it. Therefore, it is of great interest to know, whether the now proposed new IASPEI broadband magnitude standards  $mB_{BB}$  and  $M_s_{BB}$ , which differ slightly from the original Gutenberg  $mB$  and  $M_s$ , can reproduce this relationship nowadays with many more data in a wider magnitude range being now available than at Gutenberg's time.





**Fig. 3.71** Regression relations between new IASPEI standard magnitudes measured at CENC with respect to the ideal 1:1 line. N = number of (x,y) pairs, RXY = correlation coefficient, RMSO = orthogonal root-mean square error, OR = OSR, SR2 = ISR. For formulas see Tab. 3.4. Copy of Figure 11 by Bormann et al. (2009); © Seismological Society of America.

**Tab. 3.4** Relationships between magnitudes determined from global earthquake CNSN data according to the IASPEI (2005) measurement standards. The unresolved OSR with RMSO are printed in bold letters, the rms for the resolved OSR are the uncertainty (standard error) of the estimated magnitude (modified and amended according to Bormann et al. (2009).

| $M_x - M_y$<br>N     | Regression                          | Relationship                                                                                                                                                                                  | rms<br>Error                                                                                           | Figure 3.71  |
|----------------------|-------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------------------|--------------|
| mB_BB - mb<br>523    | SR1<br>SR2<br><br><b>OSR</b>        | mb $\leftarrow$ 0.86 mB_BB + 0.50<br>mB_BB $\leftarrow$ 1.02mb + 0.18<br><b>0.15 = -0.67 mB_BB + 0.74mb</b><br>mb = 0.91 mB_BB + 0.20<br>mB_BB = 1.10mb - 0.22                                | $\pm 0.18$<br>$\pm 0.20$<br><b><math>\pm 0.14</math></b><br>$\pm 0.19$<br>$\pm 0.21$                   | upper left   |
| Ms_BB - Ms_20<br>391 | SR1<br>SR2<br><br><b>OSR</b>        | Ms_20 $\leftarrow$ 1.07Ms_BB - 0.59<br>Ms_BB $\leftarrow$ 0.89Ms_20 + 0.79<br><b>0.50 = 0.74Ms_BB - 0.67Ms_20</b><br>Ms_20 = 1.10Ms_BB - 0.75<br>Ms_BB = 0.91Ms_20 + 1.47                     | $\pm 0.20$<br>$\pm 0.19$<br><b><math>\pm 0.14</math></b><br>$\pm 0.21$<br>$\pm 0.19$                   | upper right  |
| Ms_20 - mb<br>406    | SR1<br>SR2<br><br><b>OSR</b>        | mb $\leftarrow$ 0.52Ms_20 + 2.77<br>Ms_20 $\leftarrow$ 1.53mb - 3.15<br><b>2.29 = -0.48Ms_20 + 0.88mb</b><br>mb = 0.55Ms_20 + 2.61<br>Ms_20 = 1.82mb - 4.75                                   | $\pm 0.24$<br>$\pm 0.41$<br><b><math>\pm 0.21</math></b><br>$\pm 0.24$<br>$\pm 0.44$                   | middle left  |
| Ms_BB - mb<br>407    | SR1<br>SR2<br><br><b>OSR</b>        | mb $\leftarrow$ 0.57Ms_BB + 2.40<br>Ms_BB $\leftarrow$ 1.38 mb - 2.15<br><b>1.86 = -0.52 Ms_BB + 0.85mb</b><br>mb = 0.61Ms_BB + 2.18<br>Ms_BB = 1.61mb - 3.57                                 | $\pm 0.25$<br>$\pm 0.39$<br><b><math>\pm 0.22</math></b><br>$\pm 0.26$<br>$\pm 0.42$                   | middle right |
| Ms_20 - mB_BB<br>404 | SR1<br>SR2<br><br><b>OSR</b>        | mB_BB $\leftarrow$ 0.58Ms_20 + 2.79<br>Ms_20 $\leftarrow$ 1.43 mB_BB - 3.11<br><b>2.25 = -0.52Ms_20 + 0.85mB_BB</b><br>mB_BB = 0.61Ms_20 + 2.63<br>Ms_20 = 1.61mB_BB - 4.31                   | $\pm 0.24$<br>$\pm 0.38$<br><b><math>\pm 0.21</math></b><br>$\pm 0.24$<br>$\pm 0.40$                   | lower left   |
| Ms_BB - mB_BB<br>406 | <b>SR1</b><br>SR2<br><br><b>OSR</b> | <b>mB_BB <math>\leftarrow</math> 0.65Ms_BB + 2.30</b><br>Ms_BB $\leftarrow$ 1.31 mB_BB - 2.16<br><b>1.74 = -0.56Ms_BB + 0.83mB_BB</b><br>mB_BB = 0.68Ms_BB + 2.10<br>Ms_BB = 1.47mB_BB - 3.09 | <b><math>\pm 0.23</math></b><br>$\pm 0.33$<br><b><math>\pm 0.19</math></b><br>$\pm 0.23$<br>$\pm 0.34$ | lower right  |

The Chinese data presented in Fig. 3.71 (lower right) cover the magnitude range between 4 and 8.8 (with the great majority of magnitudes < 6), i.e., a much larger range than (3.94) or the slightly modified Abe and Kanamori (1980) formula based on strong earthquake records between 1953 and 1977:

$$mB = 0.65 Ms + 2.5. \quad (3.96)$$

None the less, the respective **standard regression SR1** between the two standard broadband magnitudes  $mB_{BB}$  and  $Ms_{BB}$  (**bold blue in Tab. 3.4**) is very similar to both of these formulas :

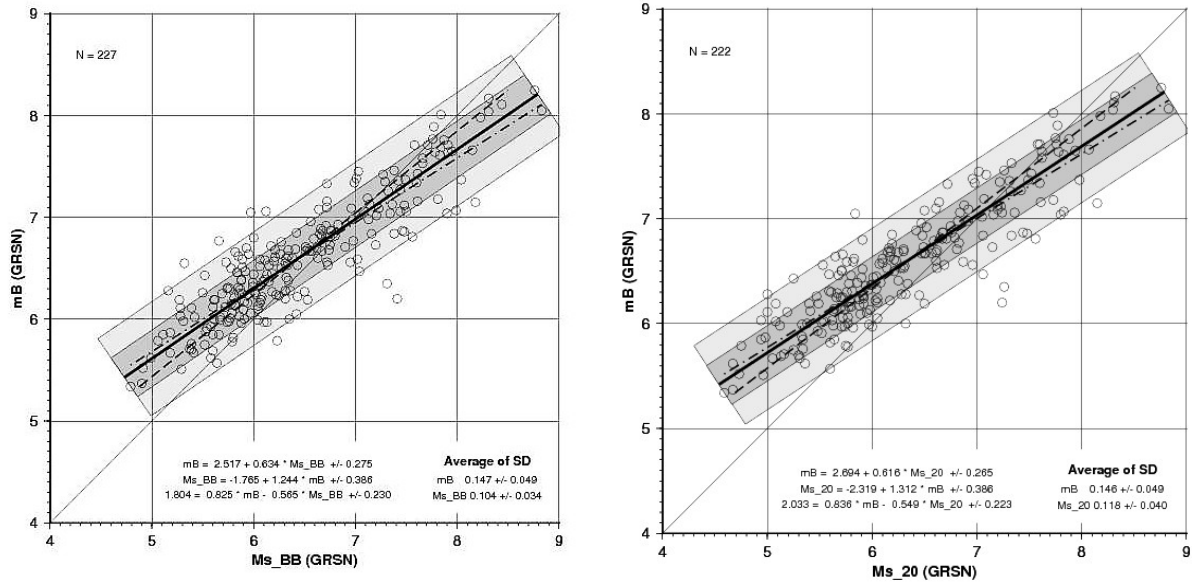
$$mB_{BB} \leftarrow 0.65 Ms_{BB} + 2.30 \quad (\text{with SD} = \pm 0.23). \quad (3.97)$$

Practically identical with (3.94), however, is a new  $mB_{BB} - Ms_{BB}$  SR1 relationship derived from a global set of 227 earthquakes in the range  $4.7 < Ms_{BB} < 8.9$  recorded by STS2 broadband seismographs of the GRSN network in Germany (see also Fig. 3.72 left):

$$mB_{BB} \leftarrow 0.63 Ms_{BB} + 2.52 \quad (\text{with SD} = \pm 0.28). \quad (3.98)$$

According to (3.97) and (3.98)  $mB_{BB} = Ms_{BB}$  for 6.5 and 6.8, respectively, i.e. close to the 6.75 according to (3.95). When converting values between  $Ms_{BB} = 6.0$  and  $8.5$ , i.e. magnitudes for which (3.95) holds, via the relationship (3.97) into  $mB_{BB}$ , then the differences with respect to (3.95) vary between  $-0.08$  and  $-0.03$  m.u. And when using (3.98) instead, the difference to conversions based on the Gutenberg-Richter formula is constant  $+0.02$  m.u. for all magnitudes and thus practically negligible. And with relationship (3.74) Bormann and Wylegalla (1975) had already shown that **even single station BB body-wave and surface-wave magnitudes from global earthquakes** in the magnitude range 4.7 to 8.5 do essentially reproduce (3.95).

Also the SR1 between  $mB_{BB}$  and  $Ms_{20}$ , with slopes between 0.58 (Chinese data; Fig. 3.71 lower left and Tab. 3.4) and 0.62 (GRSN data; Fig. 3.72 right), still match rather well with Eq. (3.95). Conversion differences for  $Ms_{20}$  between 6.0 and 8.5 range between  $-0.04$  and  $+0.13$  m.u., which is much smaller than the standard deviations for all these regressions that vary between  $\pm 0.23$  and  $\pm 0.28$  m.u.



**Fig. 3.72** Regression relationships between  $mB\_BB$  and  $Ms\_BB$  (left diagram), and  $Ms\_20$ , respectively (right diagram) based on the analysis of global earthquakes recorded by the German Regional Seismic Network (GRSN):. SR1 (dot-broken line and upper inserted formula), SR2 (broken line and middle inserted formula) and OSR (bold black line and lower inserted formula). The gray and light-gray shaded bands indicate the 1- and 2-standard deviation limits around the OSR regression line. Given are also the average standard deviations with which the event magnitudes have been estimated.

The differences, however, to the Gutenberg-Richter (1956a)  $mB$ - $Ms$  relationships is slightly larger when comparing a larger set ( $N = 931$ ) of automatically determined global earthquake  $mB(GFZ) = mB\_BB$  measured by the procedure described in Bormann and Saul (2008) and using stations globally distributed, with the respective  $Ms(USGS)$  data, which is in fact an  $Ms\_20$ . According to Fig. 3.73 the SR1 relationship is

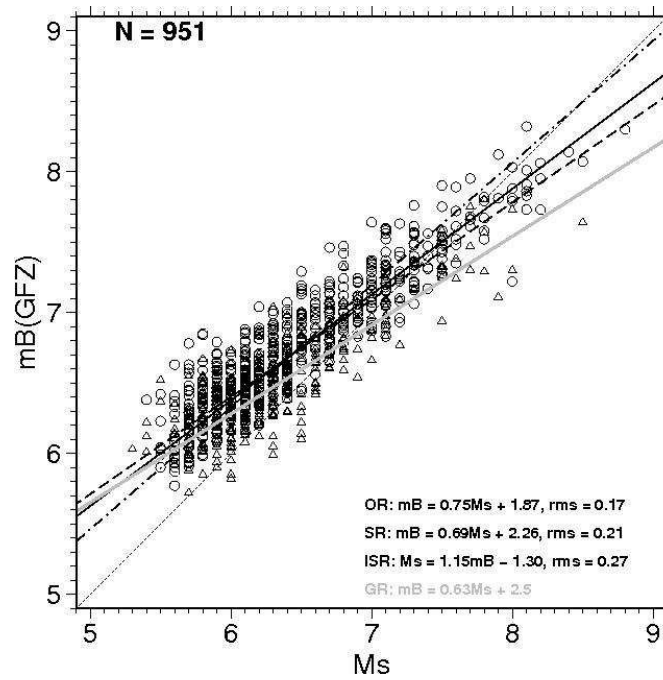
$$mB\_BB = 0.69 Ms + 2.26 \quad (\text{with } SD = \pm 0.28). \quad (3.99)$$

In the magnitude range  $Ms = Ms\_20$  between 5.5 and 8.5, conversions of  $Ms$  into  $mB$  according to (3.99) yield values that are between 0.09 and 0.27 m.u. larger than those derived by using the G-R formula (3.94), but they agree perfectly with the later by Abe and Kanamori (1980) on the basis of more data revised relationship (3.96). The conversion differences between (3.99) and (3.96) vary between  $-0.02$  and  $+0.01$  only. This is insignificant as compared to the measurement errors of these magnitudes.

Yet, the respective orthogonal regressions between  $mB$  and  $Ms$  differ clearly from these classical relationships. For the data in Fig. 3.73 the OSR is:

$$mB\_BB = 0.75 Ms + 1.87 \quad \text{with } RMSO = \pm 0.17 \quad (3.100)$$

and an  $SD = \pm 0.21$  m.u. of the  $mB\_BB$  estimates.

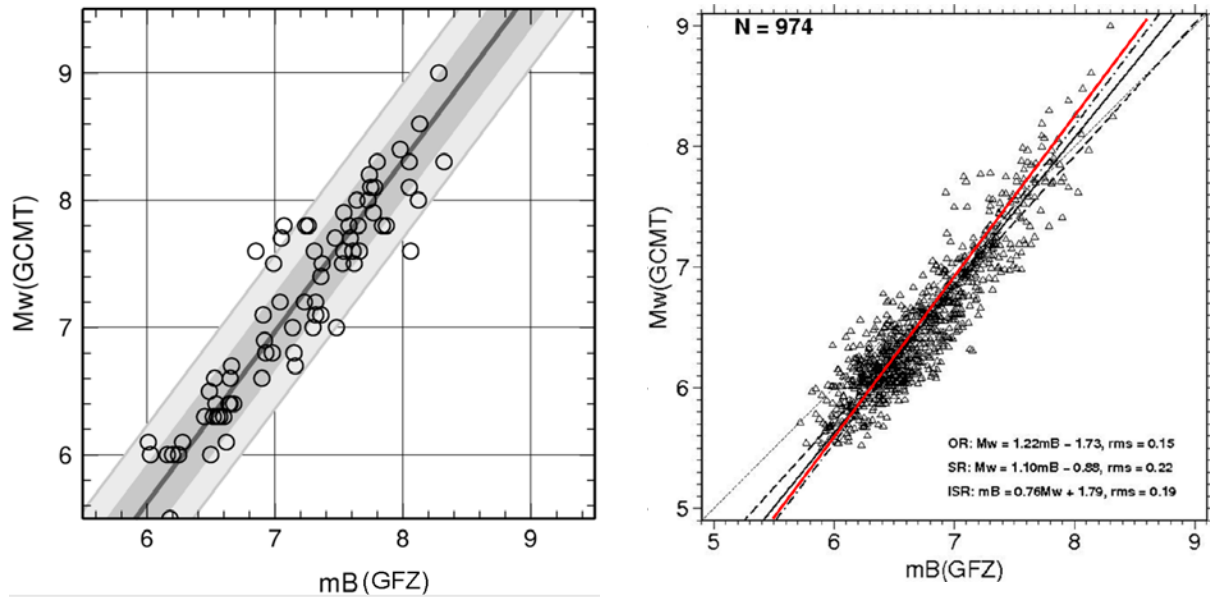


**Fig. 3.73**  $mB(GFZ) = mB\_BB$  over  $M_s(USGS) = M_s\_20$  for a global earthquake-station network data set compiled and processed by J. Saul and Di Giacomo (personal communication, 2010). Different point symbols relate to different types of source mechanism, e.g., the upright open triangles to strike-slip mechanism. The grey full line corresponds to the Gutenberg-Richter SR1 relationship (3.64).

In summary, we rest assured that the new standard magnitudes  $mB\_BB$  and  $M_s\_BB$  do not contradict with the inferences drawn by Richter (1958), and later by Kanamori and others on the basis of the Gutenberg-Richter SR1 relationship (3.94) leading to the derivation of the  $M_w$  standard formula. Modern  $M_s\_20$  and  $mB\_BB$  data, however, do better agree with the Abe and Kanamori (1980)  $mB$ - $M_s$  relationship.

#### 3.2.9.4 Linear orthogonal and standard regressions between IASPEI standard magnitudes and classical $mb$ and $M_s$ with $M_w$ and $M_e$ (P. Bormann, D. Di Giacomo and S. Wendt)

Bormann and Saul (2008) and Bormann et al. (2009), supplemented by data provided by Di Giacomo (2010, personal communication), looked into the OSR and standard regression relationships between IASPEI standard magnitudes with  $M_w$  and  $M_e$  for **global earthquakes recorded by world-wide distributed broadband stations**. Fig. 3.73 plots  $M_w(GCMT)$  over  $mB(GFZ)$  which is an automatically determined  $mB\_BB$  according to the procedure described by Bormann and Saul (2008), in the left panel for a limited early and in the right panel for a greatly extended data set compiled and processed by J. Saul and Di Giacomo (personal communication 2010).



**Fig. 3.74**  $M_w(\text{GCMT})$  over  $mB(\text{GFZ})$  for an initial limited data set (left-hand diagram) and a greatly expanded representative data set compiled and processed by J. Saul and Di Giacomo (personal communication 2010) (right hand diagram). The boundaries of the grey and light gray bands, respectively, in the left hand diagram correspond to one and two standard deviations. Larger positive outliers in  $M_w$  direction in both diagrams are relate to rare very slow tsunami earthquakes and large negative outliers to very energetic ruptures. The read line in the right-hand diagram corresponds to the OSR relation in the left-hand panel according to Bormann and Saul (2008).

The OSR relationship in the left-hand panel of Fig. 3.74 is given by Bormann and Saul (2008) as

$$M_w(\text{GCMT}) = 1.33 mB(\text{GFZ}) - 2.36, \quad (3.101)$$

with  $\text{RMSO} = \pm 0.18$  m.u. and an rms of  $\pm 0.30$  m.u. in the estimated  $M_w$ . The related standard regression SR1, however, is

$$M_w(\text{GCMT}) = 1.21 mB(\text{GFZ}) - 1.45 \text{ with rms} = \pm 0.29. \quad (3.102)$$

and SR2

$$mB(\text{GFZ}) = 0.71 M_w(\text{GCMT}) + 2.08 \text{ with rms} = \pm 0.22. \quad (3.103)$$

We should note, however, that the conversion of available instrumentally measured  $M_w$  data into not available  $mB_{\text{BB}}$  values is not of practical importance at all. Relevant is only the need (or wish) to calculate proxy  $M_w$  estimates from measured  $mB_{\text{BB}}$  or other magnitudes if proper  $M_w$  values are not at all, or - because of longer required data acquisition and processing times - not yet available, e.g., in the context of rapid early warning operations. And then, we have principally the choice to use either the SR1 or OSR regression relationship of  $M_w$  over  $mB_{\text{BB}}$  (or other magnitudes). When comparing long-period  $M_w$  data with other long-period or broadband magnitudes of comparably low initial measurement errors than in any event OSR relations are preferable because they represent more correctly the relationship

between these two magnitudes, although the uncertainty in the estimated dependent variable is of the same order as for the SR1.

In Fig. 3.74 the regression relationships are remarkably good average representations for the vast majority of events despite the very different frequency ranges considered and methodologies applied to measure  $M_w$  and  $mB$ . According to the relationships (3.101) to (3.103)  $mB(GFZ)$  values are on average equal to  $M_w$  between 6.9 and 7.2, larger than  $M_w$  by 0.3 to 0.4 at  $M_w = 6.0$ , and smaller by the same amount at  $M_w = 8.5$ . But **automatically determined  $mB(GFZ)$**  from first arriving P waves is available much earlier than  $M_w(GCMT)$  derived from the analysis of very long-period P, S and surface waves. Therefore, by using relationship (3.101), which compensates best for the systematic trend difference between  $M_w$  and  $mB$ , the latter is used, e.g. in the operational German-Indonesian Tsunami Early Warning System (GITEWS; <http://www.gitews.de>), to get first rough proxy estimates of  $M_w$  within 5 min after origin time.

The respective formulas for the larger data set have been inserted in the right-hand diagram of Fig. 3.74. The OSR relationship reads:

$$M_w(GCMT) = 1.22 mB(GFZ) - 1.73 \text{ with RMSO} = \pm 0.15. \quad (3.104)$$

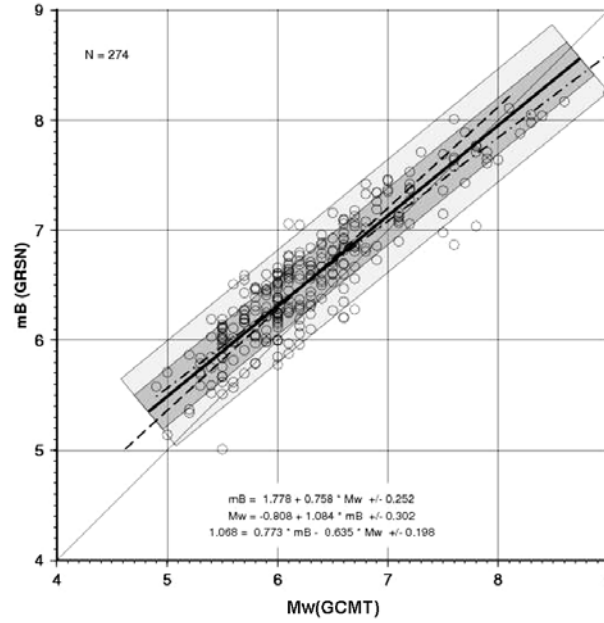
The related uncertainty of  $M_w$  proxy estimates via  $mB(GFZ)$  is  $SD = \pm 0.24$  m.u. and the average difference of  $M_w$  proxies calculated with either (3.101) or (3.104) is generally less than 0.2 m.u.

An almost identical OSR relationship has been derived by S. Wendt by comparing only 274 **manually derived  $mB_{BB}$**  values for global earthquakes recorded at stations of the German GRSN broadband network with  $M_w(GCMT)$ :

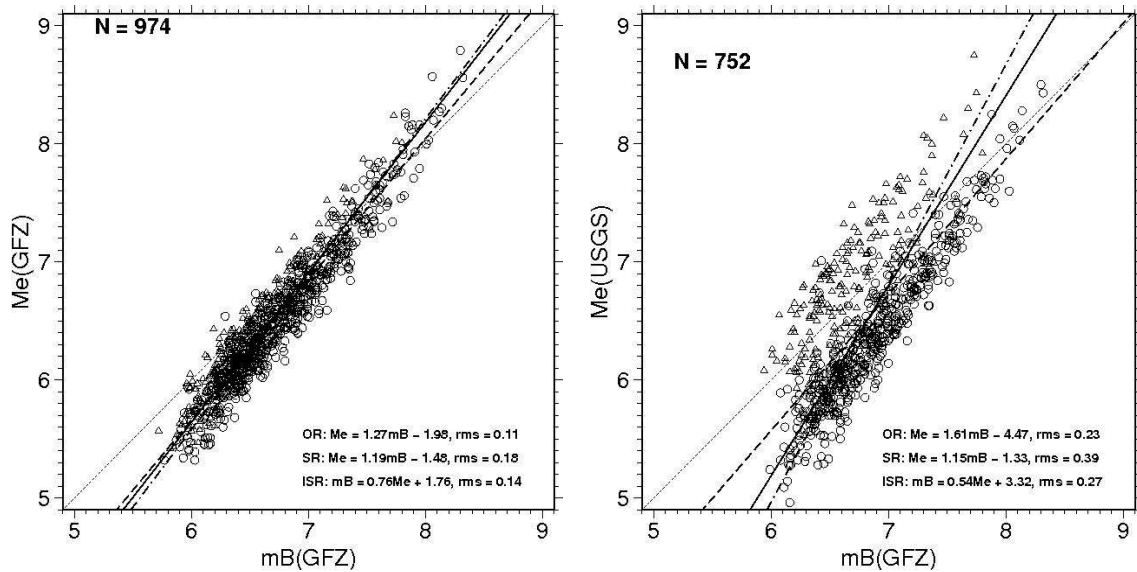
$$M_w(GCMT) = 1.22 mB(GRSN) - 1.68 \text{ with RMSO} = \pm 0.20. \quad (3.105)$$

Fig. 3.75 shows the related data plots together with the OSR, SR1 and ISR relationships and the one and two RMSO error intervals (gray and light gray bands).

Similar relationships have been derived by Di Giacomo (personal communication 2010) between  $mB(GFZ)$ ,  $Me(GFZ)$  and  $Me(USGS)$ . The former  $Me$  data have been calculated with another fully automatic procedure described by Di Giacomo et al. (2008; 2010a and 2010b) while the USGS  $Me$  data are calculated by an off-line procedure described by Choy and Boatwright (1995) and in IS 3.6. In contrast to the GFZ procedure, which does not apply source-mechanism corrections to the measured values, in agreement with other classical magnitude procedures, the USGS procedure applies such corrections. Such theoretically based corrections, however, may have the tendency to overestimate  $Me$ , especially for strike-slip events. Possible reasons have been discussed in section 3.2.7.2. Fig. 3.76 shows plots of  $Me(GFZ)$  and  $Me(USGS)$ , respectively, over  $mB(GFZ)$  with the related OSR, SR1 and  $ISR = SR2$  relationships, which are given as inserts together with their rms values.



**Fig. 3.75** Regression relationships SR1, SR2 = ISR and OSR (see inserted formulas from top to bottom) between  $mB\_BB$ , calculated from broadband records at the German GRSN network, and  $Mw(GCMT)$ . For legend see Fig. 3.72.



**Fig. 3.76** Plots of  $Me(GFZ)$  (left-hand diagram) and  $Me(USGS)$  (right hand diagram) over  $mB(GFZ)$ . Different point symbols relate to different types of source mechanism, e.g., the upright open triangles to strike-slip mechanism.

From Fig. 3.76 it is obvious that the data scatter doubles when accounting for theoretically expected yet purely model-based source-mechanism corrections, primarily for strike-slip (SS) earthquakes. Their share in the left-hand diagram are 323 and in the right-hand diagram 204. Note that in the left-hand diagram the difference between  $mB(GFZ)$ - $Me(GFZ)$  for strike slip events is  $0.13 \pm 0.19$  and for all types of source mechanisms  $0.24 \pm 0.19$ , i.e., an average difference of only 0.11 m.u. In contrast, for event magnitudes plotted in the right-hand diagram the difference between  $mB(GFZ)$ - $Me(USGS)$  for strike slip events is  $-0.15 \pm 0.34$

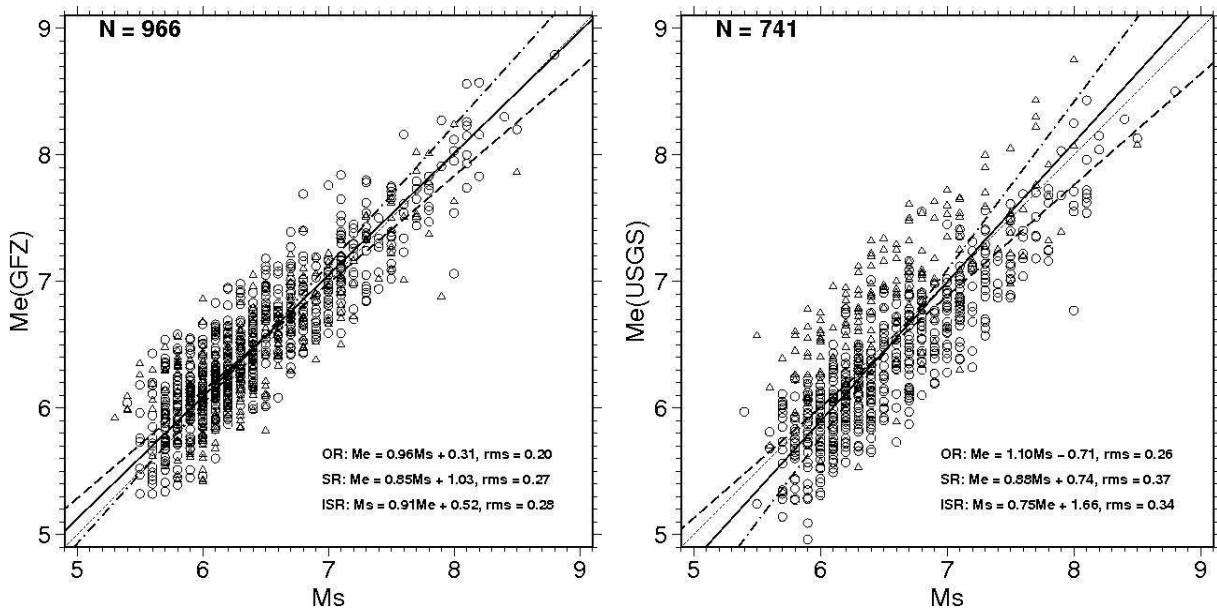


and for all types of source mechanisms  $0.50 \pm 0.24$ , i.e., an average difference of 0.65 m.u. and a greatly increase data scatter.

Proxy estimates of  $Me(GFZ)$ , however, could be made by using the OSR relationship with the automatically determined  $mB(GFZ)$ :

$$Me(GFZ) = 1.27 mB(GFZ) - 1.98 \quad \text{with RMSO} = \pm 0.11. \quad (3.106)$$

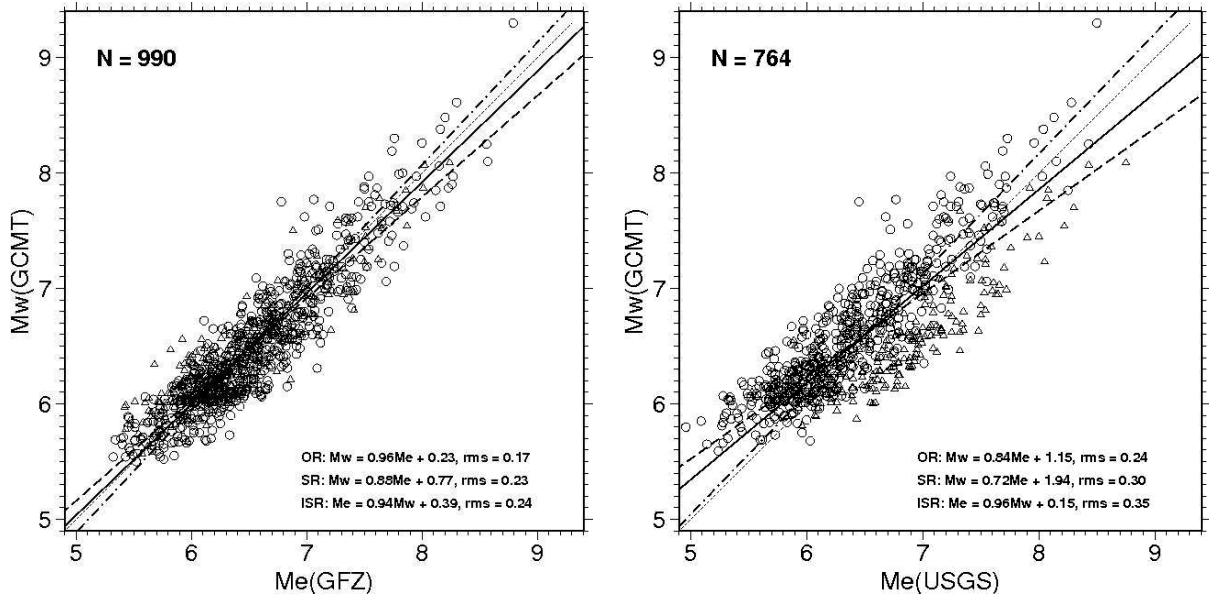
$mB(GFZ)$  is available already within the first few minutes after origin time OT and the standard deviation of the  $Me$  estimate is only  $SD = \pm 0.18$  m.u. This is much better than estimating  $Me(GFZ)$  or  $Me(USGS)$  via OSR conversion relationships with  $USGS Ms = Ms_{20}$ . Such proxy estimates would be uncertain, in terms of SD, by  $\pm 0.28$  m.u. for  $Me(GFZ)$  and, even worse,  $\pm 0.39$  m.u. for  $Me(USGS)$  (see Fig. 3.77), although  $Me(USGS)$  had been scaled to  $Ms(USGS)$  (see Choy and Boatwright, 1995; Choy et al., 2006). The very noisy relationship between  $Me$  and  $Ms$  illustrates, why Gutenberg favoured – intuitively - energy calculations via equation (3.95) based on ground motion velocity related  $(A/T)_{max}$  measurements in a wide period range and not via the Richter  $\log E_S$ - $Ms$  relationship (3.66) solely based on 20 s displacement amplitudes.



**Fig. 3.77** Regression relationships for  $Me(GFZ)$  (left diagram) and  $Me(USGS)$  (right diagram) over USGS  $Ms$ , which is in fact an  $Ms_{20}$ . See inserted formulas for OSR, SR1 and  $ISR = SR2$  with their related rms. Triangles represent strike slip earthquakes, circles all other types of mechanisms.

In view of the large difference between the two established  $Me$  procedures and results the question arises, how these two different approaches of  $Me$  measurement relate to  $Mw$ . Fig. 3.78 presents  $Mw(GCMT)$  over  $Me(GFZ)$  and  $Me(USGS)$ , respectively. Again,  $Me(GFZ)$  correlates better than  $Me(USGS)$  with  $Mw(GCMT)$  at reduced scatter. While at  $Mw = 5.5$   $Me(GFZ)$  values are on average 0.16 m.u. smaller than  $Me(USGS)$ ,  $Me(GFZ)$  is on average 0.16 m.u. larger at  $Mw = 9$ . The OSR relationship  $Mw(GCMT)$  over  $Me(GFZ)$  is on average very close to 1:1, whereas the slope of the respective relationship with  $Me(USGS)$  is clearly less than one. The scatter of  $Me$  with respect to  $Mw$  is mainly due to differences in the  $Mw$ -

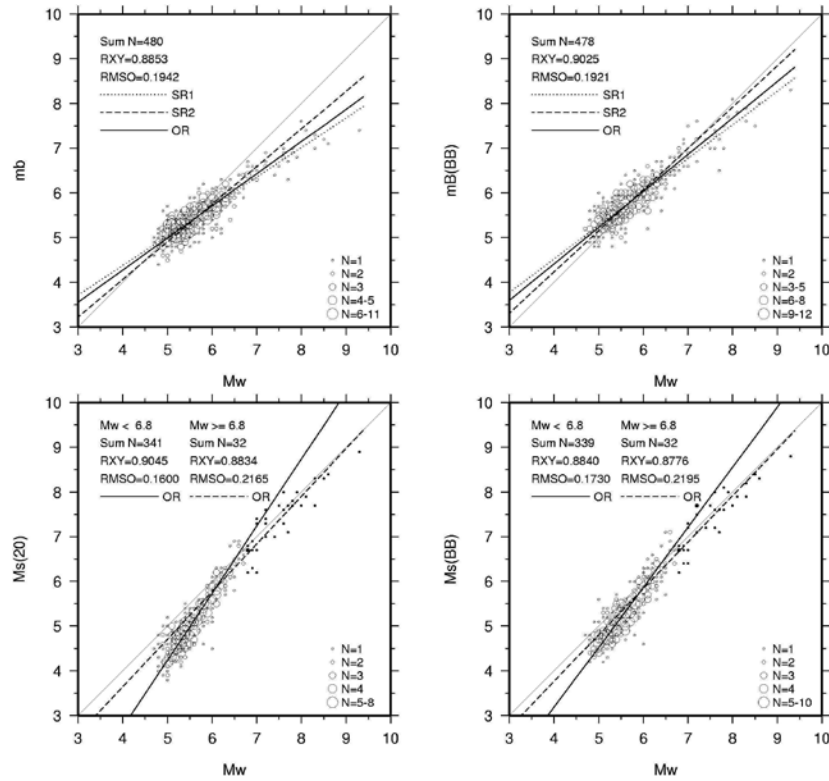
scaled release of radiated seismic energy, which is closely related to differences in seismic stress-drop and/or rupture velocity (see, e.g., Kanamori and Brodsky, 2004; Venkatamaran and Kanamori, 2004b; Bormann and Di Giacomo, 2011). It can not be excluded, however, that the much larger data scatter of  $M_e(\text{USGS})$  is at least partially due to a overcorrection for theoretically expected source-radiation effects (see section 3.2.7.2).



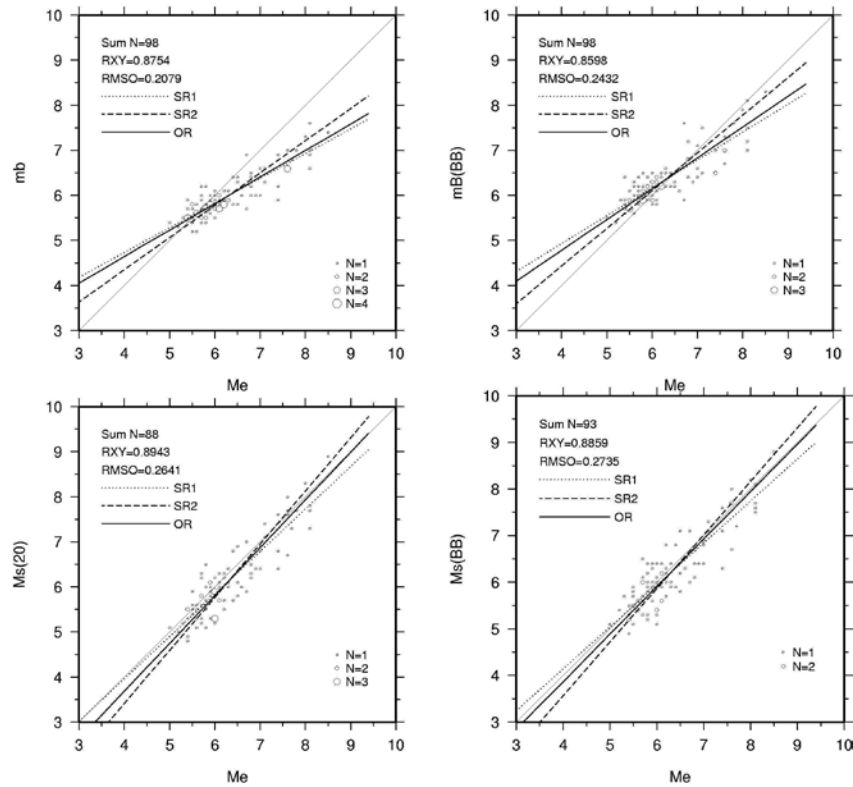
**Fig. 3.78** Regression relationships of  $M_w(\text{GCMT})$  over  $M_e(\text{GFZ})$  (left-hand diagram) and  $M_e(\text{USGS})$  (right-hand diagram), respectively. Triangles represent strike slip earthquakes, circles all other types of mechanisms.

Bormann et al. (2009) published regression relations between  $M_w(\text{GCMT})$  and  $M_e(\text{USGS})$  with all 4 teleseismic IASPEI magnitude standards (see Figs. 3.79 and 3.80 and Tab. 3.4). One should recognize, however, that magnitude conversion relationships aimed at catalog homogenization are generally not used for converting  $M_w$  or  $M_e$  to classical magnitudes but rather to use the latter, which are both more frequently and for much longer time-spans available than the more recent physically based magnitudes, into proxy estimates of  $M_w$  and  $M_e$ . And since we give preference to the use of the OSR relationships we have highlighted in Tab. 3.4 those resolved for  $M_w$  and  $M_e$ , respectively, in bold blue.

We realize that, with the exception of  $M_s_{20}$  conversions for  $M_s < 6.8$  and conversions of  $mB_{BB}$  into  $M_w$  with standard deviations  $SD \approx \pm 0.2$ , proxy estimates of  $M_w$  from other classical magnitudes have generally an  $SD \approx \pm 0.3$  and for conversions of classical magnitudes into  $M_e$  even values around  $SD \approx \pm 0.4$ , with the exception of somewhat lower standard deviations when converting  $M_s$  data.



**Fig. 3.79** SR1, SR2 and OR relationships between mb, mB\_BB, Ms\_20 and Ms\_BB over Mw(GCMT). Legend as in Fig. 3.71. (Copy of Figure 13 in Bormann et al. (2009); © Seismological Society of America).



**Fig. 3.80** SR1, SR2 and OR relationships between mb, mB\_BB, Ms\_20 and Ms\_BB over Me(USGS). Legend as in Fig. 3.71. (Copy of Figure 14 in Bormann et al. (2009); © Seismological Society of America).

**Tab. 3.5** Relationships of magnitudes determined from CNSN data according to the new IASPEI measurement standards with  $M_w$ (GCMT) and  $M_e$ (USGS), respectively. Legend as in Tab. 3.4. (modified Table 3 in Bormann et al. (2009).

| $M_x - M_y$<br>N                            | Regression                  | Relationship                                                                                                                                                                                             | rms<br>Error                                                                                           | Figure                          |
|---------------------------------------------|-----------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------------------|---------------------------------|
| $M_w - mb$<br>480                           | SR1<br>SR2<br><br><b>OR</b> | $mb \leftarrow 0.66M_w + 1.73$<br>$M_w \leftarrow 1.19mb - 0.83$<br><b><math>1.14 = -0.58M_w + 0.81mb</math></b><br>$mb = 0.72M_w + 1.40$<br><b><math>M_w = 1.39mb - 1.94</math></b>                     | $\pm 0.24$<br>$\pm 0.32$<br><b><math>\pm 0.19</math></b><br>$\pm 0.24$<br><b><math>\pm 0.33</math></b> | <b>Fig. 3.76</b><br>upper left  |
| $M_w - mB\_BB$<br>478                       | SR1<br>SR2<br><br><b>OR</b> | $mB\_BB \leftarrow 0.75M_w + 1.52$<br>$M_w \leftarrow 1.08mB\_BB - 0.59$<br><b><math>0.87 = -0.63M_w + 0.77mB\_BB</math></b><br>$mB\_BB = 0.82M_w + 1.15$<br><b><math>M_w = 1.22mB\_BB - 1.40</math></b> | $\pm 0.24$<br>$\pm 0.29$<br><b><math>\pm 0.19</math></b><br>$\pm 0.24$<br><b><math>\pm 0.30</math></b> | upper right                     |
| $M_w - M_s\_20$<br>for $M_w < 6.8$<br>341   | SR1<br><br><b>OR</b>        | $M_s\_20 \leftarrow 1.31M_w - 2.19$<br><b><math>1.81 = 0.83M_w - 0.55M_s\_20</math></b><br>$M_s\_20 = 1.50M_w - 3.27$<br><b><math>M_w = 0.667M_s\_20 + 2.18</math></b>                                   | $\pm 0.28$<br><b><math>\pm 0.16</math></b><br>$\pm 0.29$<br><b><math>\pm 0.19</math></b>               | lower left                      |
| $M_w - M_s\_20$<br>for $M_w \geq 6.8$<br>32 | SR1<br><br><b>OR</b>        | $M_s\_20 \leftarrow 0.93M_w + 0.37$<br><b><math>0.42 = 0.73M_w - 0.68M_s\_20</math></b><br>$M_s\_20 = 1.06M_w - 0.61$<br><b><math>M_w = 0.943 M_s\_20 + 0.575</math></b>                                 | $\pm 0.31$<br><b><math>\pm 0.22</math></b><br>$\pm 0.32$<br><b><math>\pm 0.30</math></b>               | lower left                      |
| $M_w - M_s\_BB$<br>for $M_w < 6.8$<br>339   | SR1<br><br><b>OR</b>        | $M_s\_BB \leftarrow 1.15M_w - 1.10$<br><b><math>1.31 = 0.80M_w - 0.60M_s\_BB</math></b><br>$M_s\_BB = 1.34M_w - 2.19$<br><b><math>M_w = 0.746M_s\_BB + 1.634</math></b>                                  | $\pm 0.28$<br><b><math>\pm 0.17</math></b><br>$\pm 0.28$<br><b><math>\pm 0.21</math></b>               | <b>Fig. 3.76</b><br>lower right |
| $M_w - M_s\_BB$<br>for $M_w \geq 6.8$<br>32 | SR1<br><br><b>OR</b>        | $M_s\_BB \leftarrow 0.91M_s + 0.60$<br><b><math>0.27 = 0.72M_w - 0.69M_s\_BB</math></b><br>$M_s\_BB = 1.04 M_w - 0.39$<br><b><math>M_w = 0.961 M_s\_BB + 0.375</math></b>                                | $\pm 0.31$<br><b><math>\pm 0.22</math></b><br>$\pm 0.32$<br><b><math>\pm 0.30</math></b>               | lower right                     |
| $M_e - mb$<br>98                            | SR1<br>SR2<br><br><b>OR</b> | $mb \leftarrow 0.55M_e + 2.54$<br>$M_e \leftarrow 1.40mb - 2.08$<br><b><math>1.97 = -0.51M_e + 0.86mb</math></b><br>$mb = 0.59M_e + 2.28$<br><b><math>M_e = 1.70mb - 3.86</math></b>                     | $\pm 0.24$<br>$\pm 0.38$<br><b><math>\pm 0.21</math></b><br>$\pm 0.24$<br><b><math>\pm 0.41</math></b> | <b>Fig. 3.77</b><br>upper left  |
| $M_e - mB\_BB$<br>98                        | SR1<br>SR2<br><br><b>OR</b> | $mB\_BB \leftarrow 0.62M_e + 2.46$<br>$M_e \leftarrow 1.20mB\_BB - 1.30$<br><b><math>1.69 = -0.56M_e + 0.83mB\_BB</math></b><br>$mB\_BB = 0.68M_e + 2.05$<br><b><math>M_e = 1.47mB\_BB - 3.42</math></b> | $\pm 0.29$<br>$\pm 0.40$<br><b><math>\pm 0.24</math></b><br>$\pm 0.29$<br><b><math>\pm 0.43</math></b> | upper right                     |

|                             |                         |                                                                                                                                                                                                    |                                                         |             |
|-----------------------------|-------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|---------------------------------------------------------|-------------|
| Me – Ms <sub>20</sub><br>93 | SR1<br>SR2<br><b>OR</b> | Ms <sub>20</sub> ← 0.94Me + 0.20<br>Me ← 0.85Ms <sub>20</sub> + 1.08<br><b>0.38 = 0.73Me - 0.69Ms<sub>20</sub></b><br>Ms <sub>20</sub> = 1.06Me - 0.53<br><b>Me = 0.943Ms<sub>20</sub> + 0.50</b>  | ±0.38<br>±0.36<br><b>±0.26</b><br>±0.3<br><b>±0.33</b>  | lower left  |
| Me – Ms <sub>BB</sub><br>93 | SR1<br>SR2<br><b>OR</b> | Ms <sub>BB</sub> ← 0.90Me + 0.53<br>Me ← 0.87Ms <sub>BB</sub> + 0.89<br><b>0.15 = 0.71Me - 0.70Ms<sub>BB</sub></b><br>Ms <sub>BB</sub> = 1.02Me - 0.21<br><b>Me = 0.980Ms<sub>BB</sub> + 0.196</b> | ±0.38<br>±0.37<br><b>±0.27</b><br>±0.38<br><b>±0.36</b> | lower right |

Other conclusions to be drawn from Figs. 3.79 and 3.80 as well as from Tab. 3.4 and formula (3.105) are:

- According to Fig. 3.79 (lower two diagrams) the relationship between Mw with Ms<sub>BB</sub> and Ms<sub>20</sub>, respectively, is better approximated by a bi-linear OSR.
- While both Ms<sub>BB</sub> and Ms<sub>20</sub> correlate with Mw > 6.5 with a **slope close to 1 (up to about 8.5) the slope is less than 1 for smaller Ms.**
- **At magnitudes < 6.5 Ms<sub>BB</sub> deviates on average less (about half) from Mw than Ms<sub>20</sub>** (for formulas see Tab.3. 4)
- The OSR between Mw and mB<sub>BB</sub> according to the Chinese data in Tab. 3.4 has the same slope of 1.22 as in the formulas (3.104) and (3.105), yields, however, in the whole magnitude range on average Mw proxy estimates that are 0.33, respectively 0..28 m.u. larger than those derived with formula (3.105).
- The orthogonal regression between Ms<sub>BB</sub> and Me in Fig. 3.79 and Tab. 3.4 has a slope near 1:

$$Ms_{BB} = 1.02 Me - 0.21 \text{ with SD} = \pm 0.38 \quad (3.107)$$

and is only somewhat larger (1.06) when plotting Ms<sub>20</sub> over Me.

- This good agreement between Me and Ms is understandable, because the magnitude formula  $Me = (\log E_s - 4.4)/1.5$  has been scaled by Choy and Boatwright (1995) to USGS surface-wave Ms in the wide range between 5.5 and 8.3 , adopting the slope of 1.5 in the Richter (1958) logE<sub>s</sub>-Ms relationship.
- This explains the almost perfect agreement of the SR1 regression mB<sub>BB</sub> - Me in Tab. 3.4:

$$mB_{BB} \leftarrow 0.62 Me + 2.46 \text{ (with SD} = \pm 0.29). \quad (3.108)$$

with the classical Gutenberg-Richter (1956) relationship  $mB = 0.63Ms + 2.5$ , because when deriving the Me formula, Choy and Boatwright (1995) substituted Ms by Me.

These quantitative comparisons between classical and new relationships illustrate that by linking the physically well-defined size parameters M<sub>0</sub> and E<sub>s</sub> with magnitude, both Mw and Me are completely tied to the classical empirical magnitudes of mB and Ms with all their possible biases that are, with the exception of saturation, not yet well known and documented.

The most recent GOR models have been derived by Di Giacomo et al. (2013b) in the framework of the ISC-GEM project for creating a New Reference Global Instrumental Earthquake Catalogue (1900-2009) (see section 3.2.9.5). For the bi-linear  $M_w$ - $M_s$  relationship they read:

$$M_w = 0.67 M_s + 2.13 \quad \text{for } M_s \leq 6.47 \quad (3.109)$$

and

$$M_w = 1.10 M_s - 0.67 \quad \text{for } M_s > 6.47 \quad (3.110)$$

and are comparable with bi-linear standard models obtained from more than 26,000 globally distributed earthquakes by Scordilis (2006):

$$M_w = 0.67(\pm 0.005) M_s + 2.07(\pm 0.03) \text{ with SD} = \pm 0.17 \text{ m.u for } 3.0 \leq M_s \leq 6.1 \quad (3.111)$$

and

$$M_w = 0.99(\pm 0.02) M_s + 0.08(\pm 0.13) \text{ with SD} = \pm 0.20 \text{ m.u for } 6.2 \leq M_s \leq 8.2 \quad (3.112)$$

or the GOR models by Das et al. (2011)

$$M_w = 0.67 M_s + 2.12 \quad \text{for } 3.0 \leq M_s \leq 6.1 \quad (3.113)$$

and

$$M_w = 1.06 M_s - 0.38 \quad \text{for } 6.2 < M_s \leq 8.4. \quad (3.114)$$

All these relationships (3.109) to (3.114), based on some 10.000 to 20.000 data points, are almost identical with the OSR relationships in Tab. 3.4 derived by Bormann et al. (2009) based on only 373 proper  $M_{s\_20}$  event magnitudes determined from Chinese National Network data:

$$M_w = 0.667 M_{s\_20} + 2.18 \quad \text{for } M_s \text{ between } 4.0 \text{ and } 6.5 \quad (3.115)$$

and

$$M_w = 0.943 M_{s\_20} + 0.575 \quad \text{for } M_s > 6.5 \text{ to } 8.8, \quad (3.116)$$

respectively.  $M_w$  proxies calculated with (3.115) differ from those derived by the currently most representative ISC-GEM relationship (3.109) in the magnitude range 4.0 to 6.5 by  $\pm 0.04$  m.u., the respective Das et al. (2011) GOR relationship (3.113) by constant -0.01 m.u. and the Scordilis (2006) relationship (3.111) by constant -0.06 m.u. The differences are larger for magnitudes  $M_s > 6.5$  where all relationships are based on much less data. Yet still  $M_w$  proxy estimates according to Das et al. (2011) agree with the respective ISC-GEM estimates up to  $M_s = 8.5$  within 0.05 m.u, and those by the Bormann et al. (2009) formula within 0.1 m.u.. With -0.18 m.u in  $M_w$  at  $M_s = 8.5$  the Scordilis relationship The largest difference.

Although much inferior as compared to the  $M_w$ - $m_B$ -BB relationships published by Bormann et al. (2009) and above [see formulas (3.101), (3.104) and 3.105) as well as Figs. 3.74, 3.75 and 3.78], in common literature only regression relationship between  $M_w$  and  $m_b$  have been published for converting body-wave magnitudes into  $M_w$  proxies. The reason is that globally operating seismological data centers such as the ISC, NEIC, and GCMT published so far, since the mid 1970s, only shortperiod  $m_b$  body-wave magnitudes.

The most recent  $M_w$ - $m_b$  linear GOR relationship, based on more than 27,000 data points, has been derived by Di Giacomo et al (2013):

$$M_w = 1.38 m_b - 1.79 \quad \text{for } 4.0 < m_b < 7.5 \quad (3.117)$$

It is rather close to the Bormann et al.(2009) OSR relationship, based on just 480 data points:

$$M_w = 1.39 m_b - 1.94 \text{ with SD} = \pm 0.33 \text{ for } 4.5 \leq m_b \leq 7.5 \quad (3.118)$$

The  $M_w$  proxy estimates via (3.117) and 3.118) differ in the range  $4.5 \leq m_b \leq 7.5$  less than 0.1 m.u.

Both relationships, however, differ strongly from the GOR relationship between  $M_w$ (GCMT) and ISC  $m_b$  values published by Wason et al. (2012) with a slope of only 1.13, and even more from the slope of the SR1 relationship by Scordilis (2006), based on more than 40,000 data points:

$$M_w = 0.85 m_b + 1.03 \quad \text{for } 3.5 \leq m_b \leq 6.2 \quad (3.119)$$

These large differences in slope are mainly due to the dataset truncation at  $m_b = 6.2$  by both Scordilis (2006) and Wason et al. (2012).

However, since the standard errors with which event  $M_w$  and  $m_b$  are typically determined ( $\leq 0.1$  m.u. and 0.2 to 0.3 m.u., respectively) are very different, ordinary OSR and GOR, based on the assumption that the ratio of initial errors of the two magnitudes is close to 1, are no longer optimal. Castellaro and Bormann (2007) showed that for standard error ratios  $< 0.7$  the inverse regression relationship, i.e., in our case  $m_b$  over  $M_w$ , is closer to optimal. Bormann et al. (2011) give  $m_b = 0.66 M_w + 1.73$  and Das et al. (2011)  $m_b = 0.65 M_w + 1.65$ . When resolved for  $M_w$  these relationships read for the Bormann et al. (2009) data

$$M_w = 1.52 m_b - 2.62 \quad (3.120)$$

and for the Das et al. (2011) data

$$M_w = 1.54 m_b - 2.54. \quad (3.121)$$

For  $m_b = 7.5$ , which is about the largest value so far measured for standard  $m_b$ , (3.120) and (3.121) would yield  $M_w$  proxy values of 8.78 and 9.01, respectively. These are reasonably large values for such large  $m_b$  values, more likely than those derived via (3.117;  $M_w = 8.56$ ) and (3.118;  $M_w = 8.49$ ). However, proxy  $M_w$  estimate via linear regression relationships with  $m_b$ , which really cannot fit well a highly non-linear data cloud (see Fig. 3.82 in the next section), generally tend to underestimate  $M_w$  for both the strongest and the weakest earthquakes.

Therefore, we do not recommend the conversion of short-period  $m_b$  into long-period  $M_w$ . If it has to be done, for whatever reason, then a non-linear relationship, as presented in the next section, should be used. Yet, in future such  $m_b$ - $M_w$  relationships should be based on new standard  $m_b$  data only. They saturate later and thus reduce the strong non-linear distribution of older  $m_b$  data that have been measured on records with often different responses and within more or less fixed limited time windows after the first P-wave arrival. As one can see from Fig. 3.79, the relation between standard  $m_b$  and  $M_w$  is reasonably linear in a rather large range of magnitudes.

Wason et al. (2012) also published a GOR relationship between  $M_e$  and  $m_b$ :

$$M_e = 1.445 m_b - 2.33. \quad (3.122)$$

It differs strongly from the OSR relationship  $M_e = 1.70m_b - 3.86$  by Bormann et al. (2009), but is very close to the related standard regression (see Tab. 3.4):

$$M_e \leftarrow 1.40 m_b - 2.08 \quad \text{with SD} = \pm 0.38 \text{ m.u.} \quad (3.123)$$

In the range of applicability of (3.122), i.e., for input values of  $5.0 < m_b < 6.5$ , the  $M_e$  proxy estimates differ less than 0.04 m.u. from those derived via (3.123). The latter relationship, however, is applicable up to  $m_b = 7.5$ .

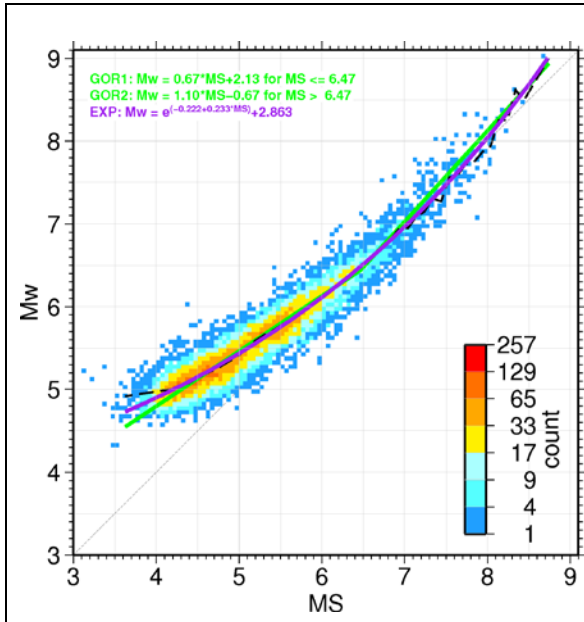
### 3.2.9.5 Non-linear ISC-GEM relationships between $M_s$ and $m_b$ with $M_w$ (D. DiGiacomo and P. Bormann)

The ISC-GEM project aimed at the establishment of a New Reference Global Instrumental Earthquake Catalogue (1900-2009) (see Storchak et al., 2013; Di Giacomo et al., 2013b) made every effort to use uniform procedures of magnitude determination during the entire period of the catalogue. Both surface-wave magnitudes  $M_s$  and short-period body-wave  $m_b$  were recalculated, benefitting from new hypocenters (Bondár et al., 2013), previously unavailable amplitude-period data digitized during this project (Di Giacomo et al., 2013a) and a more reliable algorithm for magnitude computation based on 20% alpha-trimmed median network magnitude from several stations (Bondár and Storchak, 2011). For  $M_s$  up to 1970 several thousands of station magnitudes not available before were processed.

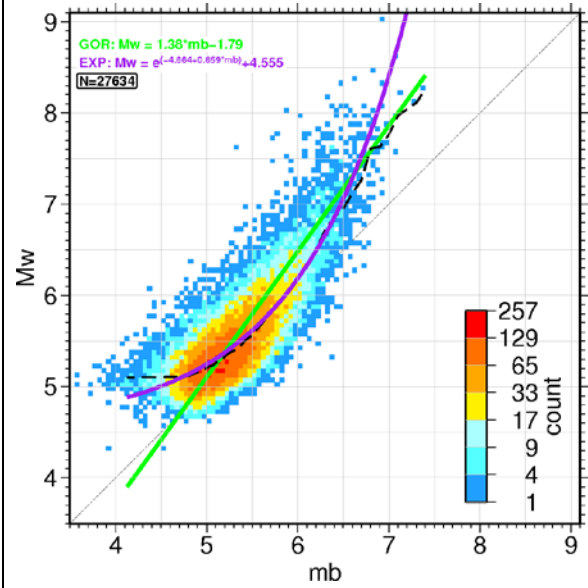
The re-computed  $M_s$  and  $m_b$  values provided an ideal basis for deriving new conversion relationships to moment magnitude  $M_w$ . Therefore, rather than using already published regression models, new empirical relationships were derived using both Generalized Orthogonal Linear and exponential non-linear least-square models to obtain  $M_w$  proxies from  $M_s$  and  $m_b$ . The new models were tested against true values of  $M_w$  and found to be preferable to the linear GOR regressions in computing  $M_w$  proxies for the new ISC-GEM catalog. For more details see Di Giacomo et al. (2013b).

The linear GOR relationships derived by Di Giacomo et al. (2013b) for converting recalculated ISC  $M_s$  and  $m_b$  values into  $M_w$  proxies have been given already in the previous section as formulas (3.109), (3.110) and (3.117), respectively. Below they are plotted together with the non-linear OLS regressions and the median values polygons through the  $M_s$ - $M_w$  and  $m_b$ - $M_w$  data clouds (Figs. 3.81 and 3.82).





**Fig. 3.81** Training dataset set  $M_s$ - $M_w$  with regression models: the exponential model is drawn in purple; the bi-linear GOR model is shown in green. The dashed black polygon curve represents the median values for separated bins. Note the good fit between the exponential model and the median values. (Copy of Fig. 14 of Di Giacomo et al., 2013b, © PEPI )



**Fig. 3.82** Training dataset set  $m_b$ - $M_w$  with regression models: Purple - the exponential model; green - the linear GOR model. The dashed black polygon curve connects the median values for separated bins. Note the good fit between the exponential model and the median values between  $m_b=4.7$  and  $6.5$ . (Copy of Fig. 17 of Di Giacomo et al., 2013b, © PEPI )

Although it is obvious from Fig. 3.81 that fitting the  $M_s$ - $M_w$  data cloud with a bi-linear orthogonal regression is already good enough for obtaining sufficiently reliable  $M_w$  proxy estimates within realistic error limits, such an approach introduces some arbitrariness in the data set separation and also a discontinuity point in the relationships derived. The transition between slope  $\sim 0.7$  and  $\sim 1$  is not sharp at all and the separation, which had been adopted by Scordilis (2006) and Das et al. (2011) at  $M_s = 6.1$ , by Bormann et al. (2009) at  $6.5$  and by Ekström and Dziewonski (1988) even at  $6.8$ , could in fact be moved anywhere in this range. Thus, data pairs in this  $M_s$  range may belong to one or another domain depending on the subjective choice of an author of how the data set was divided into the two domains. This also raises the question on how to consistently map the uncertainty in  $M_s$  to  $M_w$  proxies around the separation of the two linear trends. To approximate sufficiently well the highly non-linear trend in the  $m_b$ - $M_w$  data cloud of Fig. 3.82 one would even need at least three or even four linear GOR regression lines.

In order to avoid this problems Di Giacomo et al. (2013b) fitted a single, continuous regression curve to the two training dataset of data points, using an exponential model of the form  $M_y = e^{(a+b \cdot M_x)} + c$ . As one can see from Figs. 3.81 and 3.82, these exponential curves fit rather well with the median values for separated bins. The regression is performed using a

non-linear ordinary least square algorithm (Bates and Watts, 1988; Bates and Chambers, 1992), which is freely available with the R-language (<http://stat.ethz.ch/R-manual/R-patched/library/stats/html/nls.html> or <http://www.r-project.org/>).

Admittedly, this algorithm does not yield an optimal “orthogonal” fit which accounts for the difference in the (usually not well known) measurement errors of the two variable. Rather, the independent, given and to be converted variable is assumed to be error-free. This, however, means that its errors are projected into the dependent, to be estimated variable, thus accounting for uncertainty of the Mw proxy estimates. Moreover, we have be aware that in our case Mw(GCMT) is, in contrast to mb, already incomplete for values below 5.5. But for the purpose of avoiding censoring effects in the model for a magnitude range where Mw proxy estimates above 5.0 were still needed by the GEM project, Mw values below 5.5 had been added to the data set. We are fully aware, however, that one cannot have a reliable OLS mb-Mw model also for smaller quakes when regressing only Mw(GCMT) values over mb. Rather, as discussed already in section 3.2.9.1, mb tends to scale with Mw at values <4.5-5.0 nearly 1:1. Therefore, one should not use the following relations for converting mb and Ms values below 5.0, respectively about 4.5, to Mw.

Despite these limitations, an important advantage of the models presented below is that histogram equalization has been applied to obtain an equal number of data in magnitude bins of varying size in order to preserve the shape of the distribution in the training and validation sets. The latter were used to compare true Mw values and proxy estimates. Thus it could be proved that on average the agreement between Mw and its proxy estimates Mw(Ms) is in the magnitude range 5 to 8 practically 1:1 with rather low scatter, however bad and noisy for Mw(mb) proxies (compare Figures 15 and 18 in Di Giacomo et al., 2013). This was to be expected from the data plots in Figs. 3.81 and 3.82. Therefore, for the Global Instrumental Earthquake Catalogue (1900-2009), only for a small fraction of earthquakes without Ms measurements, usually for deep earthquakes, Mw(mb) has been calculated from mb values only between 5 and 6. The exponential model to convert Ms to proxy Mw reads as

$$M_w = e^{(-0.222+0.233 \times M_s)} + 2.863 \quad (3.124)$$

and that for converting mb into proxy Mw as

$$M_w = e^{(-4.664+0.859 \times m_b)} + 4.555. \quad (3.125)$$

After all the pros and cons of deriving and using conversion relationship as discussed above in section 3.2.9.4 and by Di Giacomo et al. (2013b), including also verification tests, it is strongly recommended to use in future for converting Ms\_20 type of surface waves from the global data centers only formula (3.124). However, one should refrain from converting mb into Mw proxies, if Ms event data are available. If not, the inferior mb derived Mw proxies should be used and interpreted only with caution. In any event, one should never mix and use with equal weight directly measured Mw and proxy values.

### 3.2.9.6 Relationships for converting local and regional magnitudes into Mw (P. Bormann)

So far we have looked only into relationships between the most common types of instrumentally measured magnitudes with emphasis on the recently approved new IASPEI standard magnitudes. They are mostly teleseismic ones. However, the great majority of events

in national-regional catalogs are local or regional events associated with primarily local (Ml) or duration (Md) magnitude types, often not yet according to standardized rules. Because of this and the regionally highly varying attenuation laws, Ml formulas, even when calibrated to the ML standard, may differ significantly (see DS 3.1). The same applies to Md formulas (3.2.4.5). Yet, also such data are nowadays converted into “unified” moment magnitudes Mw, based on regression relationships not with Mw(GCMT), which is routinely available only for  $M_w \geq 5.5$ , but to regional Mw estimates, down to much weaker events, via spectral parameter analysis (see EX 3.3) instead of fitting long-period waveforms.

A very detailed report about the derivation and use of local-regional event conversion relationships based on both local and teleseismic magnitudes as well as regional and GCMT Mw has been presented by Braunmiller et al. (2005) by way of example for Switzerland. These authors also give average uncertainties for the Mw proxy estimates via such relations. They range between  $\pm 0.2$  for Ml and  $\pm 1.0$  m.u. for the macroseismic magnitude estimates Mms of the Swiss Earthquake Service (SED). This is clearly worse than for any proper direct Mw measure. Yet, Braunmiller et al. derived also average Mw conversion relationships for Ml and Md data published by agencies in neighboring countries, which are relevant for the seismic hazard assessment on the Swiss territory too. This concerns data of the Landeserdbebendienst Baden-Württemberg (LED) and of the University of Karlsruhe (KHE) in Germany, the Laboratoire de Detection Geophysique (LDG) in France, and the Istituto Nazionale di Geofisica e Vulcanologia in Italy (INGV). Interestingly, the local magnitudes of all these institutions scale with a slope of 1 with Mw, however their constants as well as the average uncertainty of the Mw proxy estimates vary strongly, hinting to not standardized level scaling, different degree of reliability of the input Ml (see relationships 3.126), Md (see relationships 3.127), and Mm (relationship 3.128) and thus the incompatibility of the respective agency magnitudes and therefrom calculated Mw proxy estimates:

$$\begin{aligned} M_w &= M_l(\text{SED}) - 0.2 \pm 0.2 = M_l(\text{KHE}) - 0.2 \pm 0.4 = M_l(\text{LED}) - 0.3 \pm 0.4 \\ &= M_l(\text{LDG}) - 0.6 \pm 0.4 = M_l(\text{INGV}) - 0.3 \pm 0.7, \end{aligned} \quad (3.126)$$

$$M_w = M_d(\text{LDG}) - 0.8 \pm 0.5 = M_d(\text{INGV}) + 0.1 \pm 0.5, \quad (3.127)$$

$$M_w = M_{ms}(\text{SED}) \pm 0.5 - 1.0. \quad (3.128)$$

Yet, often published relationships aim not at calculating Mw proxies but at estimating via Ml  $\log M_0$ . The latter we converted with the Mw standard formula (3.68) into Mw-Ml relationships and compiled them together with other regional Mw-Ml relations in Tab. 3.6.

In summary, one realizes that in contrast to most relationships between teleseismic magnitudes with Mw there are no globally representative relationship for converting local magnitudes into Mw. Not only the slopes may differ strongly, between about 0.6 and 1.0, but even when the slopes are comparable, then the constants may differ strongly, up to half a magnitude unit or even more. Striking examples in Tab. 3.6 are the difference between the conversion formulas in Western Canada for events in the continental crust and those in the subducting slab (Ristau et al., 2005), or between NW Turkey and Italy. This highlights the need to base any such local-regional conversion formulas on own careful investigations in the respective area. There is no chance to find a priori well-fitting ones in publications related to other seismotectonic regions, although one might later find out a very close agreement with a formula for another region, as, e.g. Margaritis and Papazachos (1999) for their  $\log M_0$  -  $M_{lsm}$  (SM = strong motion) formula for Greece with that by Hanks and Kanamori (1979) (see Tab. 3.6).

**Tab. 3.6** Linear regression formulas for converting local and regional  $M_l$  into  $M_w$ .  $M_{lSM}$  = strong motion  $M_l$

| $M_w =$                  | $M_l$ range     | Region              | Reference                    |
|--------------------------|-----------------|---------------------|------------------------------|
| $0.61 M_l + 0.85$        | $0 - < 4.5$     | NW Italy            | Bindi et al. (2005)          |
| $0.79 M_l + 1.20$        | $3.5 - 5.8$     | Italy               | Castello et al. (2007)       |
| $0.78 M_l + 0.68$        | $0.5 - 6.0$     | NW Turkey           | Parolai et al. (2007)        |
| $0.80 M_l + 0.93$        | $1.0 - 6.0$     | California          | Chávez & Priestley (1985)    |
| $0.87 M_l + 0.60$        | $2.2 - 5.3$     | Israel              | Shapira & Hofstetter (1993)  |
| $1.00 M_{lSM} - 0.02$    | $3.9 - 6.6$     | Greece              | Margaris & Papazachos (1999) |
| $1.00 M_l - 0.05$        | $\approx 3 - 7$ | Southern California | Thatcher & Hanks (1973)      |
| $1.00 M_l - 0.16$        | $2.7 - 5.9$     | Switzerland         | Braunmiller et al. (2005)    |
| $1.00 M_{lcrust} + 0.06$ | $3.6 - < 5.5$   | Western Canada      | Ristau et al. (2005)         |
| $1.00 M_{lslab} + 0.58$  | $3.6 - 6.0$     | Western Canada      | Ristau et al. (2005)         |

Using general orthogonal regressions (GORs), Gasperini et al. (2013a) calibrated local magnitudes, estimated in Italy by using various methods in different periods of time from 1981 to 2010, with a set of homogeneous moment magnitudes  $M_w$ . Magnitude uncertainties, necessary for the application of GOR methods, were inferred by a trial-and-error procedure based on *a priori* information and empirical regression results. They found that these updated and uniquely processed Italian  $M_l$ - $M_w$  relationships also scale 1:1 in most cases but that their  $M_l$  values underestimate in general  $M_w$  by about 0.1- 0.2 magnitude units.

Besides dominantly linear relationships between  $M_w$  and  $M_l$  one finds occasionally also non-linear ones, e.g., those published by Gusev (1991) and Gusev and Melnikova (1992) (see also Figure 9 in IS 3.7). Another logarithmic one has been published by Wu et al. (2001) for Taiwan:

$$M_l = 4.53 \times \ln(M_w) - 2.09 \pm 0.14, \quad (3.129)$$

and a non-orthogonal non-linear one with a quadratic term included by Parolei et al. (2007) for Northwestern Turkey:

$$M_w \leftarrow (0.95 \pm 0.03) + (0.58 \pm 0.02)M_l + (0.003 \pm 0.004)M_l^2. \quad (3.130)$$

### 3.2.10 Summary remarks about magnitudes and their perspective (P. Bormann)

According to Richter (1935), magnitude was intended to be a measure of earthquake size in terms of the seismic energy  $E_s$  released by the source. Yet, more than 40 years later, Kanamori (1977), in his effort to develop a non-saturating magnitude scale, proposed to relate magnitude to the scalar seismic moment  $M_0$  and thus to the amount of “work” performed by rupturing and displacing the earthquake fault. For estimating energy via  $M_0$  assumed an average stress drop and ratio between  $E_s/M_0$ .

$E_s$ , being proportional to the squared velocity of ground motion, can theoretically be obtained by integrating spectral energy density over all frequencies contained in the transient

waveform, e.g., of the P-, S-, or surface-wave train. This procedure could not be carried out efficiently with analog recordings. Therefore, Gutenberg (1945b and c) assumed that the maximum amplitude observed in a body-wave group was a good measure of the energy density in that arrival. As classical seismographs were relatively broadband displacement sensors, he estimated ground motion velocity by dividing the maximum ground displacement of body-wave phases by their associated periods in the range between some 2 to 20 s for calculating a medium-period broadband magnitude  $m_B$  [see Eq. (3.43)]. But the largest displacement plateau amplitude occurs at somewhat longer periods than the maximum velocity amplitudes (see Fig. 3.5). None the less, according to the published formula, Gutenberg's intention was to measure  $(A/T)_{\max}$  and not  $(A_{\max}/T)$ . For estimating the magnitude  $M_s$  from surface waves, however, Gutenberg (1945a) proposed measuring just the maximum displacement amplitude in the surface-wave train around 20 s, although one finds in pre-1970s station bulletins plenty of surface-wave  $A_{\max}$  data published at periods between some 5 and 60 s (see Fig. 3.35b).. Also Gutenberg used in fact occasionally for his  $M_s$  calculations amplitude readings from bulletins made at periods as low as 12 s and as high as 23 s (Abe, 1981a; Lienkaemper, 1984).

According to Fig. 3.5, the spectral amplitudes sampled around 20 s are reasonably good estimators of the maximum of the source velocity spectrum and thus of  $E_s$  only for magnitudes between about 6 and 8.5 as well as of the maximum displacement amplitude plateau and thus of seismic moment for magnitudes  $< 7.5$ . Richter's  $M_L$  also measures only the displacement amplitude maximum in local earthquake records, typically at frequencies  $< 1$  s. Yet, according to Fig. 3.5, such amplitudes are on average both a reasonably good seismic moment estimator for magnitudes  $< 5$  (see several  $M_w$ - $M_L$  relationships in section 3.2.9.6) and a good estimator of the maximum of high-frequency energy release for magnitudes  $< 6$ .

Further, one should note that all classical calibration functions for  $M_L$ ,  $m_B$  and  $M_s$ , including also the IASPEI standard  $M_s$  formula since 1967 [see Eq. (3.36)], which accepts proper  $(A/T)_{\max}$  measurements in a wider range of periods (3 to 60 s) and distances ( $2^\circ - 160^\circ$ ), do account only for distance- (and for body waves additionally depth-) dependent attenuation but neither for any source-mechanism correction nor for a frequency-dependence of the quality factor  $Q$ . The latter seems to be justified at least for periods  $\geq 4$  s, for which the frequency dependence of  $Q$ , if any, becomes negligible or not separable from other disturbing influences on the measured amplitudes (see Chapter 2, section 2.5.4.2).

Since the pioneering works of Richter (1935), Gutenberg (1945 a, b, and c) as well as Gutenberg and Richter (1956 a, b, and c) magnitudes have become, besides phase identification, onset time readings and source locations derived therefrom, the most important earthquake parameter data published in station bulletins and national as well as international earthquake catalogs. Admittedly, seismic moment tensor solutions, scalar seismic moment  $M_0$ , radiated seismic energy  $E_s$  and strong-motion maps may be of greater scientific interest, giving a more precise, complete and physically based description of the investigated seismic source processes and their static, kinematic and dynamic effects. Yet magnitude data will remain also in future an indispensable, most frequently asked for and practically used earthquake source parameter, for both long-term statistical assessment of seismicity and seismic hazard, as well as for public announcements on earthquakes and tsunamis and for providing a first rapid event assessment to guide disaster management and relief operations. Neither the public, nor decision makers and disaster managers can comprehend and base meaningful actions on physical parameters given to them in the range of many orders of magnitude (in different physical units and with a precision of several decimals). What is relevant in the earthquake disaster context for practical actions are simple integer numbers,

such as intensity degrees related to felt earthquake shaking strength and related damages, ranging from 1 to 7 degrees in Japan or 1 to 12 degrees in most other countries (see Chapter 12), or ground accelerations in decimal fractions of the earth gravity  $g$ , or just magnitudes, always being less than 10 and given with a precision not better than one decimal. Such scales are comprehensible and relate to each other sufficiently well. A higher precision is not required, neither for warning messages nor for decisions related to response, disaster preparedness and mitigation efforts.

Yet, besides still many unsolved scientific issues, e.g., those relating to the improvement and harmonizing of the calibration functions applied, there are three major issues that complicate the proper use and assessment of magnitude data in practice:

- a) The diversity and inhomogeneity of procedures;
- b) Changes of procedures, made by agencies or stations, which have often not been well documented and/or analyzed for their effects, thus disrupting the necessary long-term stability and compatibility of magnitude data that have been published with identical symbols/names;
- c) The current trend to convert all types of magnitudes in catalogues just into  $M_w$ . Often it is not said from which type of magnitudes such proxy estimates of  $M_w$  have been derived and what the different ranges of uncertainties for such proxy estimates are.

#### **Examples for a)**

With the establishment by the United States of the WWSSN in the 1960s and 70s no more medium-period broadband systems, required for the determination of  $m_B$ , had been available in this global network of seismic stations. This global network aimed mainly at lowering the detection threshold for seismic events globally down to body-wave magnitudes of about magnitude 4 (instead of 5 to 5.5 achievable with  $m_B$  for teleseismic events). A magnitude 4 corresponds approximately to that of a well contained underground nuclear explosion (UNE) of about 1 kt TNT equivalent. Yield estimates via such magnitudes as well as best possible discrimination between explosions and earthquakes were other original priority tasks for the WWSSN. These goals were hoped to be best achieved by two band-limited seismograph responses with maximum magnification at periods between about 0.5 to 1.0 s (termed WWSSN-SP) and between about 10 and 20 s (termed WWSSN-LP). For the body waves this necessitated the introduction of another, short-period, scale, termed  $m_b$ , although the measured amplitudes were continued to be calibrated with the same Gutenberg-Richter (1956a) calibration function  $Q(\Delta, h)PV$  for medium- to long-period vertical-component P-wave amplitude readings. Additionally, in the interest of best possible discrimination between UNE or other explosions from earthquakes, it was regulated to measure  $(A/T)_{\max}$  not within the whole P-wave train, as Gutenberg did, but within the first few seconds after the P arrival. Useful, as these modifications in record response and measurement-time window (later again expanded) have been for the purpose of UNE detection and their discrimination from earthquakes on the one hand and the lowering of the global event detection threshold on the other hand, they resulted in the much earlier saturation of  $m_b$ , as compared to  $m_B$ , and thus the underestimation of the size of strong earthquakes if only short-period P-waves were analyzed. And the USGS/NEIC  $M_{s\_20}$  values, based on WWSSN-LP records and the use of the IASPEI standard  $M_s$  formula, show a distinct distance dependence and are not fully compatible with broadband  $M_s$  measured in a wider period and distance range using the same calibration function.

Similarly, other requirement or concepts for classifying the size of seismic events resulted in the development of even more magnitude scales such as the Russian K-class, the duration magnitude  $M_d$  or  $mb_{Lg}$  for near and regional events, or of  $m(P, 1\text{Hz})$ ,  $m(PKP)$ , the mantle magnitude  $M_m$ , the more physically based moment magnitude  $M_w$  and energy magnitude  $M_e$ , the fast  $M_w$  proxy estimators  $M_{wp}$ ,  $M_{wpd}$  and others for teleseismic events, or the assignment of macroseismic magnitude values  $M_{ms}$  to historical earthquakes. The reasons for their development and the specific procedures of their measurement have been outlined in detail in the sections 3.2.4 to 3.2.8. These scales have different precision and are more or less reliable mutually scaled (see, e.g., 3.2.9.2 and 3.2.9.6). This notwithstanding, magnitudes of different kinds will still persist and be required in the foreseeable future to allow the size classifications of seismic events under very different instrumental and environmental conditions and envisaged applications, as, e.g., of  $M_l$  that is measured at periods of particular relevance ( $T \approx 0.1\text{ s}$  to  $3\text{ s}$ ) for earthquake engineers, engineering seismologists and disaster managers. Their proper use, however, requires an understanding of their potentials, limitations, original definitions and mutual relationships. Some of these scales are really important complements, such as the narrow-band magnitudes  $mb$  and  $Ms_{20}$ . Their difference permits to draw essentially the same inferences on kinematic-dynamic differences between seismic events as they can now be made, even better, by comparing  $M_e$  and  $M_w$  (see section 3.2.7.2 and IS 3.5). However, for the ordinary user of magnitude data all these details and the reasons for differences and discrepancies between the different types of magnitude can just not be overlooked and reliably be assessed from the vast literature. The previous sections hopefully can help to alleviate this problem.

### **Examples for b)**

Even more confusing and discouraging are inconsistent and even temporal changes of procedures for magnitude data of the same kind, published with the same magnitude symbol. As outlined in section 3.2.4.1, it was found out only decades later that the  $M_L$  data in California have not been measured consistently. Not properly documented and investigated changes in analog and digital amplitude reading practices resulted in baseline shifts that faked temporal changes in seismicity rates which were misinterpreted as being related to the  $M_w 7.5$  Kern County earthquake of 1952. This led Habermann (1995) to state that: “Such mistakes have the potential to undermine the relationship between the seismological community and the public we serve.”

Less dramatic but still significant have been the changes made at the USGS/NEIC with respect to early years  $mb$  (see example a) when following IASPEI recommendations of 1976 to increase the measurement time window, for great earthquakes up to about one minute, and later by replacing the original WWSSN-SP response with the PDE response which has a somewhat broader relative bandwidth and steeper roll-off towards lower frequencies (see Figure 3 in IS 3.3). And the IDC of the CTBTO uses still another pre-processing filter before measuring its  $mb$ . It has an even higher frequency of peak amplification and still steeper flanks towards lower and higher frequencies. In the extreme case of the great  $M_w 9.3$  Sumatra Andaman earthquake of 2004 these different agencies reported  $mb$  values which differed up to 1.5 m.u.:  $mb(IDC) = 5.7$ ;  $mb(CENC) = 6.4$ ;  $mb(PDE) = 7.2$ .

Such discrepancies between magnitudes of the same type are alarming and unacceptable. The recently adopted IASPEI (2005, 2011, and 2013) standards for the most widely used common types of magnitudes:  $mb$ ,  $mb_{Lg}$ ,  $mb_{BB}$ ,  $M_L$ ,  $Ms_{20}$  and  $Ms_{BB}$  as well as for the calculation of  $M_w$  (see sections 3.2.3.2; 3.2.4.2; 3.2.5.1; 3.2.5.2 as well as IS 3.3) aim at eliminating or at least significantly reducing procedural differences between magnitudes of the same kind.

Magnitudes of the same type as these standard magnitudes but based on different measurement procedures which yield values that differ on average more than 0.1 m.u. from these standard magnitudes have to be published in future with a specifying nomenclature (see IS 3.2). Moreover, seismological agencies publishing magnitude data are requested to give full evidence of their procedures of magnitude determination by filling in the questionnaire in Annex 2 of IS 3.4 and by publishing this information on their related website and/or in their bulletins and deposit it at the main international seismological data centers, such as the ISC and the NEIC. With these measures it is expected to drastically reduce in the foreseeable future heterogeneity and the scatter of magnitude data and to assure improved long-term continuity and stability of magnitude data, thus increasing their value for application and research.

### **Examples for c) and how to alleviate the problem**

One of the most important application of magnitude data is in seismic hazard assessment. One of the essential steps in seismicity and hazard assessment procedures is to estimate the average annual return period of earthquakes of a given magnitude. However, such magnitude-frequency plots would be completely “blurred” if one would take into account and mix all different kinds of magnitudes with their often distinct differences in various magnitude ranges, unless there is a single type of magnitude available which represents best the event “size”. Gutenberg and Richter (1954) have been the first to assess the seismicity of the Earth by means of such magnitude-frequency plots using their single event magnitudes  $M$ . Nowadays, there is widespread consensus that the moment magnitude  $M_w$ , calculated via the measurement parameter *seismic moment*, is a physically reasonably defined non-saturating magnitude well suited for serving as such a unique event size parameter.  $M_w$  also allows to relate the seismic moment, estimated for rare strong historical earthquakes from geologic observations, to  $M_w$  determined from instrumental data (Lee, 2013). The problem, however, is that reliable magnitude-frequency relations require as long as possible a time span covered with complete data, but accurate instrumentally measured  $M_0$  and therefrom derived  $M_w$  values are available only since 1976 (see <http://www.globalcmt.org/CMTsearch.html>), in contrast to other magnitudes such as  $m_B$  and  $M_s$  which have been catalogued since 1900 (see Abe, 1981a, 1984; Di Giacomo et al., 2013a and b; Storchak et al., 2013; <http://www.isc.ac.uk/iscgem>),  $M_l$  since the 1930s and  $m_b$  since the 1960s. Thus, there is a need to make use of available conversion relationships between these different magnitudes and  $M_w$  in order to calculate for each event, for which no directly measured  $M_w$  value is available, a  $M_w$  proxy estimate.

However, no  $M_w$  proxy, based on an average statistical estimate calculated from a noisy data set, can be as good as a directly measured individual event  $M_w$ . Therefore, one should in magnitude-wise unified catalogs flag directly measured and proxy  $M_w$  values differently. This is often not done. Moreover, the conversion relationships for different types of magnitudes may differ strongly (e.g., Figs. 3.81 and 3.82 and Eqs. 3.124 and 3.125), be based on differently reliable regression procedures (e.g., linear or exponential; ordinary least squares; simple or general orthogonal; chi-square). Thus, their proxy estimates may have rather different error ranges and should then be assigned different weights in further use. Therefore, it is strongly recommended that catalogs preserve for all events the originally measured different types of magnitude data, indicate for all  $M_w$  proxies their origin (e.g., converted from  $M_l$ ,  $m_b$ ,  $M_s$  and with which relationship). Only this will allow later to reconstruct or correct values on the basis of improved measurement procedures and/or conversion relationships and permit a more correct assessment of the specific nature and



effects of past seismic event than it would be possible on the basis of their measured or assigned proxy Mw values alone.

The current version of the ISC-GEM Global Instrumental Earthquake Catalog (1900 – 2009) (Storchak et al., 2013; Di Giacomo et al., 2013b, Lee, 2013) assigns different estimated average error classes to its Mw values that range from  $\pm 0.1$  to  $> \pm 0.4$  m.u. They relate to either direct GCMT measurements, Mw proxies from Ms, Mw proxies from mb or Mw from bibliographical search (geological and other evidences). Also for the Swiss regional earthquake catalog, covering Switzerland and adjacent territories in Germany, France, Italy and Austria, a homogeneous moment tensor calibration has been performed, using conversion relationships for 13 different magnitude scales to Mw. The uncertainties for the Mw proxies estimated from data of these different original event magnitudes (mainly Ml and Md besides ISC mb and Ms) range between  $\pm 0.2$  and  $\pm 1.0$  m.u. (Braunmiller et al., 2005). Yadav et al, (2009) gave an example, how catalog pages may look like that preserve the original magnitudes, link unambiguously the used conversion relationships to respective proxy Mw estimates and calculate from a comparison of GCMT event Mw with respective proxy estimates the reliability of such proxies (average differences and standard deviations). This eases the later use and correct assessment of catalog data under very different application and research requirements which may go far beyond the need to calculate magnitude-frequency relationships for seismic hazard assessment. Otherwise, if catalogs are magnitude-wise reduced to Mw it might require for later generations a lot of efforts to reconstruct from station bulletins or other notes the related original event magnitude types, as it took Abe (1981a and 1984) years of work to reconstruct the original mB and Ms values behind the “unified” M values published by Gutenberg and Richter (1954) in “Seismicity of the Earth” or in Duda’s (1965) catalog.

In the following section we summarize a selection of important so-called scaling relationships between magnitudes and a variety of geometric as well as kinematic and dynamic source parameters. In this context we will also look into the need (or not) of revising the currently used formulas for Mw and Me determination in the light of modern data on the relationships between mB\_BB and Ms\_20 with released seismic energy.

### 3.3 Similarity conditions and seismic scaling relations (P. Bormann)

#### 3.3.1 Similarity of seismic sources and the definition and use of seismic scaling relations

The *similarity* of seismic sources over a wide range of magnitude or other “size” parameters are the precondition for the derivation of scaling relationships, e.g., between the shape and level of radiated seismic spectra as a function of frequency, as in Fig. 3.5. Similarity of earthquake ruptures exists under certain conditions of static (geometric) and dynamic similarity. With the assumption of a constant stress drop, e.g., one gets

$$W/L = k_1 \quad \text{i.e., a constant fault aspect ratio} \quad \text{and} \quad (3.131)$$

$$\bar{D}/L = k_2 \quad \text{i.e., constant strain } \alpha. \quad (3.132)$$

One can combine Eqs. (3.131) and (3.132) with the definition of the seismic moment  $M_0 = \mu \bar{D} W L = \mu k_1 k_2 L^3$  and gets  $M_0 \sim L^3$  which is valid for source dimensions smaller than the thickness of the *seismogenic layer*.

In addition there is a dynamic similarity, namely, the rise time  $t_r$  required for reaching the total displacement, i.e., the duration of the source-time function, is

$$t_r = k_3 \times L/v_{cr} \quad (3.133)$$

with  $v_{cr}$  the crack or rupture velocity (see Fig. 3.4). This is equivalent to the Eq. (3.132) of constant strain. Lay and Wallace (1995) showed that this results in period-dependent amplitudes of seismic waves which scale with the fault dimension. For periods  $T \gg t_r$  the amplitude does not depend on fault length  $L$ . This corresponds to the plateau of the "source displacement spectrum". But if  $T \ll t_r$  then the amplitudes scale as  $1/L^2$  or  $f^{-2}$  (see Fig. 3.5). This explains the saturation effect when analyzing frequencies much higher than the corner frequency of the source spectrum.

Both in earthquake and engineering seismology the term "scaling law" is widely used. According to Boore (1983) the Fourier spectrum of an earthquake record can be represented as the product of the source, propagation path and site terms. Then it would be very helpful if the entire source term could be represented by just one parameter, e.g., by assuming similarity of the spectra of all earthquakes. This was first done by Aki (1967) when investigating the dependence of the amplitude spectrum of seismic waves on source size on the basis of two different dislocation models of an earthquake source. One has been the Haskell (1966)  $\omega^{-3}$  model, in which the spectral amplitudes decay well beyond the corner frequency  $f_c$  with the third power, and the other one an  $\omega^{-2}$  model, as in Fig. 3.5. Aki considered the surface-wave magnitude  $M_s$  to be a most suitable parameter for scaling the source spectrum. Later this was commonly substituted by the scalar seismic moment  $M_0$ , as in the Geller (1976) scaling laws for body and surface waves and in Fig. 3.5, or by the moment magnitude  $M_w$ , which itself has been scaled to  $M_s$  (see section 3.2.7.1). According to Aki (1967) the theoretical  $\omega^{-2}$  model agreed better with the observed spectra than the  $\omega^{-3}$  model, which would also imply absolute saturation of spectral amplitudes for  $f \gg f_c$ , as also in the Geller (1976) models, according to which  $m_b$  would already completely saturate at 5.9 and  $M_s$  at 8.0. Yet,  $M_s$  has already been observed up to 8.9 (see, e.g., Bormann et al., 2009) and with respect to  $m_b$  Geller had only pre-1976 values available which were based on amplitude measurements within the first few cycles only, as nowadays still the  $m_b$ (IDC-CTBTO). The IASPEI standard  $m_b$ , however, which always measures the largest amplitude in the whole P-wave train, measures values up to about 7.6.

Beresnev (2001 and 2008) rightly questions the assumption of a strict similarity of earthquake source processes and even Aki (1967) hints already in his pioneering paper to "...some indications of departure from similarity". It is rather obvious from wave physics that low-frequency measures such as the scalar seismic moment  $M_0$ , being according to equation (3.1) only a static measure of earthquake size, or the moment magnitude  $M_w$  derived therefrom, cannot provide direct information about the kinematics and dynamics of the rupture process. While the low-frequency asymptote of the displacement spectrum and thus  $M_0$  are solely controlled by the final static offset of the fault rupture, the actual stress drop (see, e.g., Abe, 1982; Choy and Boatwright, 1995) and rupture velocity (Fig. 3.6 and Venkataraman and Kanamori, 2004a and b) may vary strongly for equal moment earthquakes. These latter source parameters, however, control the corner frequency and with it the relative high-frequency content of the source spectrum and thus the amount of radiated seismic energy per unit of  $M_0$

(see Fig. 3.15). Therefore, the single source parameter scaling of seismic spectra is not a viable model and we will look into the scaling problem in a wider context.

We should always keep in mind that ANY earthquake model is a (usually gross) simplification of the complex earthquake rupture process. Models tend to reduce as much as possible the number of controlling parameters, for which usually no directly measured and sufficiently accurate empirical data available and in order to make the models tractable. A good example may be the extensive debate by Scholz (1994 a and b), Romanowicz (1994), Romanowicz and Rundle (1994) and Sornette and Sornette (1994) about a paper by Scholz (1982) dealing with the scaling laws for large earthquakes. He interpreted the observation that the mean slip in large earthquakes is linearly proportional to fault length  $L$  and does not correlate with fault width  $W$  on the basis of two possible models for large earthquakes: a)  $W$  models, in which stress drop and slip are determined by fault width  $W$  and b)  $L$  models, in which these two parameters are fundamentally determined by  $L$ . But both models come to opposite conclusions with respect to mean particle motion and rise time of the source-time function. In a later paper Scholz (1997) could show, however, that such conflicts do not exist when the size distribution of faults is properly taken into account. Wang and Ou (1998), using the large data base by Wells and Coppersmith (1994), argued that  $W$  is independent of  $L$  for large earthquakes, that average displacement  $\bar{D}$  relates to  $L$  as  $\bar{D} \sim L$ , seismic moment to  $L$  as  $M_0 \sim L^2$  whereas  $\bar{D}$  is independent of  $W$ . Accordingly, they concluded that the  $L$ -model proposed by Scholz (1992) is more appropriate than the  $W$ -model suggested by Romanowicz (1992) to describe the scaling of earthquake faults.

In any event, a rectangular fault plane is nothing but a rough approximation of a more or less irregularly shaped rupture area for medium to strong size earthquakes. According to Geller (1976), empirical data for earthquakes are agreeable with assuming an average aspect ratio  $L/W = 2$  with a scatter of about 2. Purcaru and Berckhemer (1982), however, assume for great earthquakes aspect ratios even up to 30. The “true”  $L/W$  depends on the rupture dimensions with respect to the thickness  $h_{sg}$  of the seismogenic zone in the Earth crust, respectively lithosphere, which is capable of brittle fracturing.  $h_{sg}$ , however, depends on crustal thickness, age and heat flow conditions, which vary from region to region. In California, with its young crust and high heat flow,  $h_{sg}$  is only about 12 to 15 km. Deeper earthquakes hardly ever occur. The maximum width of an earthquake rupture is  $W_{max} = h_{sg}/\sin\delta$ , with  $\delta$  the fault dip. Accordingly, the ratio  $L/W$  may vary between about 1 and about 10, or even more, e.g., for mega-events such as the great Mw 9.3 Sumatra-Andaman earthquake of December 2004 with a rupture length of about 1300 km.

If one defines, according to Scholz (1982), as small earthquakes those which have a source radius  $r \leq W_{max}/2$ , then they likely can be represented reasonably well by assuming even circular fault planes with radius  $r$  in an elastic medium, whereas large earthquakes with  $r > W_{max}/2$  are better treated as rectangular ruptures with one edge at the free surface and only  $L$  growing with magnitude.

Yet, even when using only circular source models for deriving some crucial source parameters from seismic spectra, such as source radius  $r$ , rupture area  $A$ , average displacement  $\bar{D}$  and stress drop  $\Delta\sigma$ , then the calculated values may differ significantly even for identical input values of  $M_0$  and  $f_c$  estimated from identical seismic spectra. E.g., the most common models by Brune (1970) and Madariaga (1976; models I and II) yield values for these inferred parameters that differ by factors between about 1.7 and 6 (see EX 3.4). Model inherent uncertainties in this range or even larger ones should therefore be kept in mind when using the scaling relationships. The main advantage of using quantifiable models is not that

they tell us the absolute truths with respect to the derived parameters but that they reveal, in a consistent and reproducible manner, relative differences between these parameters for different events, provided that the model assumptions hold for all of them, and to assess their variability with magnitude, space and time.

In the following we present a selection of various relationships between magnitudes with physical or geometrical parameters of the seismic source, such as radiated seismic energy and seismic moment, stress drop, duration of rupture, area or length and width of rupture, fault dislocation, or between magnitudes and earthquake effects at some distance from the source, such as the area of felt shaking, or the intensity of shaking as a function of distance from the source (attenuation laws). If any of these parameters appears to be related in a systematic and predictable manner over a wide range of earthquake size, scaling “laws” based on similarity conditions may be inferred. Seismic scaling laws then allow to roughly estimate one parameter from another (e.g.,  $E_s$  from magnitude, or  $M_0$  from field evidence such as surface rupture length and/or displacement). Therefore, the knowledge of theoretically well founded scaling laws or empirical scaling relationships is of crucial importance for both probabilistic and deterministic seismic hazard analyses. They aim at assessing the future earthquake potential of a region on the basis of data from past events, dating back as far as possible.

Scaling laws are often the only way to estimate parameters of historical earthquakes for which no instrumental measurements of magnitude, seismic energy or scalar seismic moment are available. Specifically, one often has to make reasonable estimates of the size of the largest earthquake that might have occurred at or could be generated by a particular fault or fault segment and of the kind of seismic spectrum it might (have) radiate(d). However, as stated above, seismic sources differ not only in their geometrical size, shape, aspect ratio, rupture velocity and average slip. Ambient stress conditions and stress drop, the dominant modes of faulting and related seismic source spectra may also differ significantly from region to region. For instance, events of the same seismic moment may radiate seismic energies which differ by 2 to 3 orders (see sections 3.1.2.3 – 3.1.2.5, 3.2.7.2 and IS 3.5). Therefore, globally-derived scaling relations may not be appropriate for use in some areas or outside the magnitude/size ranges for which they have been derived. Regional scaling laws should be used, therefore, whenever available, particularly when inferences have to be drawn on regional seismic strain rates, and seismic hazard, the latter being mainly controlled by the frequency of occurrence and the potential of earthquakes to generate strong high-frequency motions.

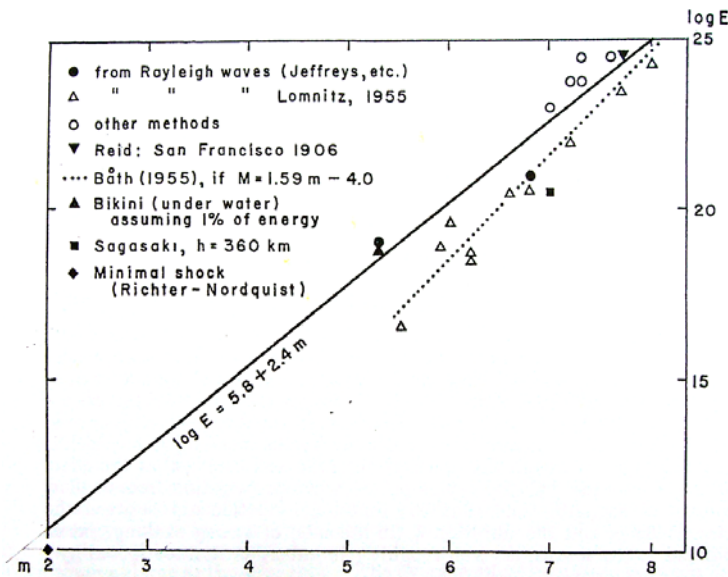
### 3.3.2 Relationships between seismic energy, seismic moment and magnitudes (P. Bormann and D. Di Giacomo)

Gutenberg (1956) proposed the following relationship between seismic energy  $E_s$  (in units of Joule ;  $1 \text{ J} = 10^7 \text{ erg}$ ) and the so-called unified body-wave magnitude  $m$ , which he measured in a wide range of periods between about 2 and 20 s and later renamed  $mB$ :

$$\log E_s = 2.4 m - 1.2. \quad (3.134)$$

Eq. (3.134) is based on a rather small data set (see Fig. 3.83). Gutenberg (1956) estimated the error in  $\log E_s$  calculated via (3.134), which is in fact not a proper standard regression relation, to be “*probably less than one unit.*” Together with  $mB = 2.5 + 0.63 M_s$  in Gutenberg and Richter (1956a), Eq. (3.134) yields

$$\log E_S = 1.5 M_s + 4.8. \quad (3.135)$$



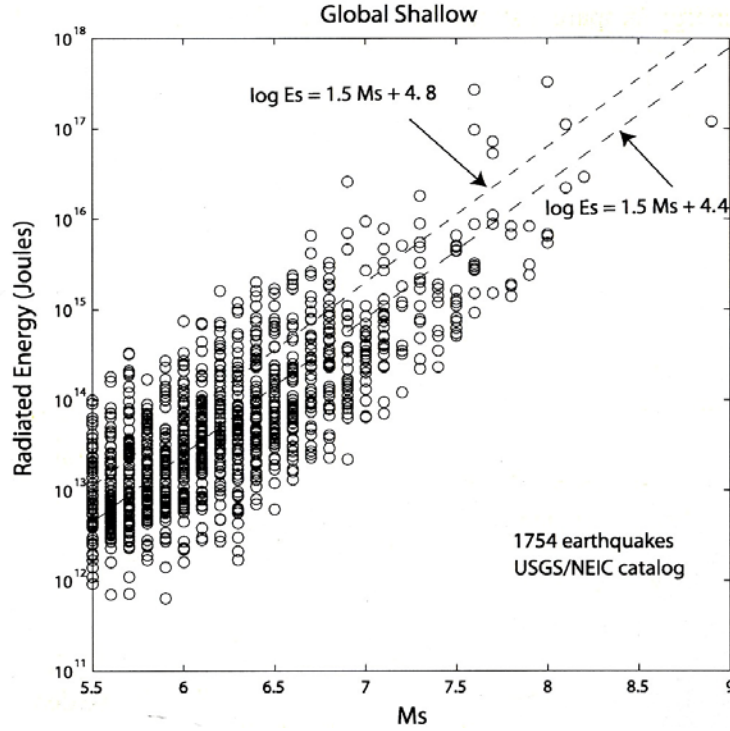
**Fig. 3.83** Primary relation, between seismic energy  $E_S$  as a function of the “unified” magnitude  $m$ , as published first by Gutenberg (1956).  $m$  is an average weighted event  $m_B$ .

Eq. (3.135) was first proposed by Richter (1958), however published with a printing error in the constant (4.4 instead of 4.8). Up to now it is the most widely applied relationship between seismic energy and magnitude and generally referred to as a Gutenberg-Richter relationship. One should be aware, however, that Gutenberg preferred and authorized with any of his own - or co-authored - publications only Eq. (3.134). In Gutenberg and Richter (1956b) he stated that “*Present studies are directed toward a scale based on the quotient ( $A/T$ ) rather than on amplitudes. ...It is believed that magnitudes determined from body waves of teleseisms are more coherent with the original scale than those from surface waves, which have been in use as the general standard*”. The reason is that  $m_B$ , determined by measuring  $A/T$  in a wide range of periods, is more directly related to ground motion velocity and thus to seismic energy released by earthquakes in a wider range of magnitude than  $M_s$ , which is measured around 20 s only. Kanamori (1977), however, when investigating the energy release in great earthquakes and developing the concept of a non-saturating seismic moment magnitude  $M_w$  (see 3.2.5.3), considered only Eq. (3.135) because of the long corner periods of strong and great earthquakes.

Later, with more  $M_s(20)$  data of the NEIC in a wider magnitude range being available as well as direct  $E_S$  determinations from analyzing broadband velocity records (see IS 3.6) Choy and Boatwright (1995) found

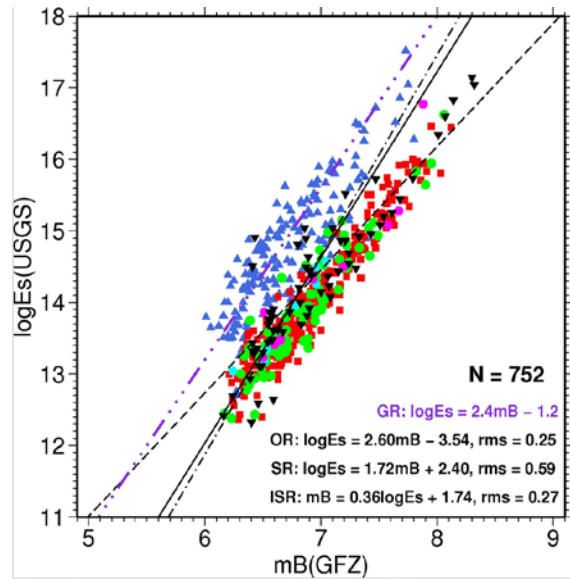
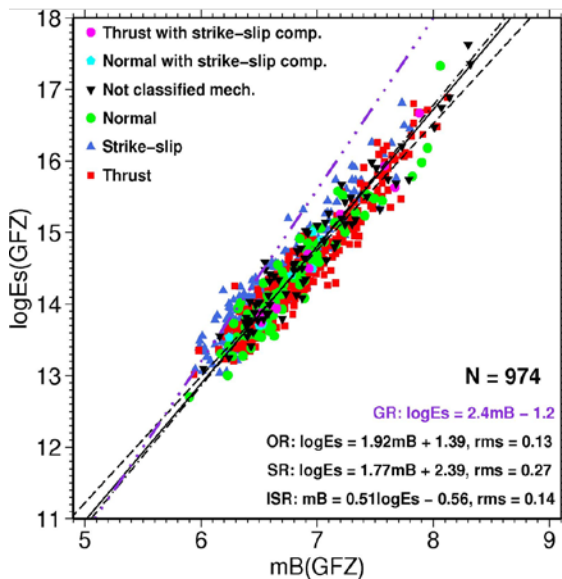
$$\log E_S = 1.5 M_s + 4.4. \quad (3.136)$$

be a better approximation to the data, confirmed also by later compilations with many more data (Choy et al., 2006; see Fig. 3.84).

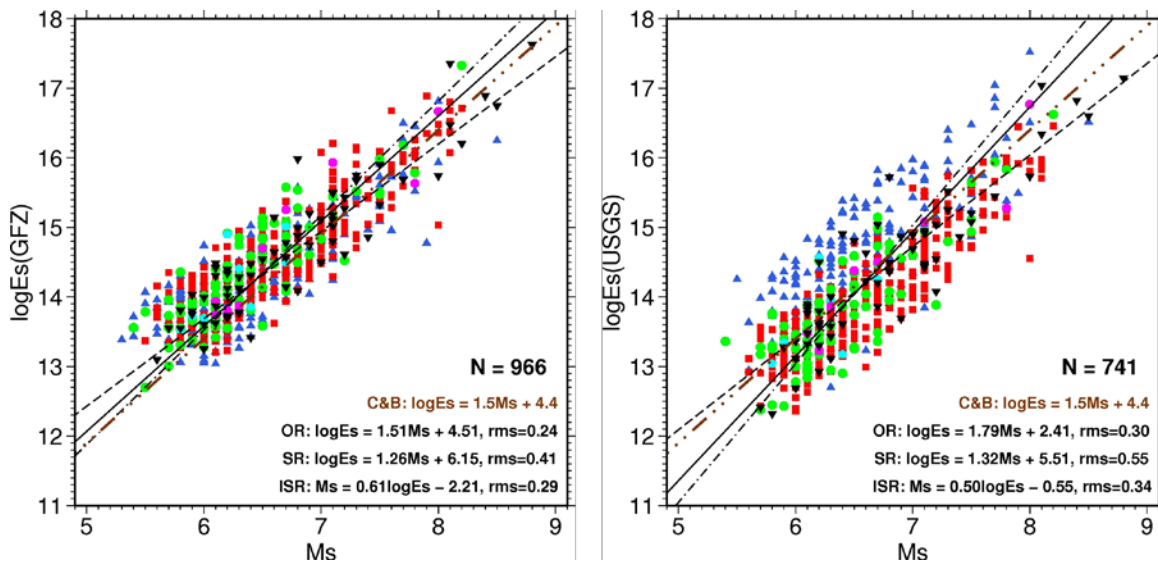


**Fig. 3.84** Relationship between radiated energy  $E_S$  and  $M_s$  as calculated by the USGS. (Copy of Figure 1 by Choy et al. (2006) in Abercrombie R., McGarr, A., and Kanamori, H. (eds): Radiated energy and the physics of earthquake faulting, *AGU Geophys. Monogr. Ser.* **170**, p. 45; © American Geophysical Union).

However, Eq. (3.136) and the respective regression line in Fig. 3.84 did not result from an optimal least square fit through the data. Rather, in the interest of assuring widest possible continuity with Eq. (3.135) and the later definition of the  $M_w$  scale, on which decades of research results rest, these authors preferred to revise only the constant by looking for the best fitting 1.5-slope line through the  $E_S$ - $M_s$  data cloud. Yet, when searching for best fitting orthogonal linear regressions through modern  $\log E_S$ ,  $mB(BB)$  and  $M_s(20)$  data, then rather different relationships have been found [see Figs. 3.85 and 3.86 and relations (3.137) to (3.140)].



**Fig. 3.85** Comparison of Gutenberg's (1956)  $\log E_s$ -mB relationship (blue hatched-3dot line) with standard (SR and ISR) and orthogonal (OR) regression relationships (black lines; hatched, hatch-dotted and full line, respectively) between  $\log E_s$  and automatically measured IASPEI standard mB\_BB values according to the GFZ procedure described by Bormann and Saul (2008). The mB data were kindly provided by J. Saul in 2010. **Left:**  $\log E_s$ (GFZ) is based on  $E_s$  determined by applying the automatic GFZ procedure for processing broadband velocity records as described by Di Giacomo et al. (2010 a and b); **Right:**  $\log E_s$ (GS) is based on  $E_s$  values published by the US Geological Survey (USGS). They are determined interactively according to the procedure described by Choy and Boatwright (1995) as well as in IS 3.6. The colored data point symbols relate to different types of source mechanism as in the legend at the upper left. For formulas see lower right inserts.



**Fig. 3.86** Comparison of Choy and Boatwright's (1995)  $\log E_s$ - $M_s$  relationship (brown hatched-3dot line) with standard and orthogonal regression relationships (symbols as in Fig. 3.87) between  $\log E_s$  values determined at the GFZ (left) and the US Geological Survey (USGS) (right), respectively, with USGS  $M_{s,20}$  values. The Richter (1958) relationship  $\log E_s = 1.5 M_s + 4.8$  would be parallel to the C&B line but 0.4  $\log E_s$  units higher.

The following conclusions can be drawn from Figs. 3.85 and 3.86:

- $\log E_s$ (GFZ) correlates much better than  $\log E_s$ (USGS) with both IASPEI standard mB(BB) and  $M_s(20)$ . The reason is that the GFZ procedure, in contrast to that of the USGS, does not apply any source mechanism corrections, which are not common for all classical magnitude scales. Such corrections to  $\log E_s$ (USGS) produce the large average off-set between the data points for strike-slip earthquakes which Choy interpretes in terms of exceptionally large apparent stress, respectively stress drop, in strike-slip environments (see IS 3.5).
- The slopes and constants of the standard OR regression relations through modern  $\log E_s$  data differ significantly from the original Gutenberg and Richter (1956a)  $\log E_s$ -mB relationship in Eq. (3.135). Remarkably, however, the latter is a rather good average fit through the  $\log E_s$ (USGS) data for strike-slip earthquakes in Fig. 3.86 right. For other types of source mechanisms, however, it would overestimate  $\log E_s$  on average by about 1 unit.

- The  $\log E_S(\text{GFZ})$ - $mB(\text{GFZ})$  OR relationship has a very high correlation coefficient  $R_{XY}$  and a small orthogonal rms (RMSO). It reads:

$$\log E_S(\text{GFZ}) = 1.92 mB(\text{GFZ}) + 1.39 \text{ with RMSO} = 0.13. \quad (3.137)$$

Values of  $\log E_S$ , calculated with (3.137) from  $mB$ , are at  $mB = 6.0$  and  $8.3$  about  $-1.3$  and  $-2.4 \log E_S$  units smaller than those derived via the Gutenberg-Richter formula (3.135), which yields already constantly values that are  $0.4$  units larger than those derived by the Choy and Boatwright (1995) relation (3.136).

- The  $\log E_S(\text{USGS})$ - $mB(\text{GFZ})$  OR relationship has a much larger RMSO:

$$\log E_S(\text{USGS}) = 2.60 mB(\text{GFZ}) - 3.54 \text{ with RMSO} = 0.30. \quad (3.138)$$

- Values of  $\log E_S$ , calculated with (3.138) from  $mB$ , are at  $mB = 6.0$  and  $8.3$  about  $-1.1$  and  $-1.7 \log E_S$  units smaller than those derived via the Gutenberg-Richter formula (3.135).
- The much larger and not random scatter of  $\log E_S(\text{USGS})$  data explains the large slope offset between the orthogonal and SR1 standard regression relations (see Fig. 3.86 right). This offset is almost negligible for the  $\log E_S(\text{GFZ})$ - $mB$  relationship (see Fig. 3.86 left). For general discussion of the relationship between SR, ISR and OR see Bormann et al. (2007) and Castellaro and Bormann (2007).
- The large difference between the regression relationships based on apparently the same type of  $\log E_S$  data highlights again the importance to specify with a unique nomenclature different procedures applied to calculate such data. Otherwise, procedural effects might be misinterpreted as being effects of real nature or the latter might be hidden due to procedural noise.
- The differences between the Richter (1958)  $\log E_S$ - $M_s$  relationship and the regression relations between different versions of  $\log E_S$  and  $M_s(20)$  are less pronounced yet still significant.
- The orthogonal regression between  $\log E_S(\text{GFZ})$  and  $M_s(20)$  of the USGS confirms almost perfectly the Choy and Boatwright (1995)  $\log E_S$ - $M_s$  relationship in Eq. (3.136). It reads:

$$\log E_S = 1.51 M_s + 4.51 \text{ with RMSO} = 0.24 \quad (3.139)$$

and yields in the whole considered magnitude range between  $M_s 5.2$  and  $8.8$  values of  $\log E_S$  that are only between  $+0.16$  and  $+0.20$  units of  $\log E_S$  larger than those calculated via the Choy and Boatwright (1995) formula (3.136). This is within the uncertainty range of energy calculations in general.

- In contrast, despite still reasonably good agreement between the orthogonal  $\log E_S(\text{USGS})$ - $M_s$  relationship

$$\log E_S = 1.79 M_s + 2.41 \text{ with RMSO} = 0.30 \quad (3.140)$$

and (3.136) for medium-size earthquakes, the differences for converting  $M_s$  into  $\log E_S$  reach for  $M_s = 5.2$  and  $8.8$  values of  $-0.42$  and  $+0.56 \log E_S$  units, respectively, which is unacceptably large. Thus, the orthogonal regression between  $\log E_S(\text{GFZ})$  and  $M_s(\text{USGS})$  constrains the Choy and Boatwright (1995) formula (3.136) much better than the orthogonal regression between  $\log E_S(\text{USGS})$  and  $M_s(\text{USGS})$  or the conditional regression applied by these authors to their data (see Fig. 3.84 and related comments).



- When substituting in the relationships (3.139) and (3.140)  $M_s$  by  $M_e$ , as done by Choy and Boatwright (1995) when deriving the currently applied  $M_e$  standard formula (3.72)  $M_e = (\log E_s - 4.4)/1.5$ , then relationship (3.139) would yield

$$M_e = (\log E_s - 4.51)/1.51 \quad (3.141)$$

and (3.140)

$$M_e = (\log E_s - 2.41)/1.79. \quad (3.142)$$

- When calculating for  $\log E_s$  values of 12 and 18, corresponding to  $M_e$  values of approximately 5 and 9, via relationship (3.141)  $M_e$  then the values are in this  $\log E_s$  range only between 0.11 m.u. and 0.14 m.u. smaller than values calculated with the current  $M_e$  formula (3.72). In contrast, relationship (3.142) yields values that differ in this range between +0.40 m.u. and -0.36 m.u.
- Moreover, the RMSO scatter of  $\log E_s$ (GFZ) over both USGS  $M_s$ (20) and  $mB$ (GFZ) is significantly smaller than that of  $\log E_s$ (USGS). This questions, within the unavoidable range of data scatter, the suitability of applying mechanism dependent corrections in the USGS procedure to the calculated  $E_s$  values, or at least the size of such corrections applied to strike-slip earthquakes and to more high-frequency data for smaller earthquakes in general.

The latter point has repeatedly been discussed, e.g., by Newman and Okal (1998), Pérez-Campos and Beroza (2001), and Bormann and Di Giacomo (2011). Also Okal and Talandier (1989) pointed out that a “*..magnitude concept, which ignores the exact focal geometry, could be a more robust measure of the true size of the event than one correcting for an expected small source excitation, but failing to account for non-geometrical effects*” such as scattering or multi-pathing of energy in inhomogeneous media. Moreover, calculated average point-source fault-plane solutions are afflicted with unavoidable uncertainties in the order of several degrees (up to about 10°). Also, the orientation and radiation pattern of sub-segments of rough non-planar faults and rupture irregularities may significantly differ from those calculated for planar point source fault-plane solutions. Such non-accountable details in real Earth and earthquake ruptures, however, effect especially higher frequencies that largely contribute to the  $E_s$  estimates, the more the smaller the earthquakes and the radiated wavelength are. This may also explain why Schweitzer and Kverna (1999) could not prove any significant influence of source radiation patterns on globally observed short-period  $m_b$  estimates.

Finally, we again make the point that it is rather arbitrary to scale the energy magnitude  $M_e$ , based on  $E_s$  estimates resulting from the integration of squared velocity broadband P-wave records in a wide range of periods between about 1 and 100 s, to narrow-band long-period surface-wave magnitude  $M_s_{20}$ . This was surely not the intention of Gutenberg (1956) when he proposed (3.134) based on body-wave magnitude measurements in a wide range of period. This inappropriate scaling can only be understood by the fact that during the decades of WWSSN dominated seismology no more medium-period broadband records were available in the “Western World” and measuring “old-fashioned” broadband  $mB$  was just out of debate. Yet, with  $mB_{BB}$ , based on velocity-broadband P-wave records, being now again a IASPEI recommended body-wave magnitude standard that “saturates” much later than short-period  $m_b$ , it is strongly recommended to reconsider the scaling of  $M_e$  to  $mB_{BB}$  instead to long-period  $M_s_{20}$ . Also Abe (1982) came to the conclusion that the size of earthquakes, both deep and shallow, is consistently quantified by the amount of seismic energy expressed by the broadband body-wave magnitude  $mB$ .

When following the practice of Kanamori (1977) and Choy and Boatwright (1995) of substituting  $M_s$  by  $M_w$ , respectively  $M_e$ , when deriving (3.68) and (3.72), then the respective  $\log E_S$ -mB relationships (3.138) and (3.139) would yield the following alternative  $M_e$  relationships:

$$M_e = (\log E_S - 1.39)/1.92 \quad (3.143)$$

and

$$M_e = (\log E_S + 3.54)/2.60. \quad (3.144)$$

Considering the same range of  $\log E_S$  between 12 and 18 as above, then, when compared with  $M_e$  values calculated by the current  $M_e$  formula (3.72), (3.143) would yield values that differ between +0.46 m.u. und -0.42 m.u., and (3.144) values that differ even more, namely up to +0.91m.u. and -0.79 m.u., respectively. At values around 7.6, however, the current  $M_s$ -scaled and new mB-scaled  $M_e$  values would agree well. This makes sense, because according to the scaling law in Fig. 3.5, right panel, the average seismic source velocity spectra have their maximum velocity amplitudes at this magnitude on average around 20 s periods.

The general conclusion to be drawn from such a reasonable rescaling of the broadband body-wave  $M_e$  to broadband mB is, however, that the current  $M_s$ -scaled  $M_e$  formula may significantly underestimate  $M_e$  at magnitudes less than 7.6 and overestimate  $M_e$  for stronger earthquakes. Yet, such a physically logical and even demanding rescaling of  $M_e$  would question many of the conclusions derived since the 1990s from  $\log E_S$  and  $M_e$  data calculated on the basis of their current scaling to  $M_s$ .

We will now refer to some earlier results and considerations on other  $\log E_S$ -magnitude relationships. From theoretical considerations, Randall (1973) derived a relationship between  $E_S$  and the local magnitude  $M_l$  which was later confirmed empirically by Seidl and Berckhemer (1982) as well as by Berckhemer and Lindendorf (1986). On the basis of direct energy calculations for earthquakes from the Friuli region, Italy, using digital broadband records of the Gräfenberg array in Germany, the latter obtained:

$$\log E_S \sim 2.0 M_l. \quad (3.145)$$

This is close to the empirical findings by Gutenberg and Richter (1956a) ( $\log E_S \sim 1.92 M_l$ ) for southern California and the more recent one by Kanamori et al. (1993). The latter found

$$\log E_S = 1.96 M_l + 2.05 \quad (3.146)$$

in the magnitude range  $1.5 < M_l < 6.0$ . For  $M_l > 6.5$   $M_l$  saturates.

For short-period body-wave magnitudes mb, Sadovsky et al. (1986) derived the relationship

$$\log E_S = 1.7 mb + 2.3 \quad (3.147)$$

that is applicable for both earthquakes and underground explosions. **Note:** According to the slopes in the above equations one unit of magnitude increase in  $M_s$ , mb,  $M_l$  or mB, respectively, corresponds to an increase of  $E_S$  by a factor of about 32, 50, 100 and 250 times, respectively!

In this context one should mention that in the countries of the former USSR the energy scale after Rautian (1960),  $K = \log E_S$  (with  $E_S$  in J), is widely used and given in the catalogs. It is based on the same elements as any other magnitude scale such as an empirical calibration

function and a reference distance (here 10 km).  $K$  relates on average to magnitude  $M$  approximately via

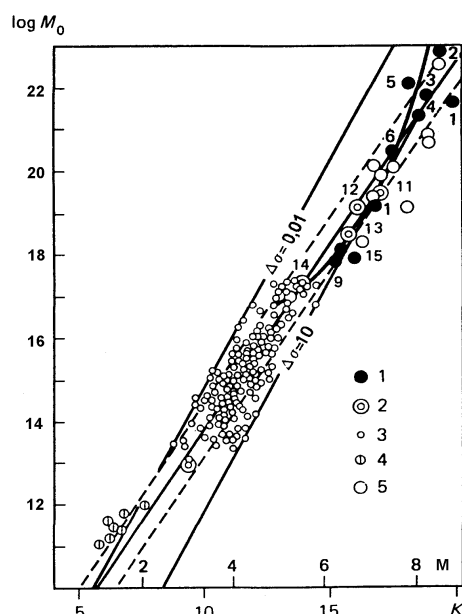
$$K = 1.8 M + 4. \quad (3.148)$$

Riznichenko (1992) summarized data and relationships published by many authors (see Fig. 3.87) between magnitude  $M$  and  $K$  on the one hand and  $\log M_0$  on the other hand. Depending on the range of distance and size,  $M$  stands here for  $M_L$ ,  $m_b$ ,  $m_B$  or  $M_s$ . For more details on  $K = \log E_s$  and other types of magnitude see section 3.2.4.6 and IS 3.7.

### 3.3.3 Relationship between $M_0$ , $E_s$ and magnitude

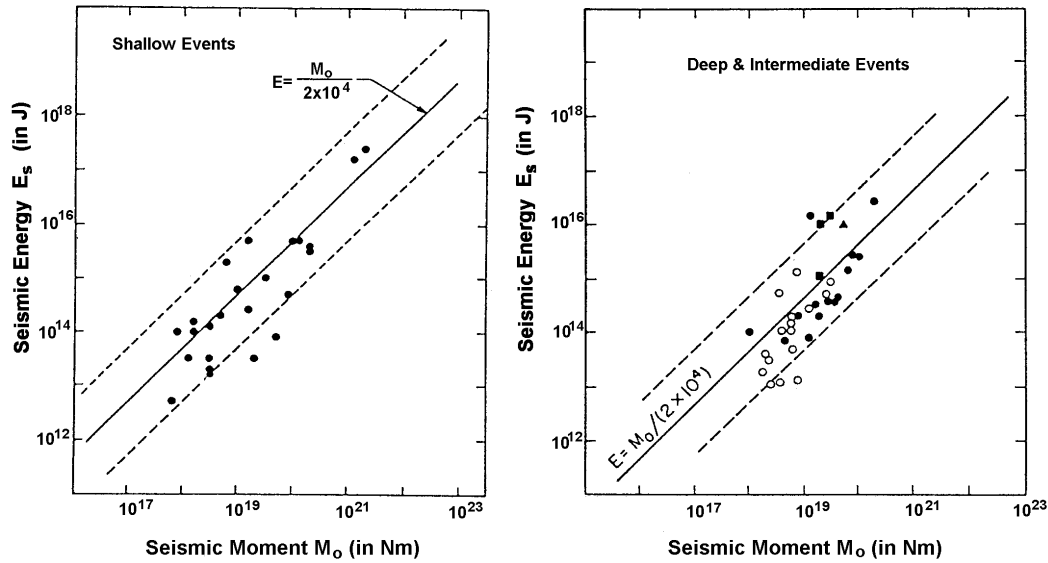
(P. Bormann and D. Di Giacomo)

Kanamori (1983) published linear relationships between  $\log E_s$  and  $\log M_0$  for both shallow and intermediate to deep events (see Fig. 3.88). These two relationships are rather similar and correspond, on average, to a ratio of moment scaled energy of  $E_s/M_0 = 5 \times 10^{-5}$ , or, according to Okal and Talandier (1989), to a so-called “slowness parameter”  $\Theta = \log (E_s/M_0)$  of -4.3. Kanamori (1977) assumed this value to be representative for global earthquake sets and used it for developing the moment magnitude scale  $M_w$ . However, as previously mentioned in the sub-sections 3.2.7.1 and 3.2.7.2 on moment and energy magnitudes, scaling laws based on a constant ratio  $E_s/M_0$  must be used with great caution. Later investigations revealed sometimes significant deviations from the Kanamori ratio of  $E_s/M_0$  (e.g., Kikuchi and Fukao, 1988; Choy and Boatwright, 1995; Bormann and Di Giacomo, 2011). Variations in  $E_s/M_0$ , respectively  $\Theta$ , are mainly due to local and regional differences in source mechanism, related stress drop, rupture velocity and time history of the rupture process.



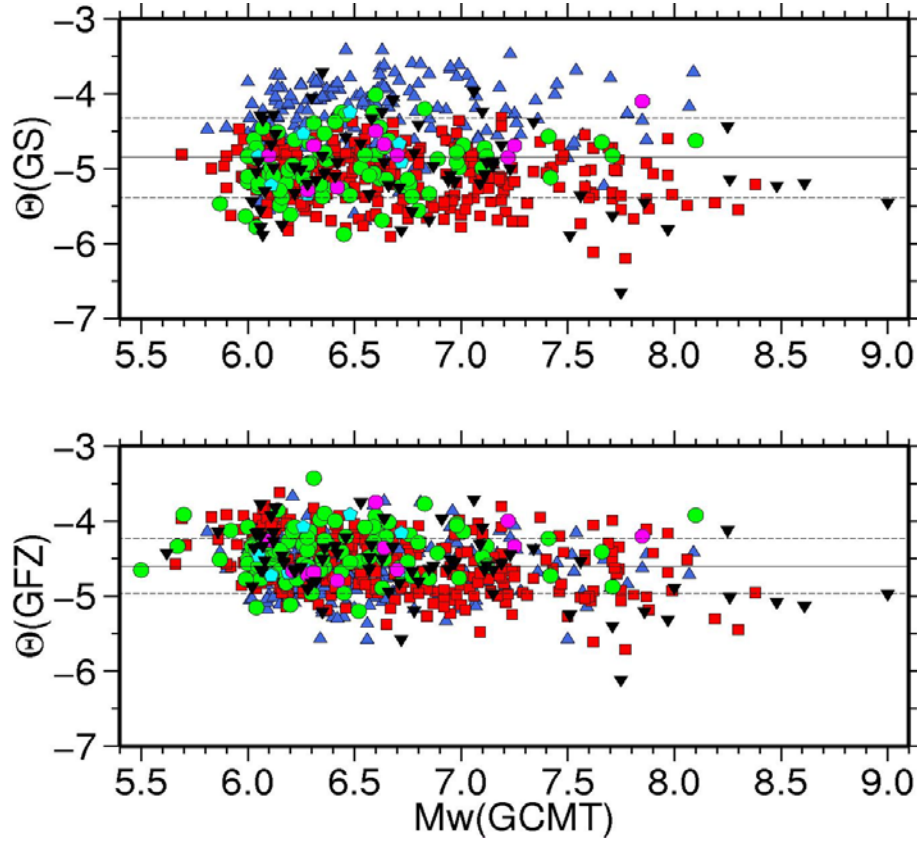
**Fig. 3.87** Correlation between seismic moment  $M_0$  (in  $\text{Nm} = \text{J}$ ), magnitude  $M$  and Rautian's (1960) energy class  $K$  according to a compilation of data from many authors. Related stress drop  $\Delta\sigma$  has been given in MPa (full straight lines). Broken lines mark the 68% confidence interval. 1 - large global earthquakes; 2 - average values for individual regions; 3 - earthquakes in the western USA; 4 - micro-earthquakes in Nevada; 5 -  $M_0$  determinations from field data;

6 to 15 - individual values from different regions (modified from Riznichenko, 1992, Fig. 1; with permission from Springer-Verlag).



**Fig. 3.88** Relations between seismic moment  $M_0$  and energy  $E_s$  for shallow events (left) and intermediate to deep events (right) according to Vassiliou and Kanamori (1982). The solid line indicates the relation  $E_s = M_0 / (2 \times 10^4)$ , which had been suggested by Kanamori (1977) on the basis of elastostatic considerations (modified from Kanamori, 1983 in *Tectonophysics*, Vol. 93, p. 191 and 192, with permission from Elsevier Science).

Fig. 3.89 depicts the  $\Theta$  values for a large global event data set, plotted over  $M_w$  and calculated as the difference between  $\log E_s$  (USGS=GS) (upper panel), respectively  $\log E_s$  (GFZ) (lower panel), and  $\log M_0$  (GCMT). Note the difference in data scatter and in average  $\Theta$  values:  $\Theta$ (Kanamori, 1977) = -4.3,  $\Theta$ (GFZ) = -4.6, and  $\Theta$ (USGS) = -4.8. For individual earthquakes,  $\Theta$  may vary between about  $-3.0 > \Theta > 7.0$  (see also Weinstein and Okal, 2005, and Lomax and Michelini, 2009a). The latter authors identified values below -5.7 to be a reliable indicator for very slow tsunami earthquakes.



**Fig. 3.89** Comparison between the  $\Theta = \log(E_s/M_0)$  values calculated as the difference between  $\log E_s(\text{USGS}=\text{GS})$  (upper panel), respectively  $\log E_s(\text{GFZ})$  (lower panel), and  $M_w(\text{GCMT})$  in the magnitude range between 5.5 and 9.0. There is a tendency of decreasing  $\Theta$  with the magnitude of strong and great, mainly subduction zone earthquakes. The different color symbol relate to different types of source mechanism, e.g., upright blue triangles to strike-slip (SS) earthquakes. Note again, as in Figs. 3.85 and 3.86, the much larger data scatter of the USGS data set with its distinct separation of SS earthquakes, which have been corrected for theoretically expected reduced radiation coefficients according to Boatwright and Choy (1986).

$M_w$  has been defined via the  $\log E_s$ - $M_s$  relationship. Following the discussions in Bormann and Di Giacomo (2011), the definition formula for  $M_w$  can be decomposed as follows:

$$M_w = (\log M_0 - \mathbf{A} \text{ const. in the } \log E_s\text{-}M_s \text{ relation} + \mathbf{B} \text{ average } \Theta) / \mathbf{C} \text{ slope of the } \log E_s\text{-}M_s \text{ relation.} \quad (3.149)$$

Since Kanamori (1977) and Hanks and Kanamori (1979) based their  $M_w$  formula on the Richter (1958)  $\log E_s$ - $M_s$  relations with a constant of 4.8 and slope 1.5, and on an assumed average  $\Theta = -4.3$ , they got the  $M_w$  relation, which reads in its standard form of writing

$$M_w = (\log M_0 - 4.8 - 4.3)/1.5 = (\log M_0 - 9.1)/1.5. \quad (3.150)$$

Yet, modern direct  $\log E_s$  determinations and their **orthogonal** regression with the respective event  $M_s(\text{USGS}) = M_{s\_20}$  data (see Fig. 3.87) as well as the average values of  $\Theta$  derived from Fig. 3.89, could not, or only partially, reproduce the constants A to C on which the current  $M_w$  formula is based. For  $\log E_s(\text{GFZ})$  we got according to formula (3.139) and Fig.

3.89 for  $A = 4.51$ , for  $B = -4.6$  and for  $C = 1.51$ . When inserting these constants into the relationship (3.149) we get on the basis of  $\log E_s(\text{GFZ})$  data the new  $M_w$  formula

$$M_w = (\log M_0 - 9.11)/1.51. \quad (3.151)$$

This is very similar with the current standard formula, yielding  $M_w$  values that differ for  $\log M_0$  between about 17 and 23 (i.e.,  $M_w \approx 5.3$  to  $9.3$ ) only within  $-0.04$  and  $-0.07$  m.u.

In contrast, the constants based on the orthogonally regressed  $\log E_s(\text{USGS})$  data according to (3.140) and on the  $\Theta$  values in Fig. 3.89 are  $A = 2.41$ ,  $B = -4.8$ , and  $C = 1.79$ , yielding

$$M_w = (\log M_0 - 7.21)/1.79. \quad (3.152)$$

$M_w$  values calculated via (3.152) with  $\log M_0$  (respectively  $M_w$ ) values as above would differ from those derived with the current  $M_w$  standard formula between  $+0.2$  und  $-0.45$  m.u. Thus, the correctness of the current  $M_w$  formula is essentially confirmed by statistically correct treated GFZ  $E_s$  data, not, however, the respective USGS  $E_s$  data. Interestingly, however, the more heuristically derived relationship  $\log E_s = 1.5 M_s + 4.4$  by Choy and Boatwright (1995) would lead with an average  $\Theta_{\text{USGS}} = -4.8$  from Fig. 3.89, upper diagram, to

$$M_w = (\log M_0 - 9.2)/1.5. \quad (3.153)$$

This is very close to (3.150) and yields  $M_w$  values that are constantly smaller than those derived with the standard formula by only  $0.07$  m.u., i.e., it matches with the latter almost as good as the  $M_w$  formula (3.151) based on regressing  $\log E_s(\text{GFZ})$  over  $M_s(\text{USGS})$ . Thus, we can state, that the current  $M_w$  standard formula is well supported also by modern  $\log E_s$  data, better than the currently used  $M_e$  formula (see the discussion above following the Figs. 3.86 and 3.87).

One should be aware, however, that the large scatter of  $\Theta$  makes global relationships between  $\log E_s$  and  $\log M_0$  often unsuitable for drawing inferences on regional differences in tectonic deformation, stress accumulation and release rates. Furthermore, scaling laws for source parameters derived from low-frequency data, such as  $M_w$  or  $M_0$ , may - with the exception for assessing the tsunami hazard - not be suitable for a realistic assessment of the risk of shaking damage. The latter is mainly caused by frequencies above  $0.3$  Hz, which are, therefore, of greatest interest for earthquake engineers.

Global relations for calculating  $M_0$  via  $M_s$  were published by Ekström and Dziewonski (1988). They plotted more than 2.300 global NEIC PDE  $M_s$  values over Harvard CMT  $\log M_0$  values and calculated for three ranges of  $\log M_0$  the regression relations for  $M_s$  vs.  $\log M_0$ . They read, with  $M_0$  in Nm (**1 Nm = 1 J =  $10^7$  dyn cm =  $10^7$  ergs**) also in all other relations below):

$$M_s = \log M_0 - 12.24 \quad \text{for} \quad M_0 < 3.2 \times 10^{24} \quad (3.154a)$$

$$M_s = -12.24 + \log M_0 - 0.088(\log M_0 - 17.5)^2 \quad \text{for} \quad 3.2 \times 10^{24} \leq M_0 \leq 2.5 \times 10^{26} \quad (3.154b)$$

$$M_s = 0.667 \log M_0 - 6.06 \quad \text{for} \quad M_0 \geq 2.5 \times 10^{26} \quad (3.154c)$$

However, Ekström and Dziewonski (1988) resolved these SR1 relations for  $\log M_0$ , which read with  $M_0$  in Nm

$$\log M_0 = M_s + 12.24 \quad \text{for} \quad M_s < 5.3, \quad (3.155a)$$

$$\log M_0 = 23.20 - (92.45 - 11.40 M_s)^{1/2} \quad \text{for} \quad 5.3 \leq M_s \leq 6.8, \quad (3.155b)$$

$$\log M_0 = 1.5 M_s + 9.14 \quad \text{for} \quad M_s > 6.8. \quad (3.155c)$$

These inverse relations, however, are strictly speaking, not fully correct since the data scatter is significant. Therefore, Ambrasey (1990a) did not recommend their application for moment prediction from  $M_s$  values. Solving (3.155a-c) for  $M_0$  may systematically overestimate moments for larger events and underestimate  $M_0$  for small events. However, Purcaru and Berckhemer (1978 and 1982) derived already earlier for constant strain drop practically the same relationship as in (3.155c), namely  $\log M_0 = 16.1 + 1.5 M_s = 16.1 + 1.5 M_E$ , with  $M_E$  = strain energy magnitude. According to their data plot of  $\log M_0$  over  $M_s$  and  $M_E$ , respectively,  $M_0$  varies for fixed values of  $M_s$  or  $M_E$  within a factor of 6.

Chen and Chen (1989) published detailed global relations of  $M_0$  with  $M_s$ ,  $m_b$  and  $M_l$  in different ranges of magnitudes, respectively seismic moment. These relationships are based on data from about 800 earthquakes in the magnitude range  $0 < M < 8.6$ . These authors also showed that their empirical data are well fit by theoretical scaling relations derived from a modified Haskell model of a rectangular fault which produces displacement spectra with three corner frequencies. Similar global scaling relations had been derived earlier by Geller (1976), also based on the Haskell (1964 and 1966)  $\omega^{-3}$  model. In both papers these relations show saturation, in our opinion too early, for  $M_l$  at about 6.3, for  $m_b$  between about 6.0 and 6.5, and for  $M_s$  between about 8.2 and 8.5 (see introductory discussion in section 3.3.1).

Chen and Chen (1989) derived the following  $\log M_0$ - $M_s$  relationships with a standard deviation of individual values  $\log M_0$  of about  $\pm 0.4$ :

$$\log M_0 = 1.0 M_s + 12.2 \quad \text{for} \quad M_s \leq 6.4, \quad (3.156a)$$

$$\log M_0 = 1.5 M_s + 9.0 \quad \text{for} \quad 6.4 < M_s \leq 7.8, \quad (3.156b)$$

$$\log M_0 = 3.0 M_s - 2.7 \quad \text{for} \quad 7.8 < M_s \leq 8.5, \text{ and} \quad (3.156c)$$

$$\text{with } M_s = 8.5 = \text{const. for } \log M_0 > 22.8 \text{ Nm.} \quad (3.156d)$$

However,  $M_s$ - $M_0$  relations (and vice versa) vary regionally. According to Ambraseys (1990) the global relations (3.155a-c) systematically underestimate  $M_s$  for events in the Alpine region of Europe and adjacent areas by 0.2 magnitude units on average. And Abercrombie (1994) discussed possible reasons for the anomalous high surface-wave magnitudes of continental earthquakes relative to their seismic moment. This illustrates the need for regional scaling of moment-magnitude relationships even in the relatively long-period range.

For  $M_0$  and body-wave magnitudes  $m_b$  (when measured at 1s period) Chen and Chen (1989) give the following global scaling relations (with saturation at  $m_b = 6.5$  for  $\log M_0 > 20.7$ ). Note, however, that this applies only to old  $m_b$  data with amplitude measurement within the first few seconds. IASPEI (2005, 2011 and 2013) standard  $m_b$  saturates in fact only around near 7.5 (see Fig. 3.51 and IS 3.3):

$$\log M_0 = 1.5 m_b + 9.0 \quad \text{for} \quad 3.8 < m_b \leq 5.2, \quad (3.157a)$$

$$\log M_0 = 3 m_b + 1.2 \quad \text{for} \quad 5.2 < m_b \leq 6.5, \quad (3.157b)$$

For the  $M_0$ - $M_L$  relationship for California the same authors assume saturation at  $M_L = 6.3$  at  $\log M_0 > 20.1$  and they approximate the underlying non-linear relationship by three linear relationships in limited magnitude ranges:

$$\log M_0 = M_L + 10.5 \quad \text{for} \quad M_L \leq 3.6, \quad (3.158a)$$

$$\log M_0 = 1.5 M_L + 8.7 \quad \text{for} \quad 3.6 < M_L \leq 5.0, \quad (3.158b)$$

$$\log M_0 = 3 M_L + 1.2 \quad \text{for} \quad 5.0 < M_L \leq 6.3. \quad (3.158c)$$

Average scaling relations among  $m_b$ ,  $M_s$  and  $M_0$  for plate-margin earthquakes have been derived by Nuttli (1985). They yield practically identical values as the equations (3.156a-c) for  $M_0$  when  $M_s$  is known while the deviations are not larger than about a factor of 2 when using  $m_b$  and Eqs. (3.157a and b)) instead.

The need for regional relationships between  $M_0$  and magnitudes is particularly evident for  $M_L$ . When calculating  $M_0$  according to Eqs. (3.157b) and (3.158c) for California and comparing them with the values calculated for a relationship given by Kim et al. (1989) for the Baltic Shield

$$\log M_0 = 1.01 M_L + 9.93 \quad \text{for} \quad 2.0 \leq M_L \leq 5.2 \quad (3.159)$$

we get for  $M_L = 2.0, 4.0$  and  $5.0$ , respectively, values for  $M_0$  which are 3.5, 5.4 and 16.6 times larger for California than for the Baltic Shield. Using instead an even more local relationship for travel paths within the Great Basin of California (Chávez and Priestley, 1985), namely

$$\log M_0 = 1.2 M_L + 10.49 \quad \text{for} \quad 1 \leq M_L \leq 6 \quad (3.160)$$

we get for the same magnitudes even 9, 21 and 32 times larger values for  $M_0$  than for the Baltic Shield according to Eq. (3.159).

Other  $M_0$ - $M_L$  relationships are:

for Southern California (Thatcher & Hanks, 1973):

$$\log M_0 = 1.5 M_L + 9.05 \quad \text{in the range } 3 \leq M_L \leq 7, \quad (3.161)$$

for  $M_L$  from strong-motion records in Greece (Margris & Papazachos):

$$\log M_0 = 1.5 M_L + 9.07 \quad \text{in the range } 3.9 \leq M_L \leq 6.6, \quad (3.162)$$

for Israel (Shapira & Hofstetter, 1993):

$$\log M_0 = 1.3 M_L + 10.0 \quad \text{in the range } 2.2 \leq M_L \leq 5.3, \quad (3.163)$$



for Northwestern Turkey (Parolai et al., 2007):

$$\log M_0 = 1.17 M_l + 10.12 \quad \text{in the range } 0.5 < M_l < 6, \quad (3.164)$$

for the Italian earthquake catalog (Castello et al., 2007):

$$\log M_0 = 1.18 M_l + 10.92 \quad \text{in the range } 3.5 \leq M_l \leq 5.8, \quad (3.165)$$

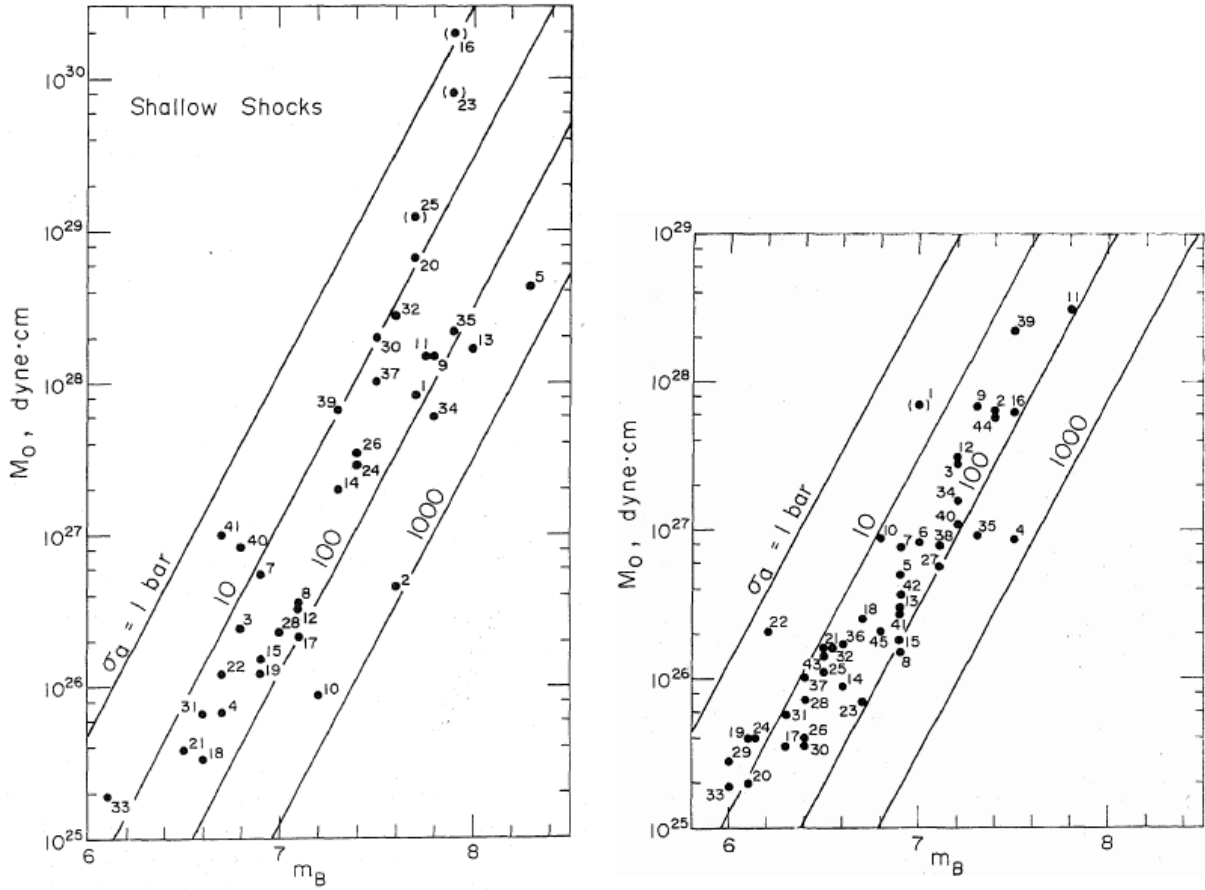
and for Northwestern Italy (Bindi et al., 2005):

$$\log M_0 = 0.95 M_l + 10.36 \quad \text{in the range } 0 < M_l < 4.5. \quad (3.166)$$

When comparing these various  $\log M_0$ - $M_l$  relationships we realize that the slope for data in the range  $0 < M_l < 7$  tends to increase with  $M_l$  from about 1 to 1.5. Accordingly, the slope for the respective  $M_w$ - $M_l$  relationship varies in this range from about 2/3 to 1 (see Tab. 3.6). The very steep slope in (3.151c), however, is mainly due to the assumption in the related scaling law calculations of a too early saturation of  $M_l$  at 6.3 already.

The slope change, which is in fact a gradual one, is mainly related to the magnitude and stress-drop dependent variability of the seismic source spectra and their corner frequencies in the range of magnitudes considered with respect to the passband of the Wood-Anderson seismometer, whereas the different constants are mainly controlled by regional differences in the attenuation laws (see, e.g., the difference in Tab. 3.6 between the slope 1  $M_w$ - $M_l$  relationships (or slope 1.5  $\log M_0$ - $M_l$  relationships) by Ristau et al. (2005) for crustal and slab earthquakes, respectively, with  $M_l \geq 3.6$  but  $< 6$  in Western Canada). Therefore, in order to avoid physically or regionally wrong conclusions to be drawn from slope differences of various  $\log M_0$  (or  $M_w$ )- $M_l$  relationships it is absolutely necessary to take the magnitude-range of their applicability into account.

While there are plenty relationships for converting  $M_l$ ,  $m_b(\text{old})$  and  $M_s(20)$  data into  $\log M_0$  (or  $M_w$ ), relationships between  $\log M_0$  (respectively  $M_w$ ) and  $m_B$  are very rare. Exceptions are the data plots by Abe (1982) of  $\log M_0$  over classical  $m_B$  for both shallow and deep earthquakes (see Fig. 3.90).



**Fig. 3.90**  $\log M_0$  (in units of  $\text{dyne cm} = 10^{-7} \text{ N m}$ ) versus  $m_B$  for shallow (left diagram) and for intermediate-depth to deep earthquakes (right diagram) according to Figures 1 and 3 of Abe (1982). The apparent stress  $\sigma_a = \mu(E_S/M_0)$  is given in units of  $\text{bar} = 0.1 \text{ MPa}$ . Note that for calculating  $M_0$  and  $C$  for intermediate-depth to deep earthquakes a three times higher rigidity (shear modulus)  $\mu$  of the source medium ( $\mu = 9 \times 10^{11} \text{ dyn/cm}^2$ ) than for shallow earthquakes has been assumed by Abe.

The lines in Fig. 3.90 have been calculated for increment values of  $\sigma_a$  with the relationship

$$\log M_0 = 2.4 m_B - 1.2 - \log(\sigma_a/\mu), \text{ with } M_0 \text{ in N m.} \quad (3.167)$$

This relation results from inserting into the Gutenberg-Richter  $\log E_S$ - $m_B$  relationship (3.134)  $E_S = M_0 \times (\sigma_a/\mu)$  and resolving it for  $\log M_0$ . A line for  $\sigma_a \approx 40 \text{ bar} = 4 \text{ MPa}$ , and thus for a stress drop  $\Delta\sigma \approx 8 \text{ MPa}$ , would be a reasonably good average fit of the data in Fig. 3.90 and correspond for shallow earthquakes with  $\mu = 3 \times 10^{11} \text{ dyn/cm}^2 = 3 \times 10^4 \text{ N m}$  to a relationship

$$\log M_0 = 2.4 m_B + 2.68 \quad (3.168)$$

However, when calculating for shallow earthquakes the linear OSR standard regression through some 974 data pair of GCMT  $\log M_0$  vs.  $m_B(\text{GFZ}) = m_B\text{_{BB}}$ , with the latter measured according to the procedure described by Bormann and Saul (2008), one gets

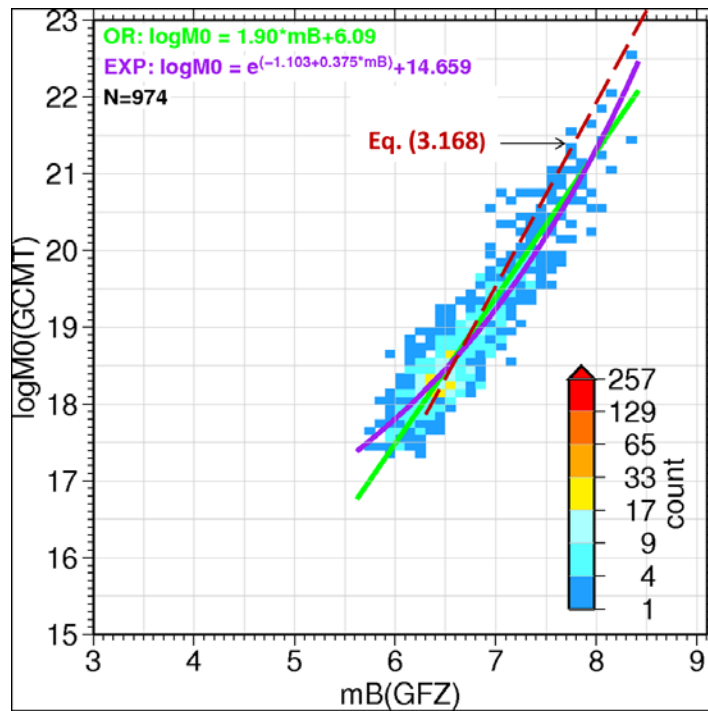
$$\log M_0 = 1.90 m_B\text{_{BB}} + 6.09, \quad (3.169)$$

and when fitting the same data with a non-linear regression relationship

$$\log M_0 = e^{(-1.103 + 0.375 \text{ mB}_{\text{BB}})} + 14.659. \quad (3.170)$$

All three relationships have been plotted together with the GCMT-GFZ data in Fig. 3.91. One realizes that within the range of the definition by data

- Eq. (3.170) approximates best the relationship between IASPEI standard  $\text{mB}_{\text{BB}}$  and GCMT  $\log M_0$ ;
- all three relationships yield identical average  $\log M_0$  proxy estimates at  $\text{mB} = 6.7$ ;
- other  $\log M_0$  proxy estimates via  $\text{mB}$  differ not more than about 0.5 units from directly measured  $\log M_0$ .



**Fig. 3.91**  $\text{mB}(\text{GFZ}) - \log M_0$  (GCMT) regression models: The exponential model [(Eq. (3.170))] is drawn in purple, the linear OSR model [(Eq. (3.169))] in green and Eq. (3.168), derived by us on the basis of early  $\log M_0$  and classical  $\text{mB}$  data by Abe (1982), as red broken line. The data distribution is shown as color-coded count cells,  $0.1 \times 0.1$  units wide.

Finally, we present in Figs. 3.92 a and b the most recent non-linear regression relationships based on 30,709  $\log M_0$ - $\text{mb}$  as well as 17,472  $\log M_0$ - $M_s$  data of the ISC-GEM Global Instrumental Earthquake Catalog (1900 – 2009). According to Di Giacomo et al. (2013) the respective OSR relationships are

$$\log M_0 = 1.95 \text{ mb} + 7.03, \quad (3.171)$$

$$\log M_0 = 1.00 M_s + 12.30 \quad \text{for } M_s < 6.47, \quad (3.172)$$

and

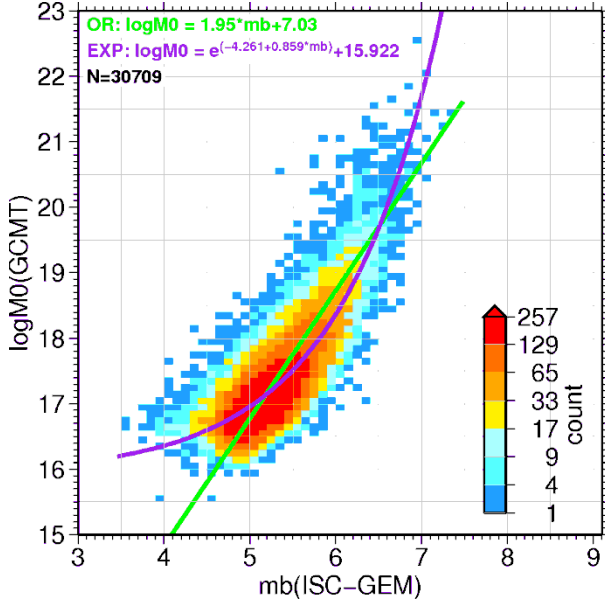
$$\log M_0 = 1.66 M_s + 7.97 \quad \text{for } M_s > 6.47, \quad (3.173)$$

and the corresponding exponential least-square relationships read

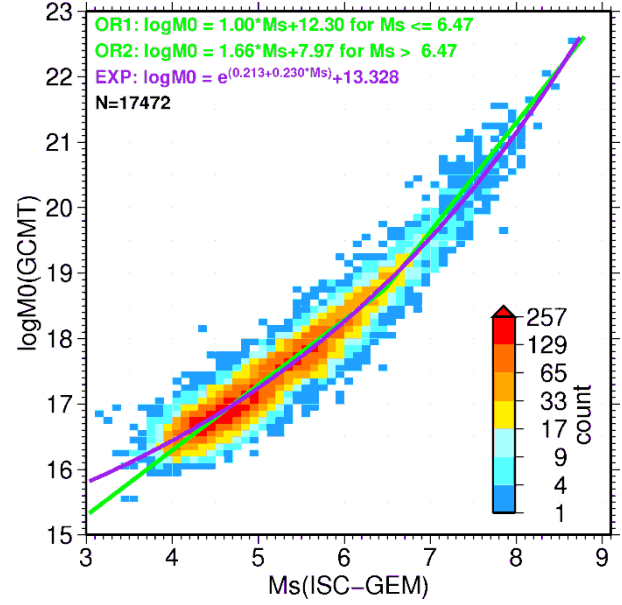
$$\log M_0 = e^{(-4,261 + 0.859 \text{ mb})} + 15.922, \quad (3.174)$$

and

$$\log M_0 = e^{(0,213 + 0.230 M_s)} + 13.28. \quad (3.175)$$



**Fig. 3.92a** mb (ISC - GEM) – log  $M_0$  (GCMT) regression models: The exponential model is drawn in purple and the linear OSR model in green. The data distribution is shown as color-coded count cells  $0.1 \times 0.1$  wide.



**Fig. 3.92b**  $M_s$  (ISC - GEM) – log  $M_0$  (GCMT) regression models: the exponential model is drawn in purple and the bi-linear OSR model in green. The data distribution is shown as color-coded count cells  $0.1 \times 0.1$  wide.

The OSR relationships between log  $M_0$  and  $M_s$ , based on the newest and most homogeneous ISC-GEM Global Instrumental Earthquake Catalogue data set (1900 – 2009) almost perfectly confirms the respective relationships published earlier by Chen and Chen (1989) for  $M_s < 6.4$  [see Eqs. (3.156a)] and of Ekström and Dziewonski (1988) for  $M_s < 5.3$  [Eq. (3.155a)]. The slope of 1.66 in (3.173), however, is larger than the slope of 1.5 in both the Ekström and Dziewonski relationship (3.155c) for  $M_s > 6.8$  and in the Chen and Chen relationship (3.156b) for  $6.4 \leq M_s < 7.8$ . Remember that Kanamori (1977), in his effort to estimate the seismic energy released by strong earthquakes and to develop a non-saturating magnitude scale, calibrated  $M_0$  to  $M_s$  via the Richter (1958)  $\log E_s$ - $M_s$  relationship, knowing about the close relationship between log  $M_0$  and  $M_s$  in the range  $6.5 < M_s < 8$  and assuming a constant average  $E_s/M_0$  ratio (see section 3.2.7.1). Having now plenty of direct  $M_0$  measurements available one could scale them directly to  $M_s$  without the detour over an assumed average  $E_s/M_0$  ratio. This was done by Choy and Boatwright (1995) when developing the energy magnitude scale  $M_e$  by relating directly measured  $E_s$  to  $M_s$  (see section 3.2.7.2). Accordingly, by substituting in (3.173)  $M_s$  by  $M_w$  and resolving it for  $M_w$  one gets

$$M_w = (\log M_0 - 7.97)/1.66. \quad (3.176)$$

When compared with the Kanamori (1977) based IASPEI standard  $M_w$  scale formula (3.68),  $M_w = (\log M_0 - 9.1)/1.5$ , (3.176) would yield for  $\log M_0 = 16$   $M_w = 4.84$  and for  $\log M_0 = 23$

$M_w = 9.05$ , instead of  $M_w = 4.60$  and  $9.27$ , respectively, when using (3.68). Thus the difference between  $M_w$  values calculated with the established and the new formula would be within this range of  $\log M_0$ , approximately corresponding to  $4.5 < M_w < 9.3$ , generally less than 0.25 m.u. This is still a reasonably good agreement but none the less worth considering a change to an  $M_w$  scale that is directly scaled to  $M_s$  for strong earthquakes and not via a  $\log E_S$ - $M_s$  relationship with a constant slope of 1.5 and an assumed constant  $E_S/M_0$  ratio.

The latter is theoretically required when the omega-squared similarity model (as in Fig. 3.5) holds on average. Deviations are then only related to differences in stress drop. Data of Abercrombie (1995) and of the TERRAscope TriNet (Kanamori 2001), however, seemed to indicate that average  $E_S/M_0$  values drop for  $M_w < 5$  below  $10^{-4}$  and even below  $10^{-6}$  for  $M_w < 1$ . However, Oth et al. (2010), analyzing masses of Japanese KiK-net data in the magnitude range  $2.7 \leq M_{JMA} \leq 8$ , could not confirm such an average scale dependence of  $E_S/M_0$  within this range, despite the large individual scatter between about  $4 \times 10^{-4}$  and  $8 \times 10^{-7}$ , which these authors attribute solely to variations in stress drop. Whether this similarity holds down to  $M_w = -4.5$  (Kwiatek et al., 2010), however, and thus the applicability of the standard  $M_w$  formula (3.68) for converting  $M_0$  into moment magnitudes even 11 magnitude units outside the range of its original definition based on  $M_s$  data, seems to be more than questionable.

### 3.3.4 Scaling relations of magnitudes, $M_0$ and $E_S$ with fault parameters (P. Bormann)

Scaling relations of magnitude, seismic moment and energy with fault parameters are used in two ways:

- 1) to get a rough estimate of relevant fault parameter when  $M$ ,  $M_0$  or  $E_S$  of the event are known from the evaluation of instrumental recordings; or
- 2) in order to get a magnitude, moment and/or energy estimates for historic or even prehistoric events for which no recordings are available but for which some fault parameters such as the length of surface rupture and/or amount of surface displacement can still be determined from field evidence.

The latter is particularly important for improved assessment of seismic hazard and for estimating the maximum possible earthquake, especially in areas with long mean recurrence times for strong seismic events. Of particular importance for hazard assessment are also relationships between macroseismic intensity,  $I$ , and magnitude,  $M$ , on the one hand (see Eqs. in 3.2.6.6) and between ground acceleration and  $I$  or  $M$ , on the other hand. Unfortunately, the measured maximum accelerations for equal values of intensity  $I$  scatter in the whole range of  $I = III$  to  $IX$  by about two orders of magnitude (Ambraseys, 1975). The reason for this scatter is many-fold, e.g., human perception is strongest for frequencies around 3 Hz while acceleration and damage might be strongest for more high frequent ground motions. Also, damage depends not only on the peak value of acceleration (PGA) but also depends on its frequency with respect to the natural period of the shaken structures and on the duration of strong ground shaking. Moreover, especially in real-time applications for damage prediction and assessment, peak ground velocity (PGV) data have proven to be much more important than PGA data, especially for mid-rise and high-rise buildings that are typical of modern societies (e.g., Wu et al., 2003). And some structures are particularly vulnerable to large ground displacements and not so much to high accelerations, e.g., pipelines or very long free-spanning bridges.

Relationships between  $M_0$ ,  $M_s$ , and  $E_s$  with various fault parameters are mostly based on model assumptions on the fault geometry, rupture velocity and time history, ambient stress or stress drop, etc. But sometimes these fault parameters can, at least partially, be confirmed or constrained by field evidence or by petrophysical laboratory experiments. As for other scaling relations discussed above, global relationships can give only a rough orientation since the scatter of data is considerable due to regional variability. Whenever possible, regional relationships should be developed.

Sadovsky et al. (1986) found that for both crustal earthquakes and underground explosions the following relationship holds between seismic energy  $E_s$  (in erg) and the seismic source volume  $V_{\text{source}}$  (in  $\text{cm}^3$ ):

$$\log E_s = 3 + \log V_{\text{source}} \quad (3.177)$$

with  $V_{\text{source}}$  for earthquakes being estimated from the linear dimensions of the aftershock zone. This means that the critical energy density for both natural and artificial crustal seismic sources is about equal, roughly  $10^3 \text{ erg/cm}^3$  or  $100 \text{ J/m}^3$ . It does not depend on the energy released by the event.  $E_s$  increases only because of the volume increase of the source. Accordingly, it is not the type of seismic source but the properties of the medium that play the decisive role in the formation of the seismic wave field. However, local and regional differences in ambient stress and related stress drop  $\Delta\sigma \approx 2\mu E_s/M_0$  may modify this conclusion.

Fig. 3.93 shows the relation between seismic moment  $M_0$  and the area  $A_r$  of fault rupture as published by Kanamori and Anderson (1975).  $A_r$  is controlled by the stress drop  $\Delta\sigma$ ; as  $\Delta\sigma$  increases for a given rupture area,  $M_0$  becomes larger. One recognizes that intraplate earthquakes have on average a higher stress drop (around 10 MPa = 100 bars) than interplate events (around 3 MPa). The data in Fig. 3.93 are also well fit by the average relationship suggested by Abe (1975), namely:

$$M_0 = 1.33 \times 10^{15} A_r^{3/2} \quad (3.178)$$

which is nearly identical with the relation by Purcaru and Berckhemer (1982):

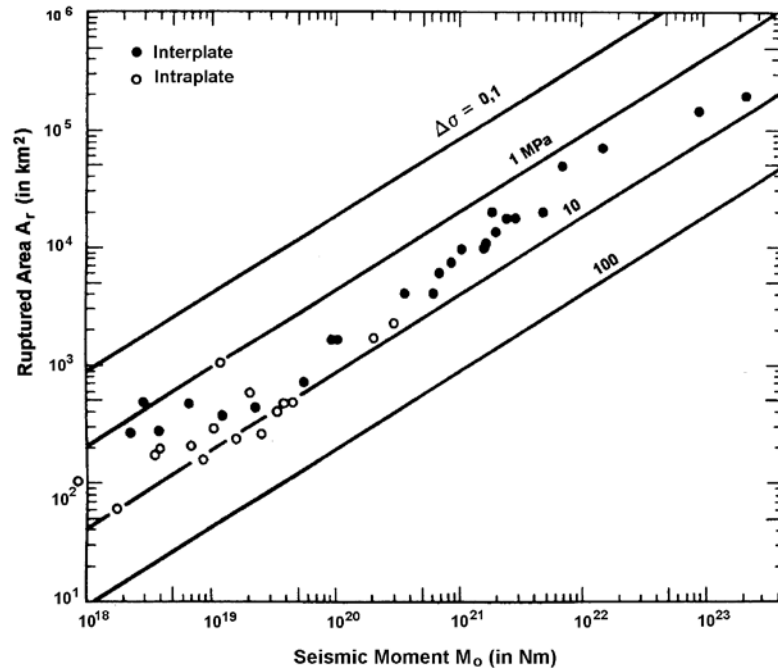
$$\log M_0 = (1.5 \pm 0.02) \log A_r + (15.25 \pm 0.05) \quad (3.179)$$

with  $M_0$  in Nm and  $A_r$  in  $\text{km}^2$ . Eq. (3.179) corresponds to the theoretical scaling relation derived by Chen and Chen (1989) for a modified Haskell model with the assumption  $L = 2W$  ( $L$  - length and  $W$  - width of fault rupture,  $A_r = LW = 0.5 L^2$ ) and an average displacement  $\bar{D} = 4.0 \times 10^{-5} L$ . Note that empirical data indicate also other aspect ratios  $L/W$  up to about 30 (e.g., Purcaru and Berckhemer, 1982).

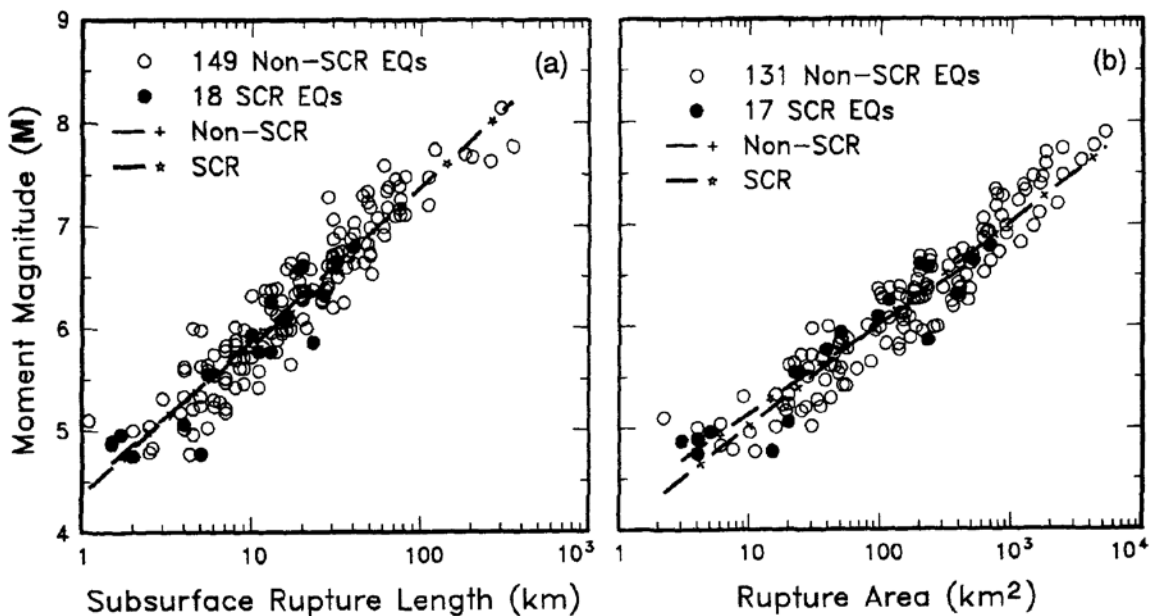
Wells and Coppersmith (1994) gave another relation between moment magnitude and  $A_r$ , derived from a very comprehensive data base of source parameters for historical shallow-focus earthquakes ( $h < 40 \text{ km}$ ) in continental interplate or intraplate environments:

$$M_w = (0.98 \pm 0.03) \log A_r + (4.07 \pm 0.06) \quad (3.180)$$

For the related data see Fig. 3.95b.



**Fig. 3.94** Relation between area of fault rupture  $A_r$  and seismic moment  $M_0$  for inter- and intraplate earthquakes. The solid lines give the respective relationships for different stress drop  $\Delta\sigma$  (in MPa;  $1 \text{ Pa} = 10^{-5} \text{ bars}$ ) (modified from Kanamori and Anderson, Theoretical basis of some empirical relations in seismology, Bull. Seism. Soc. Am., Vol. 65, p. 1077, Fig. 2, 1975; © Seismological Society of America).

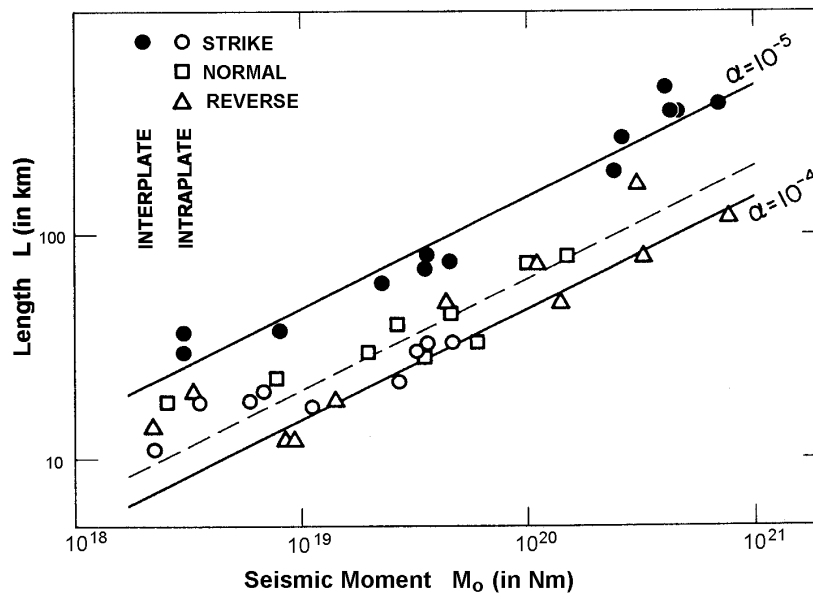


**Fig. 3.95** Data plots and regression lines for the relationship between moment magnitude and surface rupture length (left panel), respectively rupture area (right panel), for earthquakes in stable continental regions (SCR) and non-SCR according to Wells and Coppersmith (1994), Bull. Seism. Soc. Am., 54(4), Fig. 17. © Seismological Society of America.

There also exist linear log-log relationships between  $L$  and  $M_0$ . One of such relations, published by Wells and Coppersmith (1994), has been plotted in the left diagram of Fig. 3.95 for earthquakes in stable continental regions (SCR) and non-stable ones. For another average relationship between  $M_w$  and the surface rupture length SRL of strike slip, reverse and normal faulting earthquakes (77 events) these authors give:

$$M_w = 1.16 \log \text{SRL}(\text{km}) + 5.08. \quad (3.181)$$

According to Scholz et al. (1986),  $L$  is for a given seismic moment on average about 6 times larger for interplate (strike-slip) events than for intraplate ones (see Fig. 3.96) and the ratio  $\alpha$  between average fault displacement (slip)  $\bar{D}$  and fault length  $L$ ,  $\alpha \approx 1 \times 10^{-5}$  for interplate and  $\alpha \approx 6 \times 10^{-5}$  for intraplate events. Since this result is independent of the type of fault mechanism, it implies that intraplate faults have a higher frictional strength (and thus stress drop) than plate boundary faults but smaller length for the same seismic moment.



**Fig. 3.96** Fault length  $L$  versus seismic moment  $M_0$  for large inter- and intraplate earthquakes. The solid lines give the respective relationship for the ratio  $\alpha = \bar{D}/L$  (modified from Scholz et al. (1986), Scaling differences between large interplate and intraplate earthquakes, Bull. Seism. Soc. Am., Vol. 76, No. 1, p. 68, Fig. 1; © Seismological Society of America).

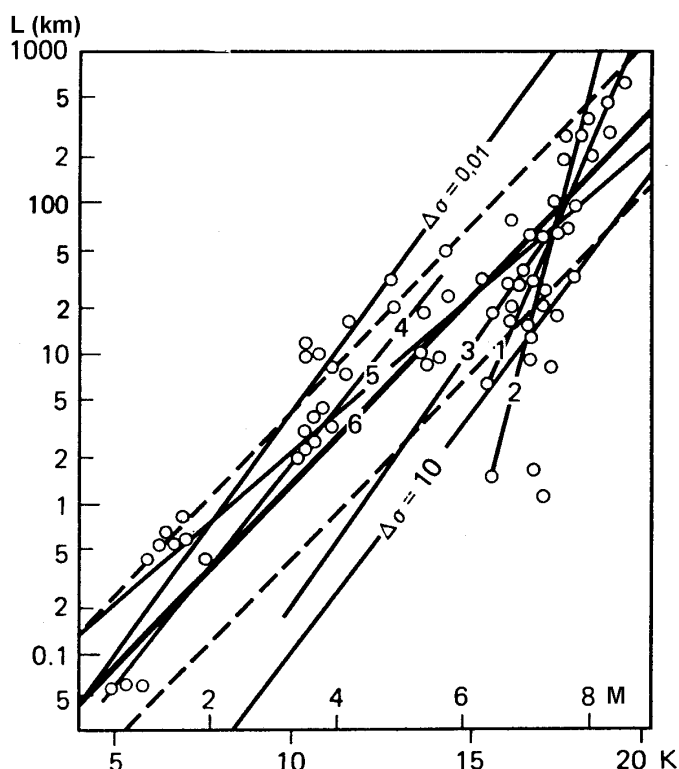
The slope of the curves in Fig. 3.96 is 0.5. This corresponds to a relation  $M_0 \sim L^2$  (Scholz 1982; Pegeler and Das, 1996) which is only valid for large earthquakes ( $M >$  about 6.5 to 7). Then the width  $W$  of the fault is already saturated, i.e., has reached its largest extension within the seismogenic zone of brittle fracturing in the crust, respectively lithosphere. Depending on heat flow and composition, the seismogenic zone in the crust is about 10 to 30 km thick. Accordingly, for larger earthquakes, the growth of the fault area with increasing  $M_0$  is in the length direction only.

Recently, there has been some serious debate on the scaling of large earthquakes and their ratio  $\alpha$  (Scholz, 1994a and b and 1997; Romanowicz 1994; Romanowicz and Rundle, 1993 and 1994; Sornette and Sornette, 1994; Wang and Ou, 1998). Romanowicz (1992), who prefers to scale slip not with length but with width, even gives a relationship of  $M_0 \sim L$  in case



of very large earthquakes. In contrast, Hanks (1977) showed that earthquakes with rupture dimensions smaller than the thickness of the seismogenic layer scale according to  $M_0 \sim L^3$  which is equivalent to Eq. (3.177).

Accordingly, in agreement also with a data compilation by Riznichenko (1992), shown in Fig. 3.97, often rather different correlation relationships between source length  $L$ , magnitude  $M$  and energetic class  $K$  have been published by various authors for events in different seismotectonic environments.



**Fig. 3.97** Correlation of source length  $L$  with magnitude  $M$  and energetic class  $K$  according to data from various sources (e.g., curve 1 by Tocher, 1958, curve 2 by Iida, 1959; curve 6 average by Riznichenko, 1992; Russian original 1985). Thin straight lines: related stress drops  $\Delta\sigma$  are given in MPa; broken lines mark the limits of the 68% confidence interval with respect to the average curve 6 (modified from Riznichenko, 1992, Fig. 3; with permission of Springer-Verlag).

Ambraseys (1988) published relationships derived from the dimensions of fault surface ruptures for Eastern Mediterranean and Middle Eastern earthquakes (with  $L$  - observed surface rupture length in km,  $\bar{D}$  - relative fault displacement in cm,  $M_{SC}$  - predicted surface-wave magnitudes):

$$M_{SC} = 1.43 \log L + 4.63 \quad (3.182)$$

and

$$M_{SC} = 0.4 \log (L^{1.58} \bar{D}^2) + 1.1. \quad (3.183)$$

They yield results which are in good agreement with those by Nowroozi (1985) for Iran but they differ significantly from the respective relations given by Tocher (1958) for Western USA and from Iida (1959) for Japan (see curves 1 and 2 in Fig. 3.97).

Khromovskikh (1989) analyzed available data for more than 100 events of different faulting types from different seismotectonic regions of the Earth. He derived 7 different relationships between magnitude  $M$  and the length  $L$  of the rupture zone, amongst them those for the following regions:

a) the Circum-Pacific belt:  $M = (0.96 \pm 0.25) \log L + (5.70 \pm 0.34)$  (3.184)

b) the Alpine fold belt:  $M = (1.09 \pm 0.28) \log L + (5.39 \pm 0.42)$  (3.185)

c) rejuvenated platforms:  $M = (1.25 \pm 0.19) \log L + (5.45 \pm 0.28)$  (3.186)

and compared them with respective relationships of other authors for similar areas.

Other relationships for estimating  $L$  (in km), when  $M_s$  is known, were derived by Chen and Chen (1989) on the basis of their general scaling law using a modified Haskell model for a rectangular fault. These relationships clearly show the effect of width saturation:

$$\log L = M_s/3 - 0.873 \quad \text{for} \quad M_s \leq 6.4 \quad (3.187)$$

$$\log L = M_s/2 - 1.94 \quad \text{for} \quad 6.4 < M_s \leq 7.8 \quad (3.188)$$

$$\log L = M_s - 5.84 \quad \text{for} \quad 7.8 < M_s \leq 8.5. \quad (3.189)$$

The same authors also gave similar relations between the average dislocation  $\bar{D}$  (in m) and  $M_s$ , namely:

$$\log \bar{D} = M_s/3 - 2.271 \quad \text{for} \quad M_s \leq 6.4 \quad (3.190)$$

$$\log \bar{D} = M_s/2 - 3.34 \quad \text{for} \quad 6.4 < M_s \leq 7.8 \quad \text{and} \quad (3.191)$$

$$\log \bar{D} = M_s - 7.24 \quad \text{for} \quad 7.8 < M_s \leq 8.5 \quad (3.192)$$

while Chinnery (1969) derived from still sparse empirical data a linear relation between magnitude  $M$  and  $\log \bar{D}$  (with  $\bar{D}$  in m) for the whole range  $3 < M < 8.5$

$$M = 1.32 \log \bar{D} + 6.27 \quad (3.193)$$

which changes to

$$M = 1.04 \log \bar{D} + 6.96 \quad (3.194)$$

when only large magnitude events are considered.

Probably best established are the relations which Wells and Coppersmith (1994) have determined for shallow-focus (crustal) continental interplate or intraplate earthquakes on the basis of a rather comprehensive data base of historical events. Since most of these relations for strike-slip, reverse and normal faulting events were not statistically different (at a 95% level of significance) their average relations for all slip types are considered to be appropriate

for most applications. Best established are the relationships between moment magnitude  $M_w$  and rupture area [see Eq. (3.174 and Fig. 3.95 right)], surface rupture length (SRL) (see Fig. 3.95 left) and subsurface rupture length (RLD) (both in km). They have the strongest correlations ( $R_{XY} = 0.89$  to  $0.95$ ) and the least data scatter:

$$M_w = (1.16 \pm 0.07) \log (\text{SRL}) + (5.08 \pm 0.10) \quad (3.195)$$

$$M_w = (1.49 \pm 0.04) \log (\text{RLD}) + (4.38 \pm 0.06) \quad (3.196)$$

$$\log (\text{SLR}) = (0.69 \pm 0.04) M_w - (3.22 \pm 0.27) \quad (3.197)$$

$$\log (\text{RLD}) = (0.59 \pm 0.02) M_w - (2.44 \pm 0.11) \quad (3.198)$$

When comparing Eqs. (3.197) and (3.198) it follows that in general the observed surface rupture length is only about 75% of the subsurface rupture length inferred from modeling.

The correlations between  $M_w$  and  $\bar{D}$  as well as  $\bar{D}$  and SLR are somewhat smaller ( $R_{XY} = 0.71$  to  $0.78$ ):

$$M_w = (0.82 \pm 0.10) \log \bar{D} + (6.693 \pm 0.05) \quad (3.199)$$

$$\log \bar{D} = (0.69 \pm 0.08) M_w - (4.80 \pm 0.57) \quad (3.200)$$

$$\log \bar{D} = (0.88 \pm 0.11) \log (\text{SLR}) - (1.43 \pm 0.18) \quad (3.201)$$

$$\log (\text{SLR}) = (0.57 \pm 0.07) \log \bar{D} + (1.61 \pm 0.04). \quad (3.202)$$

Wells and Coppersmith (1994) reasoned that the weaker correlation may reflect the wide range of displacement values for a given rupture length (differences up to a factor 50 in their data set!). These authors also give relations between SLR and the maximum surface displacement which is, on average, twice the observed average surface displacement while the average subsurface slip ranges between the maximum and average surface displacement.

Chen and Chen (1989) also derived from their scaling law the following average values:

- rupture velocity  $v_r = 2.65$  km/s;
- total rupture time  $T_r$  (in s) =  $0.35$  (s/km)  $\times$   $L$  (km); (3.203)
- slip velocity  $dD/dt$  varying between  $2.87$  and  $11.43$  m/s.

However,  $v_r$  and  $dD/dt$  usually vary along the fault during the fracture process. From teleseismic studies we obtain usually only spatially and temporally averaged point source values of fault motion but the actual co-seismic slip is largely controlled by spatial heterogeneities along the fault rupture (see Figs. 3.9-3.12). Large slip velocities over  $10$  m/s suggest very high local stress drop of more than  $10$  MPa (Yomogida and Nakata, 1994). On the other hand, sometimes very slow earthquakes may occur with very large seismic moment but low seismic energy radiation (e.g., "tsunami earthquakes"; Kanamori, 1972; Polet and

Kanamori, 2009). This has special relevance when deriving scaling relations suitable for the prediction of strong ground motions (e.g., Fukushima, 1996).

Scaling relationships between fault parameters, especially between  $D$  and  $L$ , are also controlled by the fault growth history, by age and by whether the event can be considered to be single and rare or composite and frequent (e.g., Dawers et al., 1993; Tumarkin et al., 1994). There exist also scaling relations between fault length and recurrence interval which are of particular relevance for seismic hazard assessment (e.g., Marrett, 1994).

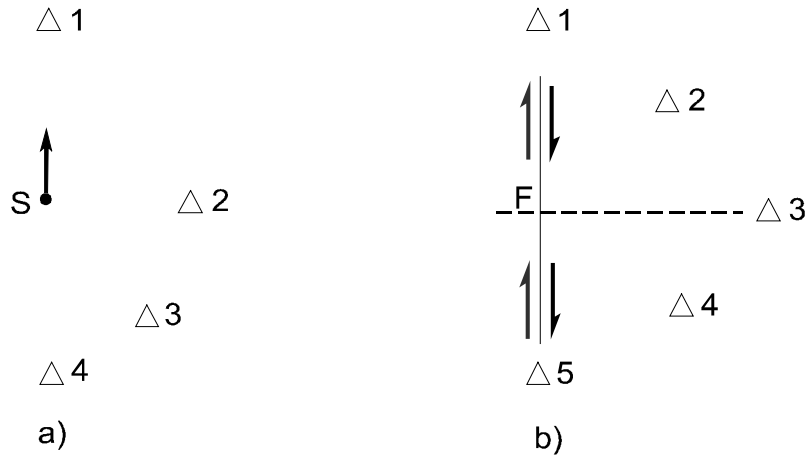
Using Eqs. (3.182), (3.184)-(3.186) and (3.195), one gets for a surface rupture length of 100 km magnitudes  $M = 7.5, 7.7, 7.6, 7.95$  and  $7.4$ , respectively. When knowing the  $M_s$  or  $M_w$  and calculating  $L$  and  $\bar{D}$  according to Eqs. (3.187-3.188), (3.189-3.190), (3.197) and (3.200), one gets for magnitude 7.0  $L = 36$  km and 41 km,  $\bar{D} = 1.4$  m and 1.1 m and for magnitude 8.0  $L = 145$  km and 200 km,  $\bar{D} = 3.8$  m and 5.2 m. The good agreement of the calculated values for magnitudes 7 and the stronger disagreement for magnitudes 8 are obviously due to the growing difference between  $M_s$  (used in the relations by Chen and Chen, 1989) and  $M_w$  (used in the relations by Wells and Coppersmith, 1994) for  $M_s > 7$  (saturation effect). For the rupture duration we get according to Eq. (3.203) for  $M_s = 7$  and 8 and the respective rupture lengths (3.188) and (3.189) approximately 13 s and 51 s, respectively. If we use, however, the empirically derived average rupture duration relation (3.3) according to Bormann and Saul (2009a) the average rupture duration of earthquakes with  $M = 7$  and 8 would be about 25 s and 100 s, respectively.

### 3.4 Determination of fault-plane solutions (P. Bormann and S. Wendt)

#### 3.4.1 Introduction

The direction (polarity) and amplitude of motion of a seismic wave arriving at a distant station depends both on the wave type considered and the position of the station relative to the motion in the earthquake source. This is illustrated by Figs. 3.98a and b.

Fig. 3.98a represents a linear displacement of a point source  $S$  while Fig. 3.98b depicts a right lateral (dextral) shear dislocation along a fault plane  $F$ . Shear dislocations are the most common model to explain earthquake fault ruptures. Note that in the discussion below we consider the source to be a point source with rupture dimension much smaller than the distance to the stations and also much smaller than the wave length considered. First we look into the situation depicted in Fig. 3.98a. When  $S$  moves towards  $\Delta 1$  then this station will observe a *compressional* (+) P-wave arrival (i.e., the first motion is *away* from  $S$ ),  $\Delta 4$  will record a P wave of opposite sign (-), a *dilatation* (i.e., first motion *towards*  $S$ ), and station  $\Delta 2$  will receive no P wave at all. On the other hand, S waves, which are polarized parallel to the displacement of  $S$  and perpendicular to the direction of wave propagation, will be recorded at  $\Delta 2$  but not at  $\Delta 1$  and  $\Delta 4$  while station  $\Delta 3$  will receive both P and S waves.



**Fig. 3.98** Direction of source displacement with respect to seismic stations  $\Delta_i$  for a) a single force at point S and b) a fault rupture F. Note that in the discussion below we consider the source to be a point source with a rupture dimension much smaller than the distance to the stations (Figure by courtesy of M. Baumbach †, 2001).

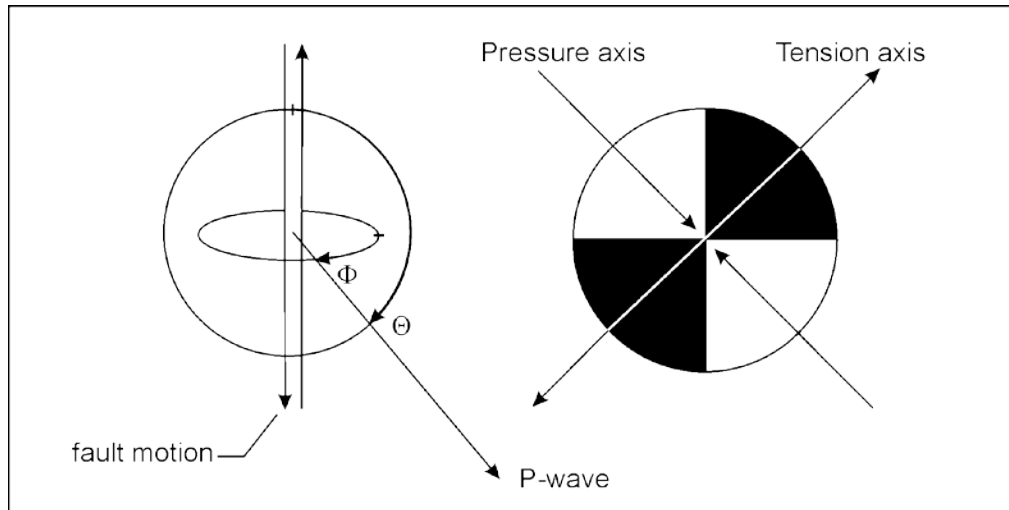
Somewhat different is the case of a fault rupture (Fig. 3.98b). At stations  $\Delta_1$  and  $\Delta_5$ , which are positioned in the strike direction of the fault, the opposite signs of P motion on both side of the fault will cancel, i.e., no P waves will be observed. The latter also applies for station  $\Delta_3$  which is sited perpendicular to the fault. On the other hand, stations  $\Delta_2$  and  $\Delta_4$ , which are positioned at an angle of  $45^\circ$  with respect to the fault, will record the P-wave motions with maximum amplitudes but opposite sign. This becomes clear also from Fig. 3.100a. It shows the different polarities and the amplitude "lobes" in the four quadrants. The length of the displacement arrows is proportional to the P-wave amplitudes observed in different directions from the fault. Accordingly, by observing the sense of first motions of P waves at many stations at different azimuths with respect to the source it will be possible to deduce a "fault-plane solution". But because of the symmetry of the first-motion patterns, two potential rupture planes, perpendicular to each other, can be constructed. Thus, on the basis of polarity data alone, an ambiguity will remain as to which one was the acting fault plane. This can only be decided either by field evidence on the orientation and nature of seismotectonic faults (see, e.g., Figure 7 and discussion in EX 3.3) or by taking into account additional data on azimuthal amplitude and frequency, respectively wave-form patterns (see Glossary: *directivity*, *directivity focusing*, *directivity pulse*). The latter are due to *Doppler* and constructive wave interference effects caused by the moving and waves radiating source.

In accordance with the above, the amplitude distribution of P waves,  $A_P$ , for a point source with pure double-couple shear mechanism is described in a spherical co-ordinate system  $(\theta, \phi)$  (Aki and Richards, 1980; see Fig. 3.99) by

$$A_P(\theta, \phi) = \cos \phi \sin 2\theta. \quad (3.204)$$

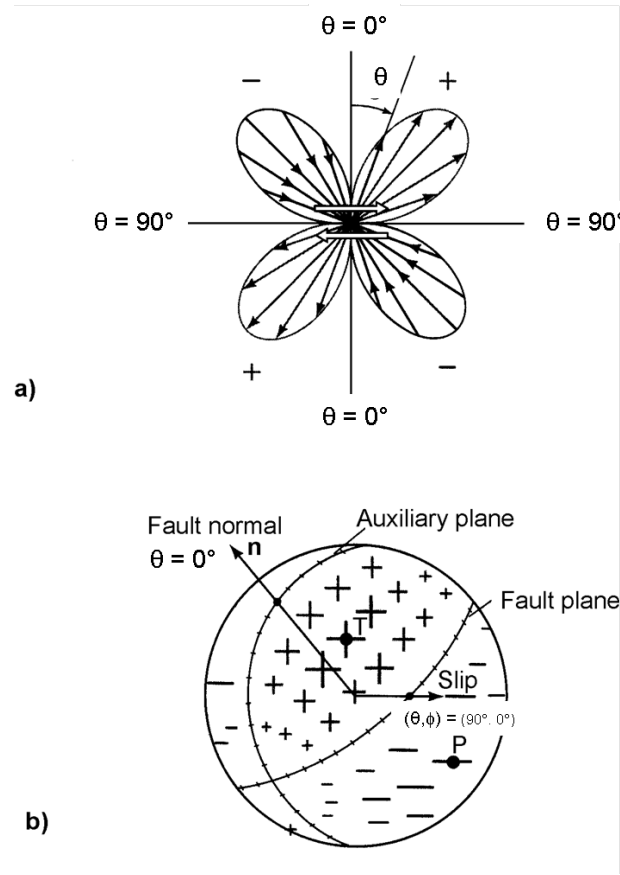
This expression divides the focal sphere into four quadrants. The focal sphere for a seismic point source is conceived of as a sphere of arbitrarily small radius centered on the source. Within each quadrant the sign of the P-wave first motion (polarity) does not change but amplitudes are large in the center of the quadrant and small (or zero) near to (or at) the fault plane and the auxiliary plane, which is perpendicular to the fault plane. The nodal lines for P

waves, on which  $A_p(\theta, \phi) = \cos \phi \sin 2\theta = 0$ , separate the quadrants. They coincide with the horizontal projection of the two orthogonal fault planes traces through the focal sphere. Opposite quadrants have the same polarity, neighboring quadrants different polarities. Note that *compressional first-motion* signals are observed at stations falling into the *tension quadrant* of the source (with tension force directed away from the point source) while *dilatational first-motion signals* are observed at stations falling into the respective *compression quadrant* (with pressure force directed towards the point source).



**Fig. 3.99** Map view of P-wave radiation pattern for a shear fault.  $\theta$  is the azimuth in the plane while  $\phi$  is in fact three-dimensional. See also Fig. 3.98b. Black areas: polarity +, i.e., **upward** P-wave first motion in vertical component records; white areas: -, i.e., **downward** P-wave first motion in vertical component records (modified according to a Figure kindly provided by M. Baumbach †, 2001).

The position of the quadrants on the focal sphere depends on the orientation of the active fault and of the slip direction in space. This is illustrated by Fig. 3.100, which shows the P-wave radiation pattern for a thrust event with some strike-slip component. Thus, the estimation of the P-wave first motion polarities and their back-projection onto the focal sphere allows us to identify the type of focal mechanism of a shear event (termed fault-plane solution). The only problem is that the hypocenter and the seismic ray path from the source to the individual stations must be known. This may be difficult for rays propagating in a heterogeneous Earth with 2-D or even 3-D velocity structure.



**Fig. 3.100** Radiation pattern of the radial displacement component (P wave) due to a double-couple source: a) for a plane of constant azimuth (with lobe amplitudes proportional to  $\sin 2\theta$ ) and b) over a sphere centered on the origin. Plus and minus signs of various sizes denote the alternating amplitude variations (with the spherical coordinates  $\theta$  and  $\phi$ ) of outward (+; extensional quadrant) and inward (-; compressional quadrant) directed motions of the source. The fault plane and auxiliary plane are nodal lines on which  $\cos \phi \sin 2\theta = 0$ . The pair of open arrows in a) at the center denotes the shear dislocation and **P** and **T** in b) mark the penetration points of the pressure and tension axes, respectively, through the focal sphere. (modified from Aki and Richards 1980; with kind permission of the authors).

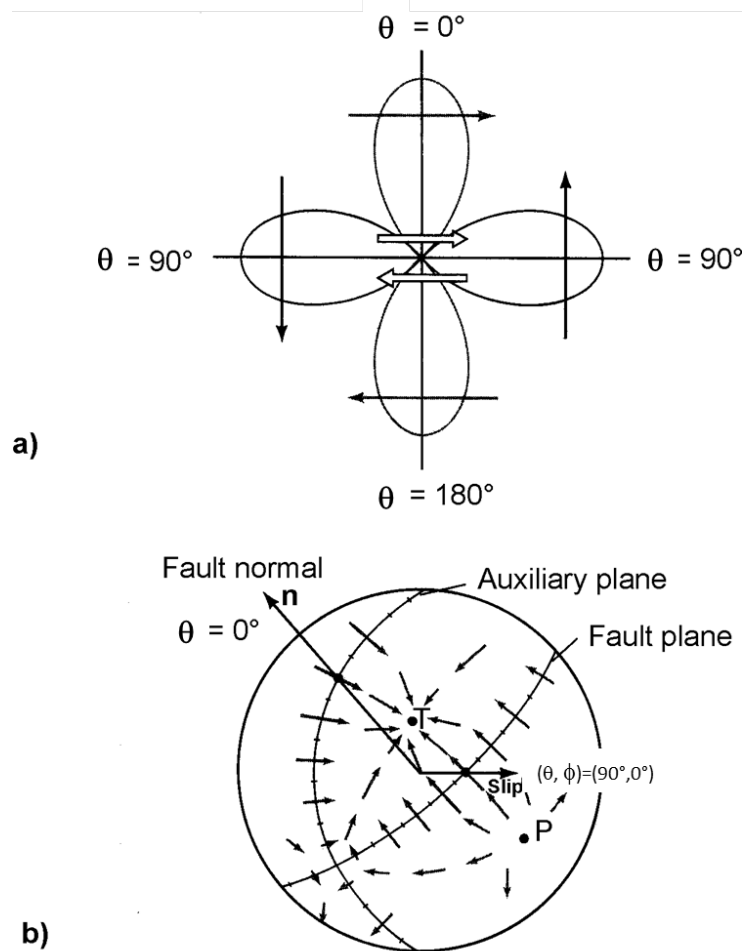
Fault-plane solutions based on P-wave first motion polarities will be better constrained if additionally the different radiation pattern of S-waves displacement amplitudes is taken into account. In the case of a double-couple mechanism, the S-wave amplitude pattern follows the relationship (see Aki and Richards, 1980)

$$\mathbf{A}_S = \cos 2\theta \cos \phi \boldsymbol{\theta} - \cos \theta \sin \phi \boldsymbol{\phi} \quad (3.205)$$

with  $\boldsymbol{\theta}$  and  $\boldsymbol{\phi}$  - unit vectors in  $\theta$  and  $\phi$  direction,  $\mathbf{A}_S$  - shear-wave displacement vector.

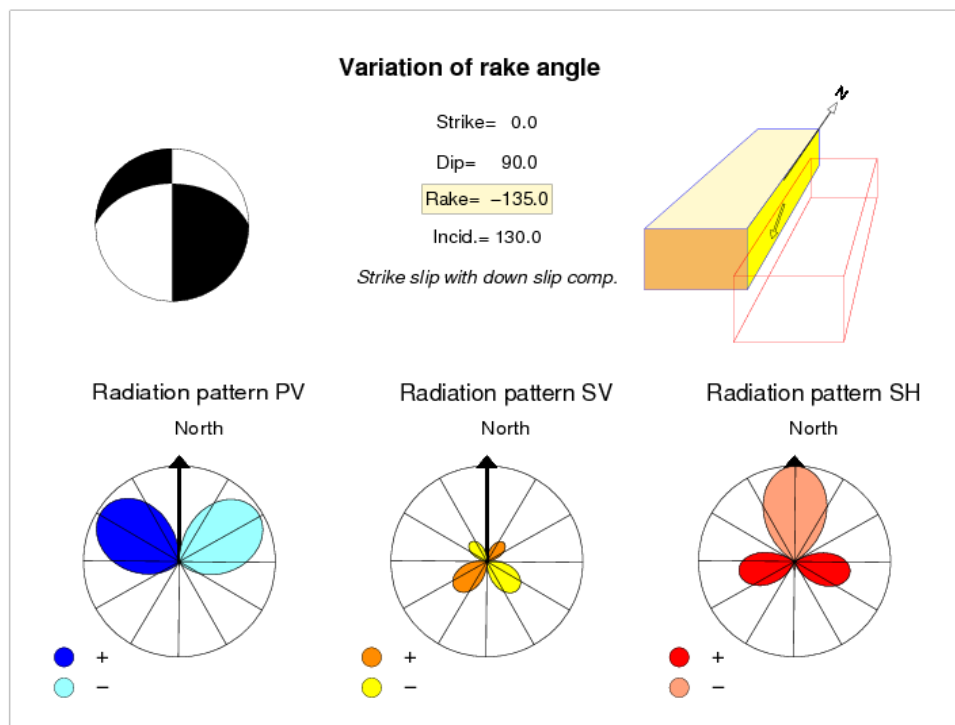
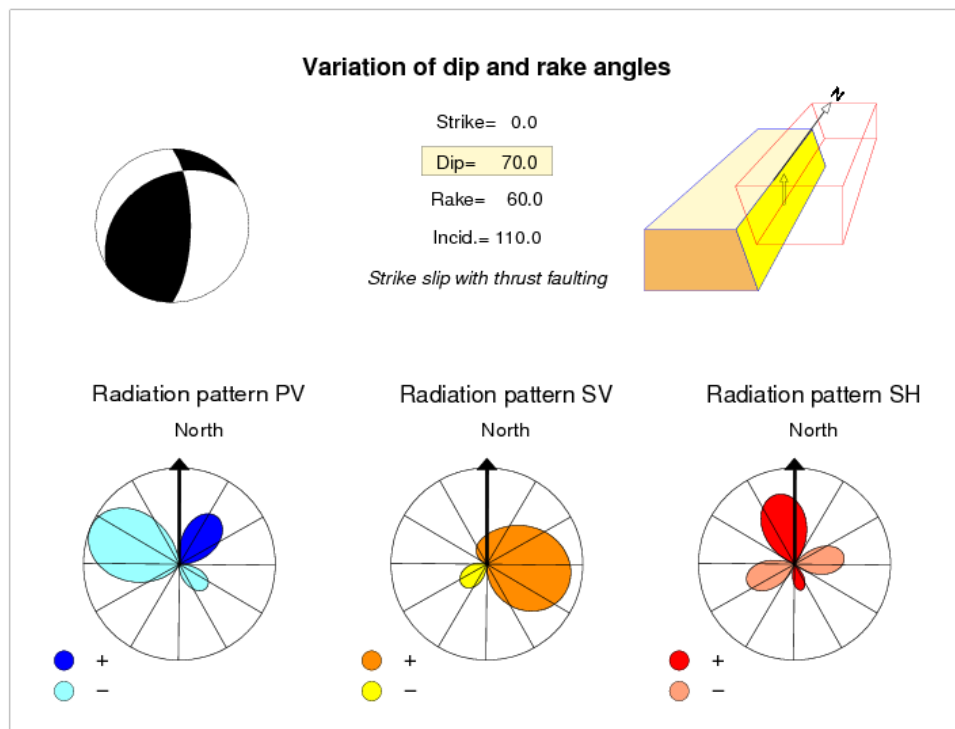
An example is given in Fig. 3.101 for the same fault-plane solution as shown in Fig. 3.100 for P waves. Fig. 3.102 shows the calculated azimuth-dependent vertical component PV, as well as vertically polarized SV and horizontally polarized SH amplitude lobe patterns for two different combinations of **strike, dip and rake** angles of a seismic fault as well as different incidence angles *Incid* of the seismic rays at the station. *Incid* corresponds in the case of a shallow source in a laterally homogeneous 1-D Earth model to the take-off angles of the

considered seismic ray from the source. And Fig. 3.103 shows the P- and S-wave radiation pattern lobes for a local earthquake with respect to the azimuth towards two seismic stations of a local network in Germany and their related records, which show striking differences in their P/S amplitude ratios. For a more general introduction into source directivity and the radiation patterns of earthquake fault mechanisms see section 2.4 in IS 1.1. It provides additional figures plus animations that illustrate the great variability of amplitude radiation and directivity patterns, depending on the controlling fault and rupture parameter combinations.

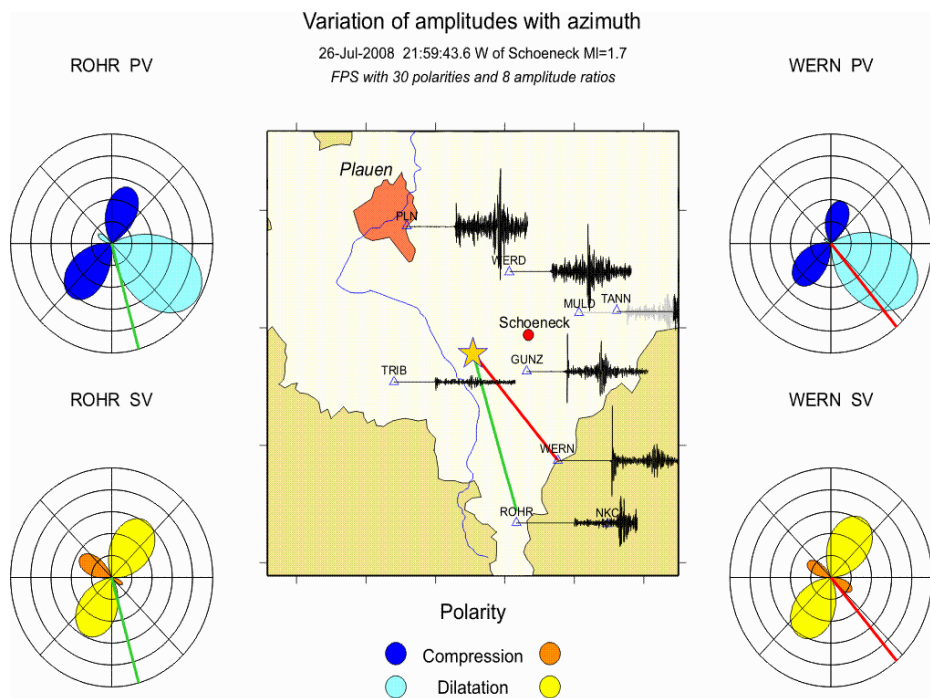


**Fig. 3.101** Radiation pattern of the transverse displacement component (S wave) due to a double-couple source. a) in the plane  $\{\phi = 0, \phi = \pi\}$  and b) over a sphere centered on the origin. Arrows imposed on each lobe in a) show the direction of particle displacement associated with the lobe while the arrows with varying size and direction in the spherical surface in b) indicate the variation of the transverse motions with  $\theta$  and  $\phi$ . P and T mark the penetration points of the pressure and tension axes, respectively, through the focal sphere. There are no nodal lines as in Fig. 3.100 but only nodal points where there is zero motion. The nodal point for transverse motion at  $(\theta, \phi) = (45^\circ, 0^\circ)$  at T is a maximum in the pattern for longitudinal motion (see Fig.3.100) while the maximum transverse motion (e.g., at  $\theta = 0$ ) occurs on a nodal line for the longitudinal motion. The pair of arrows in a) at the center denotes the shear dislocation (modified from Aki and Richards 1980; with kind permission of the authors).





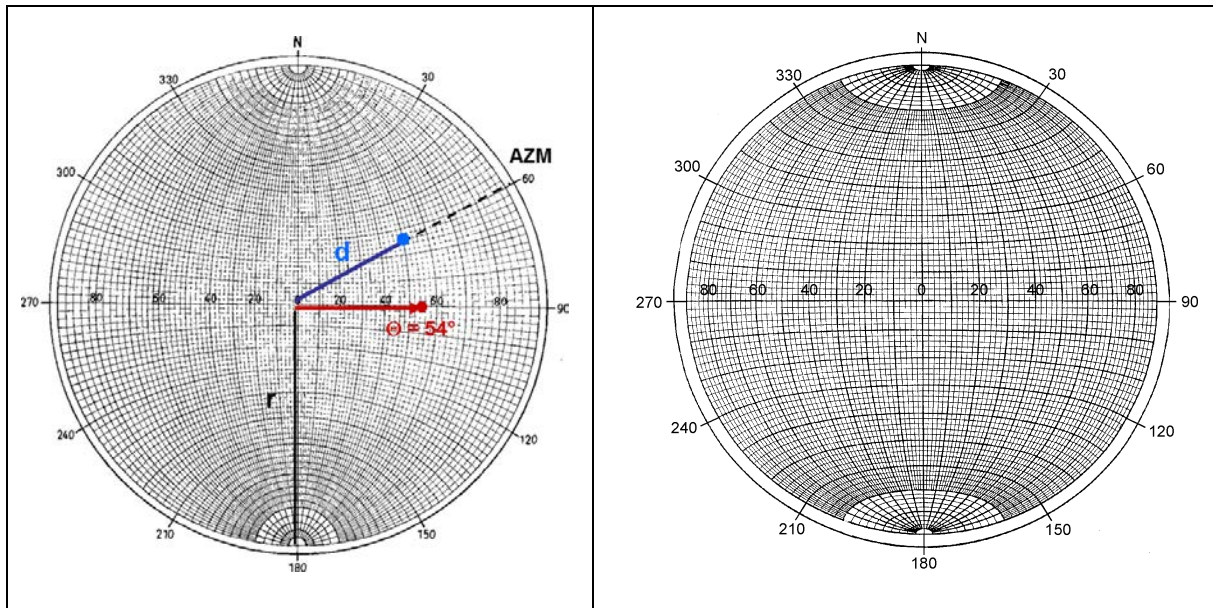
**Fig. 3.102** Radiation patterns of PV, SV and SH amplitudes for two earthquake ruptures with different fault **dip**, **rake** (= slip direction) and **take-off angle** of the seismic rays at the source (which is in the case of a shallow seismic source in a laterally homogeneous 1-D Earth model equal with the incidence angle **Incid** at the receiving seismic station).



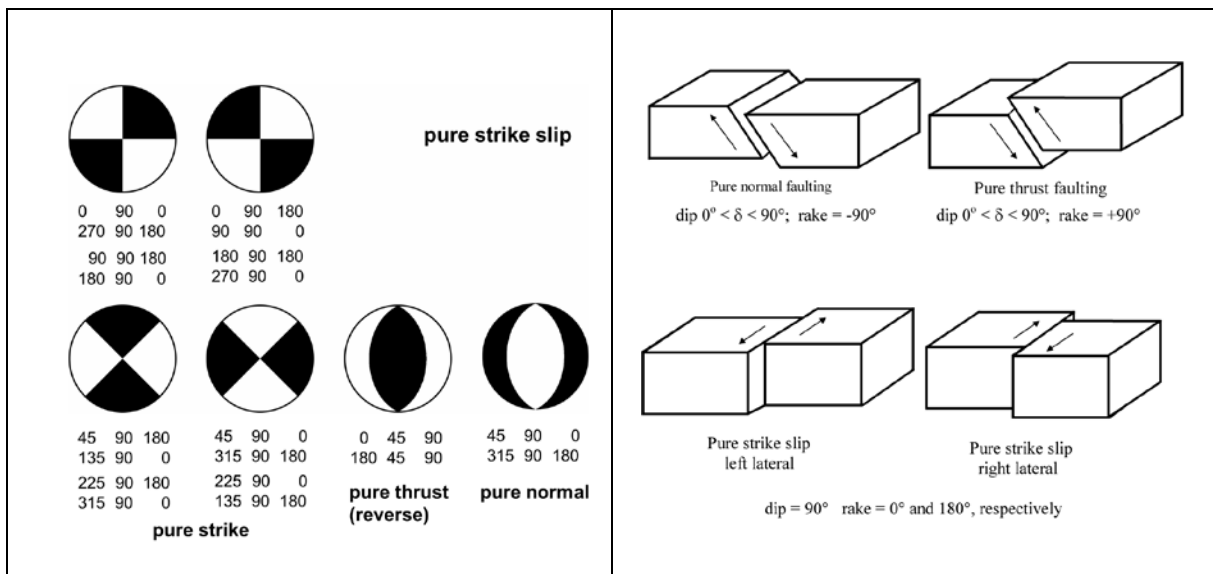
**Fig. 3.103** Comparison of theoretically calculated radiation patterns for an earthquake with well constrained fault plane solution and slip direction with the really observed amplitude ratios between vertical component P and S amplitudes at stations of a local seismic network in Germany. The source depth of the earthquake has been 12.1 km and the epicenter (respectively hypocenter) distances of stations WERN and ROHR were 13.5 km (18.1 km) and 17.5 km (21.3 km).

### 3.4.2 Determination of fault-plane solutions from P-wave polarities

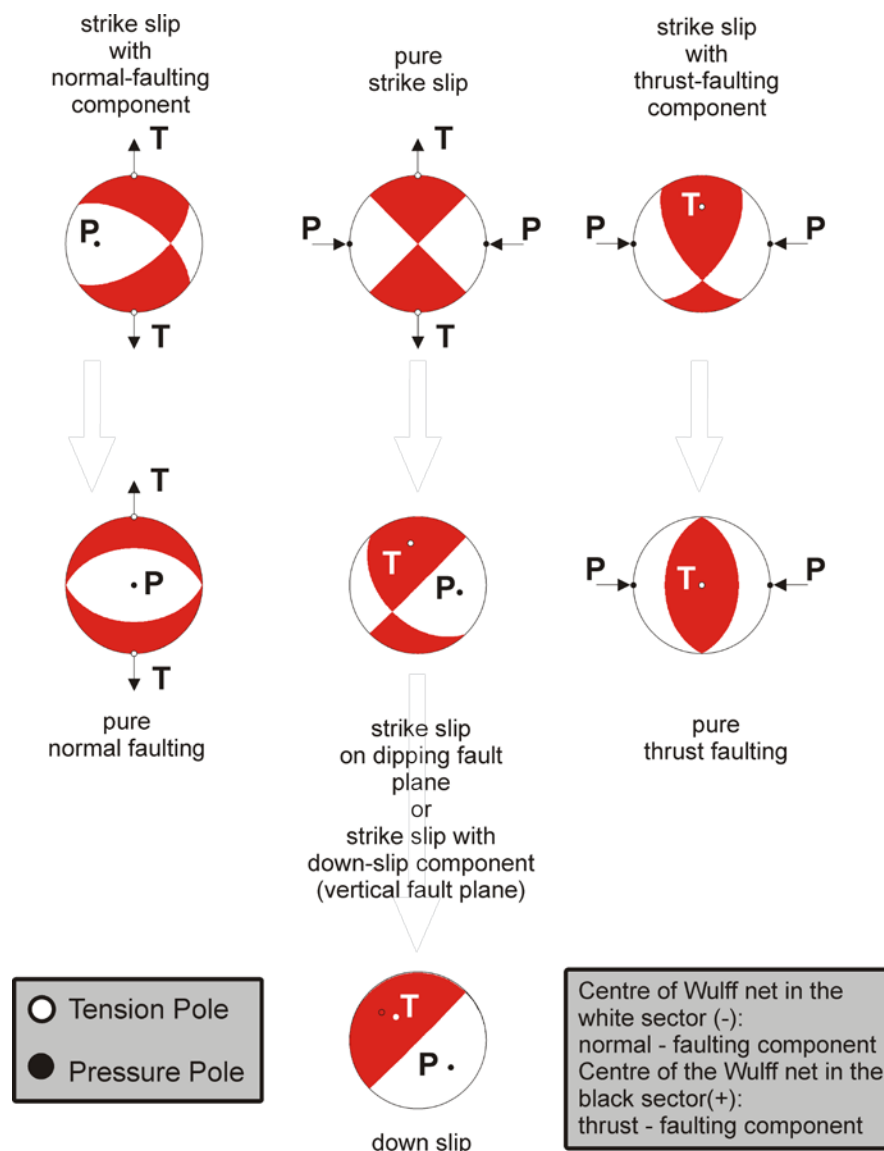
Most fault-plane solutions are still based on P-wave polarity readings only, especially for smaller earthquakes, although a steadily growing number results nowadays from moment tensor solutions via waveform fitting (see Figs. 3.66 and 3.67 as well as IS 3.8, 3.10 and 3.11). Advanced event location and seismogram analysis programs allow plotting of measured polarities on either an equal-angle Wulff net or a Lambert-Schmidt equal area projection (Figs. 3.104; see also Aki and Richards, 1980, Vol. 1, p. 109-110), yet without showing the net grids itself. Then they search for the optimal separation of the quadrants with plus and minus polarities. The quadrants are separated by the two potential fault planes which are, in the assumed idealized rupture model in a homogeneous full-space, perpendicular to each other. As the final result, usually only the black-and-white, or in different colors shaded quadrants of the horizontal 2-D focal-sphere representations, nick-named also as “beach-ball” solutions, are then plotted and presented together with the respective numbers for the strike, dip and rake of the two fault planes. For the basic types of strike-slip, normal and thrust faulting this is shown in Fig. 3.105 and for mixed types of faulting in Fig. 3.106. In the latter figure additionally the poles of the axes of maximum pressure P and maximum tension T have been marked. In this idealized rupture model they are assumed to be situated in the center of the negative and positive polarity quadrants, respectively.



**Fig. 3.104 Left:** Equal angle Wulff net, for which holds  $d = r \cdot \tan(\Theta/2)$  with  $r$  – radius of the net and  $d$  – distance from the network center of a data point with a given azimuth  $AZM$  and take-off angle  $\Theta$ . **Right:** equal area Lambert-Schmidt net with  $d = r \cdot \sin(\Theta/2)$ .



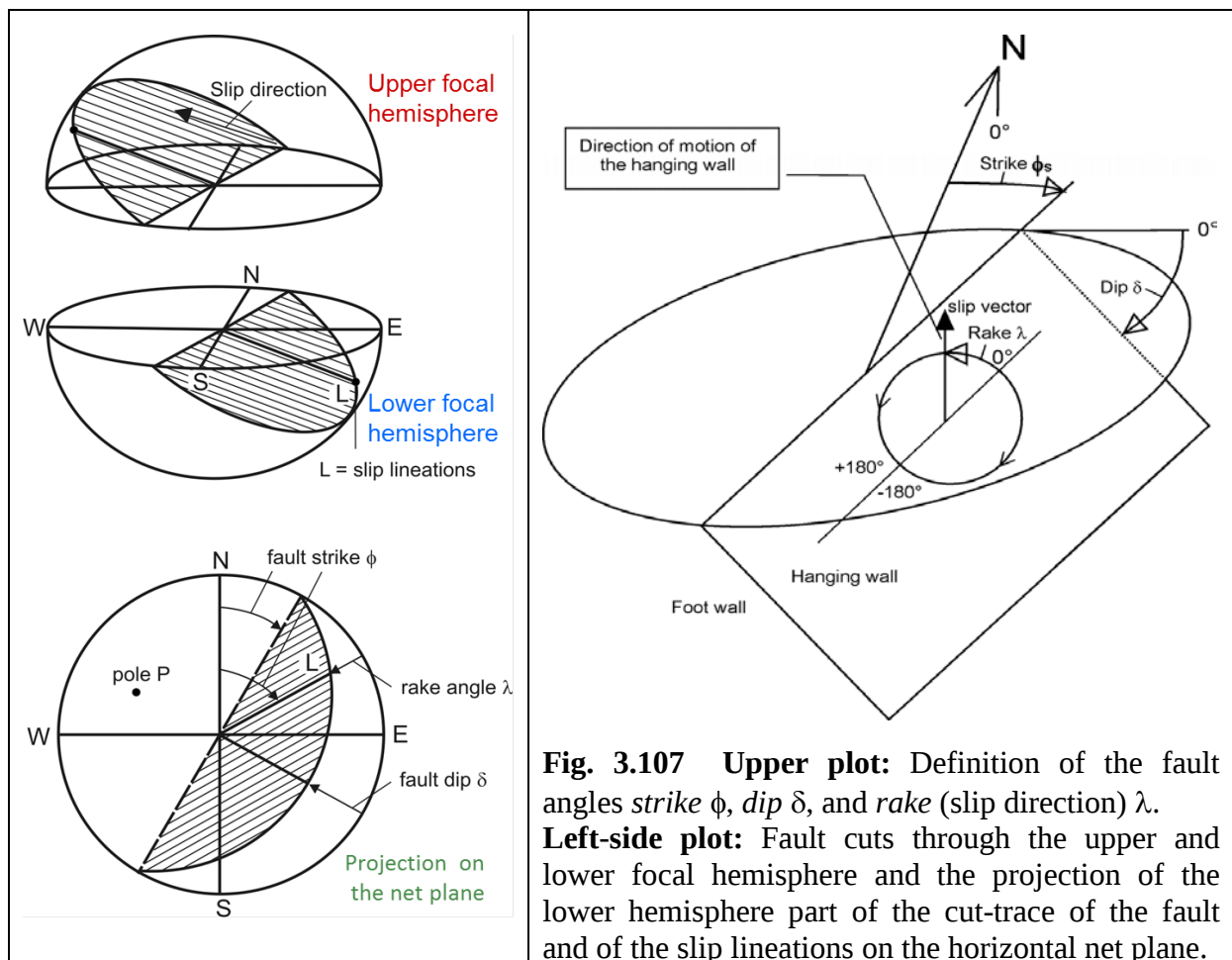
**Fig. 3.105 Left:** “Beach-ball” representations of basic types of faulting together with the respective numbers of strike, dip and rake angle for the respective fault plane and auxiliary plane, which is a potential fault plane as well. **Right:** Cartoons of the related motions of crustal blocks. Note that a pure strike-slip rupture may occur also on a fault that does not dip vertically and reverse and normal ruptures on planes with other dips than 45 degrees. Moreover, especially for ruptures near to the Earth surface the complementary fault planes may in reality not intersect at a right angle (e.g., by steepening of the dip of bended listric faults towards the surface). See also discussion under 3.1.2.6.



**Fig. 3.106** “Beach-ball” presentation of the net projections of the fault plane **cut-traces**, the quadrants of different polarities, and of the penetration points (poles) of the P- and T-axes through the **lower focal hemisphere** for different faulting mechanisms. White sectors correspond to negative and black sectors to positive P-wave first-motion polarities. Note that mixed types of faulting occur when the rake angle  $\lambda \neq 0, 180^\circ$  or  $\pm 90^\circ$ , e.g., in the case of normal faulting with a strike-slip component or strike-slip with a thrust component. Also, dip angles  $\delta$  may vary between  $0^\circ < \delta \leq 90^\circ$  (modified Figure after M. Baumbach †, 2001).

Two important terms, printed in bold letters in Fig. 3.106, have to be explained, namely the notions **cut-trace** and **lower focal hemisphere**. Our problem is that the fault plane cuts through a conceived 3-D “focal sphere”, which can be separated into an upper and a lower one. All cut traces of the fault through the focal sphere are great circles. In order to reconstruct the orientation of the fault plane in space and to measure its strike, dip and rake angles in a 2-D

stereoscopic net we have to project the great circle cut trace of the fault through the focal sphere onto the horizontally placed net. The degree of bending of the projected cut traces depends on the fault dip. Vertically dipping faults are represented by straight lines running through the center of the net, whereas half-circles along the perimeter of the net represent a horizontal fault. However, from Fig. 3.107 we realize that a single fault would yield two such cut-trace projections, one resulting from the cut through the upper hemisphere and the other from the cut through the lower hemisphere. These two traces are mirrored on the strike direction of the fault, resulting in an  $180^\circ$  ambiguity. To avoid this, one has first to make up one's mind which data should be projected to the horizontal diagram, those related to the lower hemisphere or to the upper hemisphere. This, however, depends on the distance range of observation. In the local range, the P-wave first motion will usually be related to  $P_g$ , which has left the source upward through the upper focal hemisphere. However, stations at distances larger than about 150 to 250 km will record  $P_n$  as the first P-wave, which has left the source through the lower focal hemisphere. The same applies to all teleseismic P-wave rays. However, with EX 3.2 we demonstrate that even in the local distance range, considering only records at distances less than 50 km, the first arriving P-wave first may already be a critically refracted P at a pronounced velocity discontinuity in the upper crust and thus of a head-wave type as  $P_n$  which has left the lower focal hemisphere. If one then wishes to make use of the data of both upper and lower hemisphere rays in order to have a better azimuth and take-off angle distribution of data in the diagram then one has to correct the data from the non-priority hemisphere accordingly. How the corrections of the take-off angles and azimuths of the “wrong” hemisphere rays have to be calculated is illustrated with Figures 1 and 2 of EX 3.2.



Although black-box computed fault-plane solutions are very comfortable, they are educationally not very revealing about how the procedure really works, on which assumptions it is based and how accurate the fault-plane solutions are. In order to get at least a feeling for it as well as an appreciation for the cumbersome tedious work accomplished by the early “fathers” of systematic investigations of fault-plane solutions and their relationships to seismotectonics, one should have done oneself at least once a few fault-plane solutions manually. We outline in the following the basic principles and provide with EX 3.2 an exercise with detailed step-by-step guidance, solutions and discussions.

A good seismogram analysis software, used for phase onset and polarity picking as well as the event location, can produce as a spin-off output also a file with the following data (see, e.g., Table 1 in EX 3.2): Station code, epicenter distance, azimuth AZM of the station with respect to the source, take-off angle  $\Theta$  of the seismic ray from the source, and the P-wave first motion polarities + (or U = upward motion in the record) or – (or D = downward motion in the record). These polarities have to be plotted on either the Wulff or the Lambert net at the correct position for the given AZM and  $\Theta$  of seismic ray towards the station (as shown in Fig. 3.104, left side, for one observation in the Wulff net). Principally, both the Wulff net and the Lambert net projections yield quantitatively the same results. The latter, however, has the advantage that it provides visually a less cluttered plot of data with take-off angles less than  $45^\circ$ .

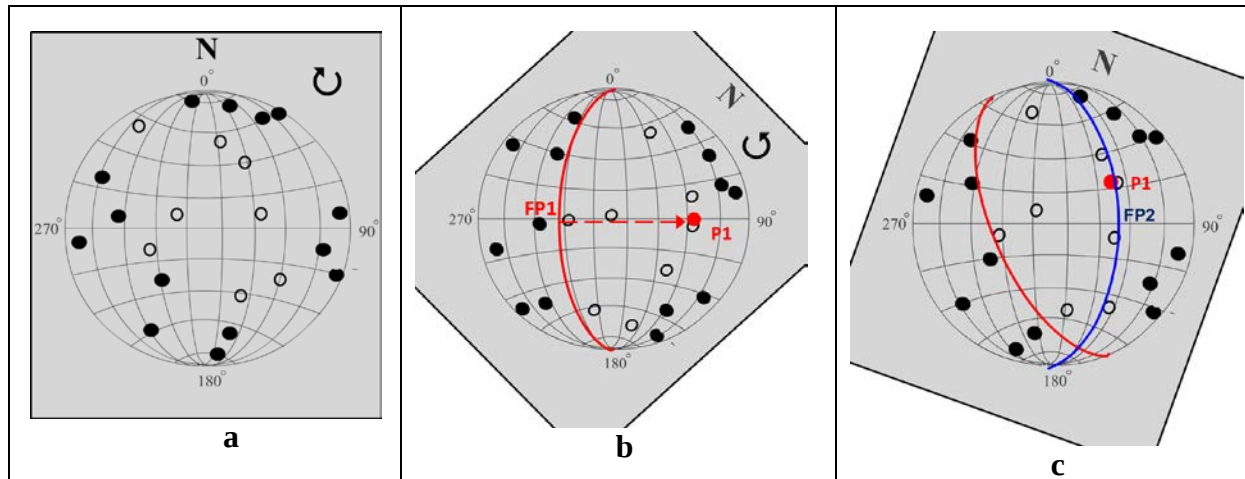
As a first step we should pin a transparent sheet with a needle to the net center and mark the N, E, S, and W direction as well as the network perimeter on the transparent sheet. Next we transfer all data points according to their AZM and  $\Theta$  values on the transparent sheet (shown in gray on Fig. 3.108a) and mark them there with their respective polarities in an unambiguous signature. Then the transparent sheet should be rotated over the net center until one finds on the underlying net a great circle meridian which separates best the + and – signs (or related symbols as in Fig. 3.108). By clockwise rotation, we find in our demonstration example a meridian on the left side which separates well the compressional first motions from the dilatational ones. This great circle trace should be drawn with a marker pen on the transparent sheet and termed fault plane one = FP1 (red in Fig. 3.108b).

Next we search for the complementary second (or auxiliary) fault plane trace FP2, remembering that it should be situated  $90^\circ$  apart from FP1 in the case of rupture in a homogeneous full-space. Therefore, we should mark on the transparent sheet the pole P1 of FP1, through which FP2 has to go. P1 is perpendicular from the middle of FP1,  $90^\circ$  away towards the right (red dot in Fig. 3.108b). By counter clockwise back rotation of the transparent sheet we find then indeed another great circle, marked in blue in Fig. 3.108c, which goes through P1 and separates at least the majority of black dots from the majority of the open circles on the right side. FP1 and FP2 cross each other in the lower part of the net at an angle of  $90^\circ$ . Accordingly, the pole P2 of FP2 would be placed on FP1,  $90^\circ$  upwards from this crossing point of the two fault planes. Note that all angles have to be measured on great circles.

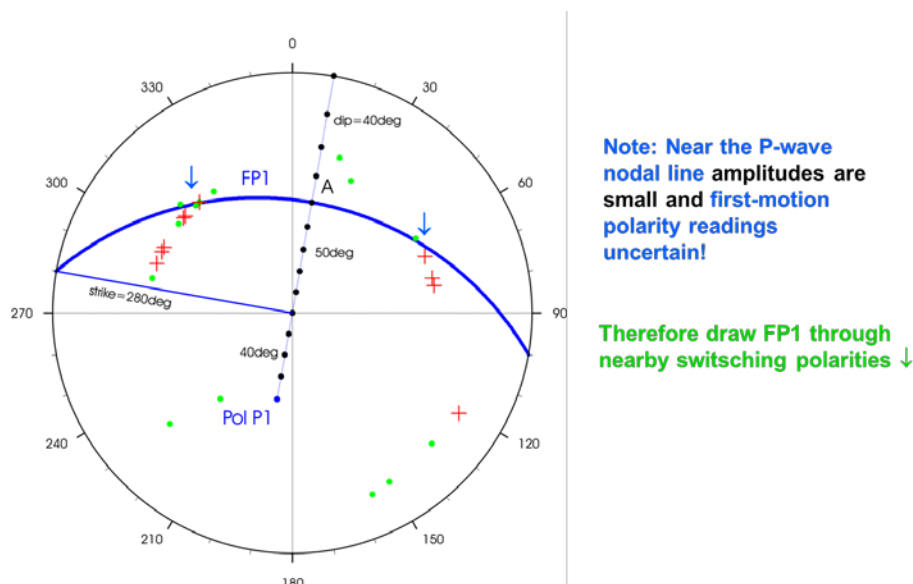
You will realize that sometimes polarities + and – are rather close to each other, respectively very close to the fault plane trace which is supposed to separate them. As in Fig. 3.108c, we might even find some polarities to be on apparently the wrong side of the fault, in the wrong quadrant. This, however, is unavoidable. We should remember that a fault plane is a nodal line for P waves (see Fig. 3.100), along which P-wave amplitudes vanish or are very small. Accordingly, if recorded at all, they have a very low signal-to-noise ratio. This may result in



wrong reading of the polarity of the usually weaker first half swing. Therefore, nearby placed opposite polarities in the data plot are even “strategic points” to look for, because they are likely to be situated close to the nodal lines. This is also illustrated below by one example from EX 3.2 (Fig. 3.109). In fact, computer programs for finding the optimal fault-plane cut traces through polarity plots aim at minimizing both their distance from conflicting polarity clusters as well as of the number of wrong polarities in the respective quadrants.



**Fig. 3.108** Illustration of the first essential steps when manually searching for the two fault plane traces that associate at least the majority of read polarities correctly with compressional and dilatational quadrants. FP1 = fault plane 1, FP2 = auxiliary second fault plane, P1= pole of FP1. Modified and amended from Andresen (Fault plane solutions <http://www.learninggeoscience.net/free/00071>).

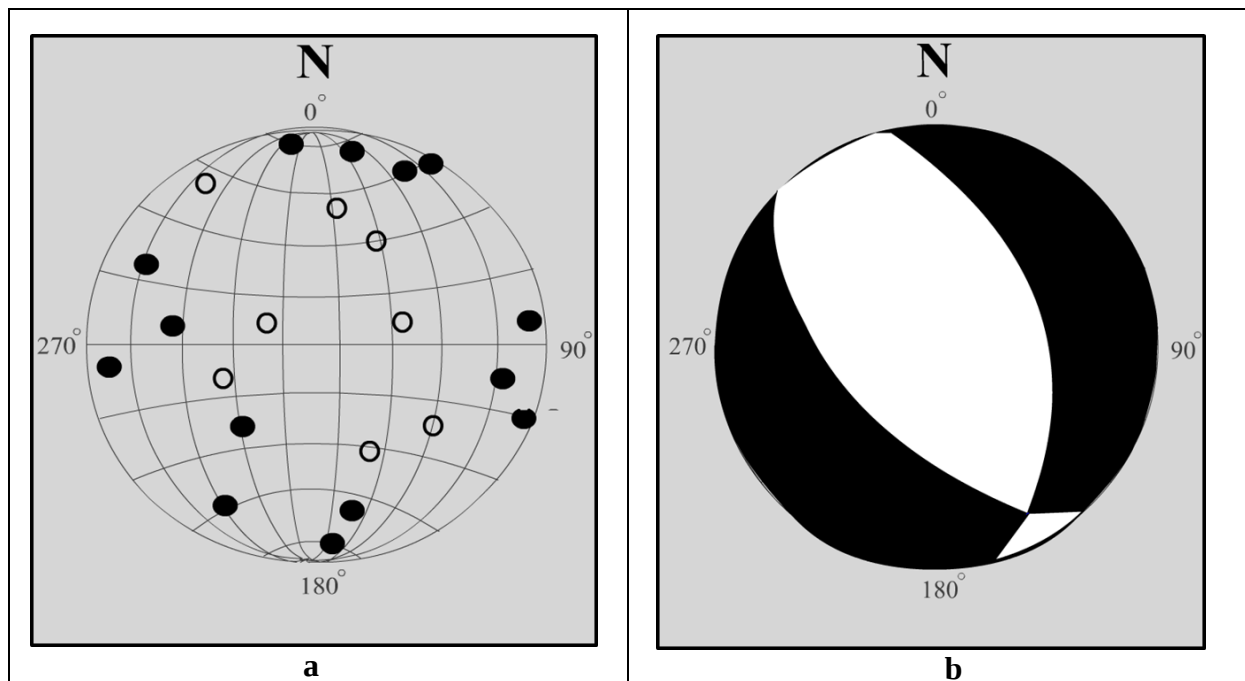


**Fig. 3.109** Example for another data set from EX 3.2 to illustrate how FP1 has been found as an appropriate great circle which passes through two clusters of nearby switching polarities, (see blue arrows) and at the same time separating well the majority of + and – (green dots) first motions in the upper part of the data plot. Shown is also how to find the pole P1 and how to measure the dip and the strike of FP1.

In summary, for obtaining a fault-plane solution, basically three steps are required:

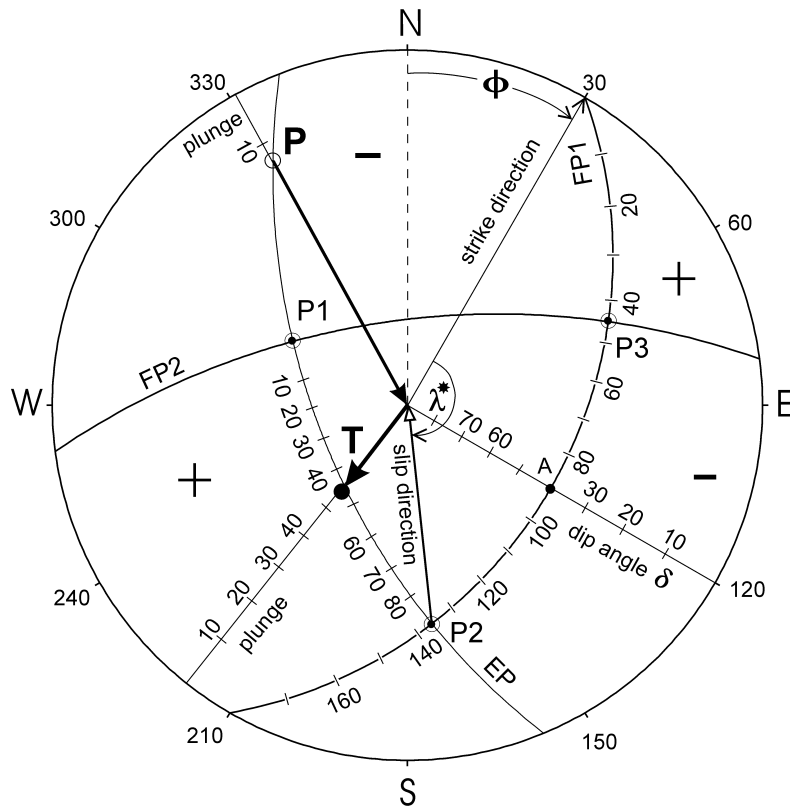
- (1) Calculating the positions of the penetration points of the seismic rays through the focal sphere which are defined by the ray azimuth AZM and the take-off (incidence) angle  $\Theta$  of the ray from the source.
- (2) Marking these penetration points through the upper or lower hemisphere in a horizontal projection of that sphere using different symbols for compressional and dilatational first arrivals. Usually, lower hemisphere projections are used. Rays which have left the upper hemisphere have then to be transformed into their equivalent lower hemisphere ray. This is possible because of spherical symmetry of the radiation pattern (see Fig. 3.100).
- (3) Partitioning the projection of the lower focal sphere by two perpendicular great circles which separate all (or at least most) of the + and - arrivals in different quadrants.

When moving back the marked N direction on the transparent overlay sheet in Fig. 3.108c to the north direction of the stereographic net then we get the beach-ball fault-plan solution depicted in Fig. 3.110. And Fig. 3.111 shows, how the strike, dip, and rake (slip) angles of the two potentially acting faults FP1 and FP2 as well as the azimuth and plunge angles of the pressure and tension axes P and T can then be read on the net diagram. P and T are situated on the equatorial plane EP midway between FP1 and FP2. For more details and how one can decide by way of field evidence, which of the two calculated fault planes has likely be the acting one, see EX 3.2.



**Fig. 3.110 Left:** Polarity data plot in a Lambert-Schmidt net with black dots representing compressional P-wave first arrivals and open circles dilatational signals, respectively; **Right:** the “beach-ball” fault-plane solution deduced from this polarity pattern. It represents a normal faulting with a small strike-slip component along a NNW, respectively SSE striking fault.





**Fig. 3.111** Determination of the fault plane parameters  $\phi$ ,  $\delta$  and  $\lambda$  in the net diagram. The polarity distribution, slip direction and projection of FP1 in this diagram corresponds qualitatively to the faulting case depicted in Fig. 3.107 upper right, namely to a thrust earthquake with a strike-slip component. For abbreviations used see text. **Note:**  $\lambda^* = 180^\circ - \lambda$  when the center of the net lies in the tension (+) quadrant (i.e., event with thrust component) or  $\lambda^* = -\lambda$  when the center of the net lies in the pressure quadrant (i.e., event with normal faulting component). P1, P2 and P3 mark the positions of the poles of the planes FP1 (fault plane), FP2 (auxiliary plane) and EP (equatorial plane) in their net projections. All three planes are perpendicular to each other and intersect in the poles of the respective third plane, i.e., FP1 and FP2 in P3, FP1 and EP in P2 etc. **P** and **T** are the penetration points (poles) of the pressure and tension axes, respectively, through the focal sphere. + and – signs mark the quadrants with compressional and dilatational P-wave first motions.

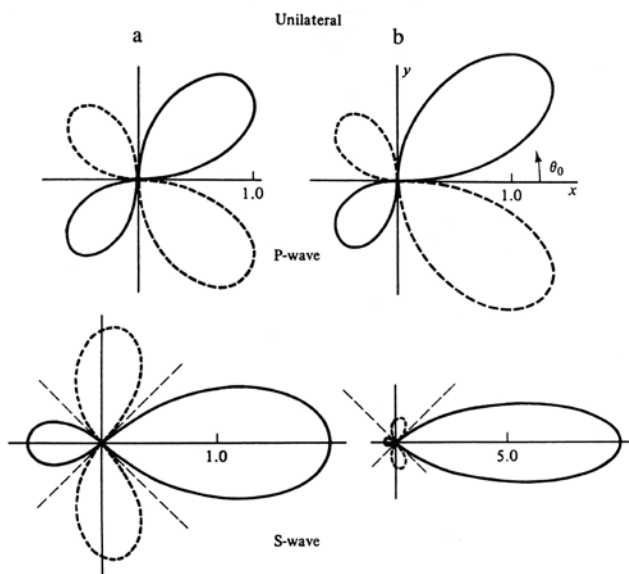
For quick proper understanding of published beach-ball solutions, one should just remember a few “rules of thumb”, which are well documented by Fig. 3.106:

Note that on the basis of polarity readings alone it cannot be decided whether FP1 or FP2 was the active fault.

- If the two potential fault planes cross each other within the beach ball or net, then there is some strike-slip component involved;
- In the case of pure strike-slip the two fault traces cross as straight lines in the center of the net, respectively the beach-ball;
- If this center is situated in a dilatation or in a compression quadrant then there is a normal faulting, respectively a thrust faulting component involved;

- In the case of pure normal or thrust faulting, either the P axis or the T axis is placed in the net or beach-ball center and the two fault planes do not cross each other within the perimeter (see Fig. 3.106);
- For proper determination of the strike directions of the two faults think yourself standing in the net center with the FP1 fault trace towards your right-hand side. Then look forward along the fault trace until it meets the perimeter. The strike direction of FP1 is then the angle clockwise from north which you read on the perimeter of the net. For finding  $\phi(\text{FP2})$  you have to turn around at the net center so as to place now FP2 at your right-hand side, and then look again towards the end point of FP2 on the perimeter. For the fault-plane solution in Fig. 3.111 you thus get  $\phi(\text{FP1}) = 30^\circ$  and  $\phi(\text{FP2}) = 157^\circ$ ;
- On the basis of polarity readings alone it cannot be decided whether FP1 or FP2 was the active fault.

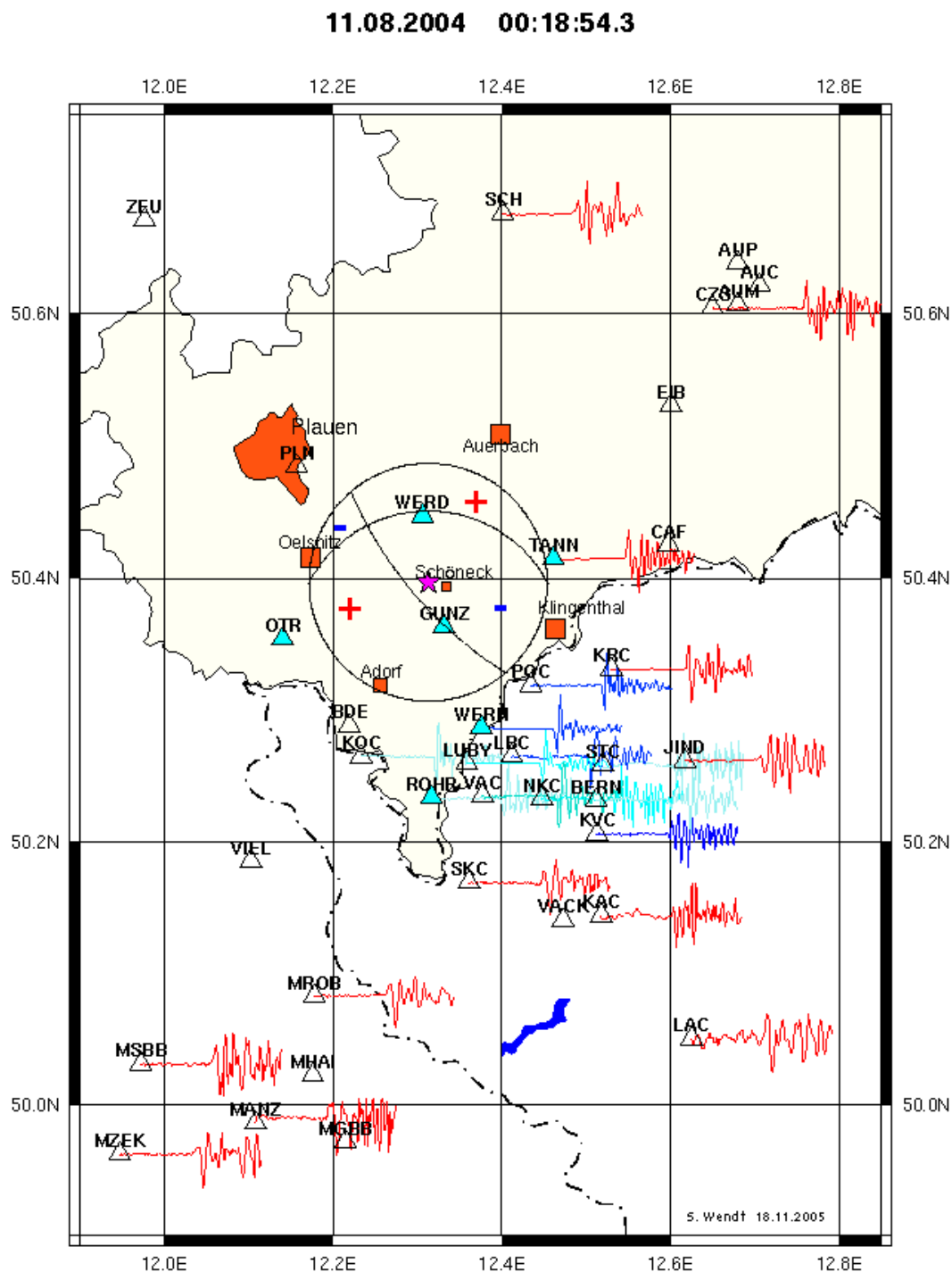
However, discrimination from seismological data alone is still possible but requires additional study of the *directivity effects* such as azimuthal variation of frequency (*Doppler effect*), amplitudes and/or waveforms (see, e.g., Fig. 3.112 and related discussions and figures in Lay and Wallace, 1995, and in section 2.4 in IS 1.1). For sufficiently large shocks these effects can more easily be studied in low-frequency teleseismic recordings while in the local distance range, high-frequency waveforms and amplitudes may be strongly influenced by resonance effects due to near-surface low-velocity layers (see Chapter 14).



**Fig. 3.112** Example for an amplitude directivity effects: Variability of P- and SH-wave amplitude for a fault rupture propagating unilaterally from left to right with a ratio between rupture and shear wave velocity  $v_r/v_s = 0.5$  (left column) and  $v_r/v_s = 0.9$  (right column), respectively. (From Kasahara, 1981; © Cambridge University Press).

Moreover, seismotectonic considerations or field evidence from surface rupture in the case of strong shallow earthquakes may also allow to resolve this ambiguity (see EX 3.2, Figure 7 and related discussion).

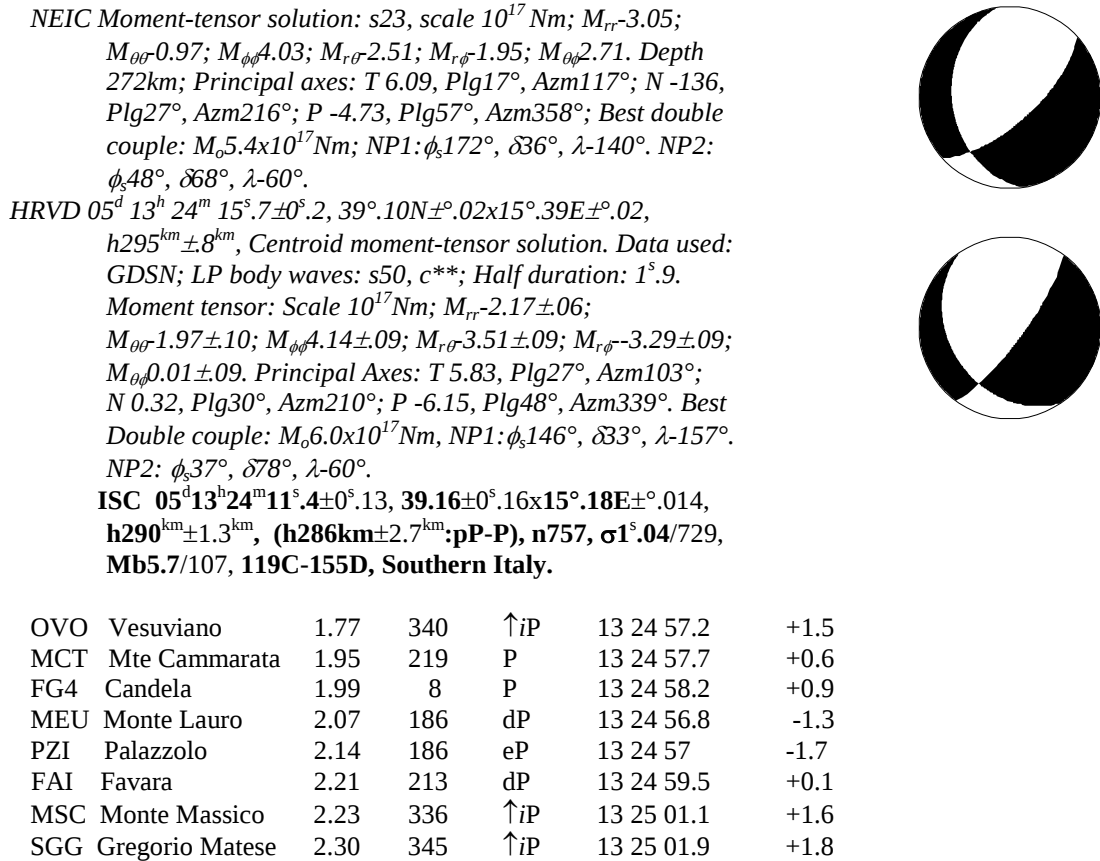
Finally, we present with Fig. 3.113 an example for the fault-plane solution derived for a local earthquake in the Vogtland swarm earthquake region in Germany by means of polarity readings. One can nicely associate the positive first-motion polarity traces (in red) and the respective negative record traces (in blue and turquoise) with the related + and – quadrants of the fault-plane solution plotted over the epicenter of this earthquake.



**Fig. 3.113** Vertical component recordings of Pg first arrivals at seismic stations in Germany from a local earthquake at epicentre distances between 10 km and 55 km in different azimuths and the „beach ball“ fault-plane solution derived from first motion polarities plotted over the epicenter. The acting fault plane had a strike of about 150° north. It was a normal faulting rupture with a left-lateral strike-slip component.

### 3.4.3 Accuracy of fault-plane solutions

Fault planes determined by eye-fit to the polarity data may be uncertain by about  $\pm 10^\circ$ . This is acceptable. Even computer assisted best fits to the data will produce different acceptable solutions within about the same error range with only slightly different standard deviations (e.g., Figure 3.114 with NEIC and HRVD solutions, respectively).



**Figure 3.114** Typical section of an ISC bulletin (left) with NEIC (USGS National Earthquake Information Center) and Harvard University (HRVD) moment-tensor fault-plane solutions (right) for the Italy deep earthquake (h = 286 km) of Jan. 05, 1994. Columns 3 to 5 of the bulletin give the following data: 3 - epicentral distance in degrees, 4 - azimuth AZM in degrees, 5 - phase code and polarity.

In addition, one has to be aware that different fitting algorithm or error-minimization procedures may produce different results within this range of uncertainty for the same data. A poor distribution of seismograph stations (resulting in insufficient polarity data for the net diagram), erroneous polarity readings and differences in model assumptions (e.g., in the velocity models used) may result in still larger deviations between the model solutions and the actual fault planes. One should also be aware that the assumed constant angular ( $45^\circ$ ) relationship between the fault plane on the one hand and the pressure and tension axis on the other hand is true in fact only in the case of a fresh rupture in a homogeneous isotropic full-space without internal friction. It may not be correct in the stress environment of real tectonic situations with prefractured rock, distinct anisotropic rock fabric and internal friction. Then **P** and **T**  $\neq \sigma_1$  and  $-\sigma_3$ , respectively, is likely (see dynamics of faulting by Anderson, 1951, and discussion in section 3.1.2.6: *Parameters which describe and control the source mechanism*).

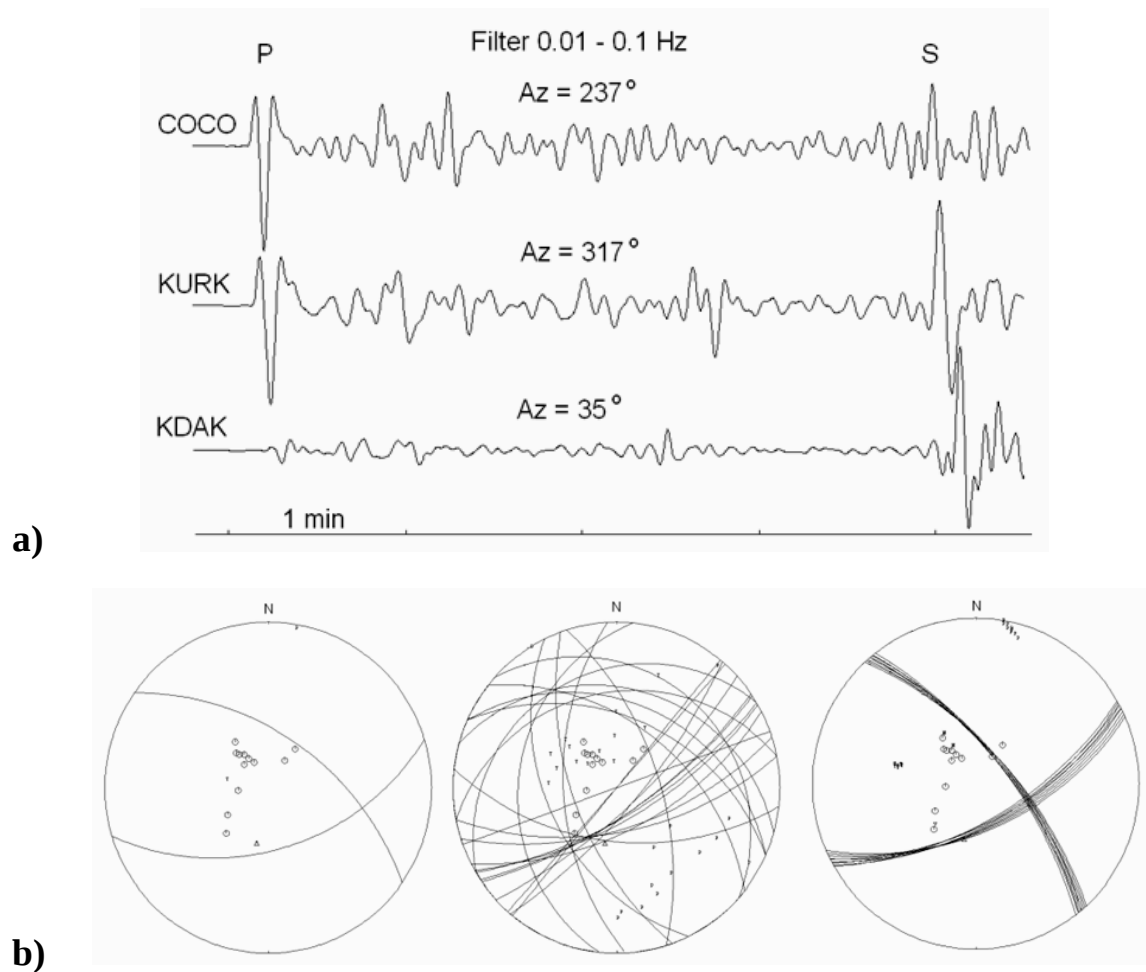
### 3.4.4 Computer-assisted fault-plane solutions (P. Bormann)

There exist quite a number of computer programs for the determination of both single and joint fault-plane solutions from first-motion data (e.g., Brillinger et al., 1980; Bufo and Udías, 1984; Udías and Bufo, 1988, and others referred to below). In most applications for local earthquakes homogeneous flat-layered velocity models are acceptable, i.e., layers with constant velocities and velocity discontinuities at the boundaries. The majority of location programs (e.g., HYPO71 by Lee and Lahr, 1975; HYPOELLIPSE by Lahr, 1989 and 2003; HYPOINVERSE by Klein, 1978, 1985 and 2002) are based on this type of velocity model. However, HYPOINVERSE and HYPOELLIPSE do also accept layers with linear velocity gradients. Moreover, HYPOELLIPSE may locate local events with predefined travel-time tables, too. During the location procedure the ray paths to the stations are calculated. The azimuth AZM and the take-off angle  $\Theta$  at which the P wave, arriving at a given station, leaves the focal sphere are listed in the output files. The remaining problem to be solved is to find the distribution of P-polarities on the focal sphere and to estimate the angles describing the focal mechanism.

The computer program FPFIT (Reasenberger and Oppenheimer, 1985) calculates double-couple fault-plane solutions based on P-wave polarity readings. It accepts as input the output files of the localization programs HYPO71, HYPOELLIPSE and HYPOINVERSE. The inversion is accomplished through a grid-search procedure that finds the source model by *minimizing a normalized weighted sum of first-motion polarity discrepancies*. Two weighting factors are incorporated in the minimization. One of them reflects the estimated variance of the data while the other one is based on the absolute value of the P-wave radiation amplitude. In addition to the minimum-misfit solution, FPFIT finds alternative solutions corresponding to significant relative misfit minima. The existence of several minima may be due to insufficient number of polarity readings, localization errors, polarity misreadings or an inadequate velocity model (e.g., not modeled refractions) resulting in an incorrect position of the P-wave first-motion polarities on the focal sphere. One has also to be aware that it sometimes may happen that the seismometer component outputs have been wrongly plugged at a given station, resulting in systematically wrong polarity readings by such a station. In the case of models which perfectly fit the data, FPFIT applies an additional constraint. Its effect is to maximize the distance sum between the observation points and the nodal planes on the focal sphere. The display program FPLOT shows the final fault-plane solution and the estimated uncertainty in terms of the range of possible orientations of the pressure and tension axes which is consistent with the data. This USGS software for Unix and Linux platforms is also available from the ORFEUS Software Library.

While the above programs accept only the output files of the hypocenter localization programs for local events, another widely used program package for seismogram analysis, SEISAN (Havskov, 1996; Havskov and Ottemöller, 1999) uses a modified version of the program HYPOCENTER (Lienert et al., 1988; Lienert, 1991; Lienert and Havskov, 1995). Its version 9.2 can be downloaded via the link *Download programs and files* on the front page or directly via <http://folk.uib.no/lot081/seisan.html>. The main modifications are that it can also accept secondary phases and locate teleseismic events. The output files are used in conjunction with the programs FOCMEC (Snoke et al., 1984; current version 2008 can be downloaded from the IRIS software library), FPFIT and HASH (see below) for the determination of the fault plane parameters by using both P-wave first-motion polarities and P/S amplitude ratios, both in the local, regional and teleseismic distance range.

In the case of sparse networks or weak events, the number of polarity data may be too small and/or their distribution in distance and azimuth not appropriate for reliable estimation of fault-plane solutions on the basis of P-wave polarity readings alone. In this case P-, SV- and SH-amplitudes can be used in addition to polarities in order to get more stable and better constrained, i.e., less ambiguous fault-plane solutions. This is due to the difference in the azimuth and take-off angle dependent P-wave and S-wave polarity and amplitude patterns for a given source mechanism (see Figs. 3.100 – 3.103). Fig. 3.115 illustrates this for a teleseismic deep earthquake.



**Fig. 3.115 a)** Vertical component P and S-wave records from a deep earthquake near Japan (28 Sept. 2007, mb(PDE) = 6.7, h = 260 km). The three stations have similar epicentral distances ( $\Delta = 54 - 58^\circ$ ) and all recorded a compressional P-wave first motion. This does not tell much about the source mechanism, in contrast to the S/P amplitude ratio which strongly differs with azimuth Az.

**b)** The fault-plane solutions FPS obtained for this earthquake. **Left:** The FPS derived from the Harvard moment tensor inversion, here plotted together with 13 available P-wave polarity readings which all fall within one quadrant. **Middle:** Completely inconsistent possible FPSs obtained by a grid search for the 13 polarity readings, all but one being compressional. P and T are the related possible pressure and tension axes **Right:** FPSs obtained by using the 13 polarities and 5 amplitude ratios from 3 stations. Triangle is dilation, hexagon compression and H and V indicate horizontal and vertical component amplitude ratios. Copied with kind agreement of the authors from Havskov and Ottemöller (2010); © Springer Publishers.

The program FOCMEC allows to calculate best fitting double-couple fault-plane solutions from P, SH and SV polarities and/or SV/P, SH/P or SV/SH amplitude ratios provided that the ratios are corrected to the focal sphere by taking into account geometrical spreading, attenuation and free-surface effects. For surface correction the program FREESURF, which is supplied together with FOCMEC, can be used. The applied Q-model has to be specified according to the regional attenuation conditions or related corrections. When adopting a constant  $V_P/V_S$  velocity ratio, the geometrical spreading is the same for P and S waves and absolute changes in amplitude cancel each other in the above amplitude ratios. Head waves and amplitude changes at velocity boundaries require special treatment. The solution is obtained by grid search over strike, dip and slip of the double-couple source. The program FOCPLT, also provided together with FOCMEC, allows to plot upper or lower hemisphere projections of the focal sphere and to show the data, i.e., the fault planes, together with the poles of the pressure (P) and tension (T) axes for SH and SV waves. Note that S-wave amplitudes are zero in the direction of P and T.

A more recent USGS program is HASH 1.0 for Unix platforms. It also computes double-couple earthquake focal mechanisms from both P-wave first motion polarities and, optionally S/P amplitude ratios.

FMSI is a program developed by Gephart (1990). It is a focal mechanism stress inversion package using earthquake focal mechanisms and fault/slickenside data. Kikuchi and Kanamori (1982 and 1986) developed a program for teleseismic body-wave inversion aimed at deriving information about the fault kinematics and/or fault mechanics, and Another program for interactive moment tensor retrieval, termed ISOLA-GUI (Authors: Jiri Zahradnik and Efthimios Sokos) can be downloaded via [seismo.geology.upatras.gr/isola/](http://seismo.geology.upatras.gr/isola/).

Other important software tools are presented in the International Handbook on Earthquake and Engineering Seismology (Lee et al., 2003) (TDMT\_INVC, Time Domain Seismic Moment Tensor Inversion Code by Douglas Dreger; [ftp://www.orfeus-eu.org/pub/software/iaspei2003/8511.html](http://ftp://www.orfeus-eu.org/pub/software/iaspei2003/8511.html)) and in “Computer Programs in Seismology” (Charles J. Ammon and George Randall; regional moment tensor inversion set of programs; <http://www.eas.slu.edu/eqc/eqccps.html>). For more information and links to programs consult the ORFEUS Software Library (<http://www.orfeus-eu.org/Softwarelib.html>) and IRIS (<http://www.iris.washington.edu>).

### **3.4.5 Estimating $M_0$ , the size of rupture area, average slip and stress-drop from measured seismic spectra (P. Bormann)**

Besides deriving some rough estimates of the point source type of rupture, i.e., of the fault-plane solution and average slip direction by analyzing the azimuthal direction of P-wave first motion polarities and/or S/P amplitude ratios additional information about the seismic moment, the size of the rupture area, the average slip and stress drop can be derived by analyzing the seismic spectra of the recorded P and/or S waves. However, with reference to the discussions in section 3.3.1 on seismic scaling laws one should be aware that all these inferences on not directly measurable source parameters are model-based and none of the available models can correctly account for all the parameter and rupture complexities in the real inhomogeneous Earth lithosphere. Therefore, all calculations aimed at deriving these additional parameter informations, have to make assumptions. We shortly demonstrate this

here by way of example, also with reference to the related exercise EX 3.4 and the conclusions drawn from it.

With reference to Fig. 3.5 and to formulas (3.1) and (3.2) the scalar seismic moment is written

$$M_0 = \mu \bar{D} A = 4\pi d \rho v_{p,s}^3 u_0 / R_{\theta,\varphi}^{p,s} \quad (3.206)$$

and thus can be calculated when measuring the spectral displacement plateau amplitude  $u_0$  of the radiated seismic source spectrum. However, this is not directly available, but is available only via the spectra recorded at some distance and specific azimuth away from the source. Therefore, in order to get a reasonable realistic estimate of the primary source spectrum one has to correct it for wave propagation and source radiation effects. This, however, requires either reliable knowledge or good average model assumptions about the

- hypocenter distance  $d$  of the source;
- distribution of density  $\rho$  and of the P- or S-wave velocity  $v_p$  or  $v_s$  in the more or less inhomogenous Earth;
- frequency-dependent intrinsic and scattering attenuation of the recorded wave amplitudes for correcting the measured spectral amplitudes;
- surface amplification factor that depends on the incidence angle of the considered seismic wave at the recording station and which is in fact again frequency dependent (see Fig. 2.7 in Chapter 2);
- source radiation coefficient  $R$  with respect to the seismic ray recorded at the given seismic station, which is different for P and S waves and depends on the take-off angle and azimuth under which considered seismic ray has left the source.

It is obvious, that most of these factors are only roughly known and that some of the errors, e.g. in the velocity model, propagate even with the third power (Eq. (3.200)).

Additionally, the recorded waveforms of well defined non-dispersive seismic body phases have a limited duration of only a few seconds for near earthquakes, up to a few minutes at best for very strong teleseismic events. This limits the spectral range that can be analyzed and makes the recorded spectra rather noisy (see Fig. 3.116). This limits in turn also the accuracy with which  $u_0$  and the corner frequency  $f_c$  can be measured on real spectra.  $f_c$  is related to the geometric size of the rupture and instrumental for calculating either the radius of assumed circular ruptures (usually for smaller sources) or via even more vaguely estimated additional corner frequencies in the sloping part of the spectrum also of the width  $W$  and length  $L$  of the rupture plane of larger earthquakes. Knowing either  $R$  or  $L$  plus  $W$ , one may roughly estimate the rupture area via the relationship

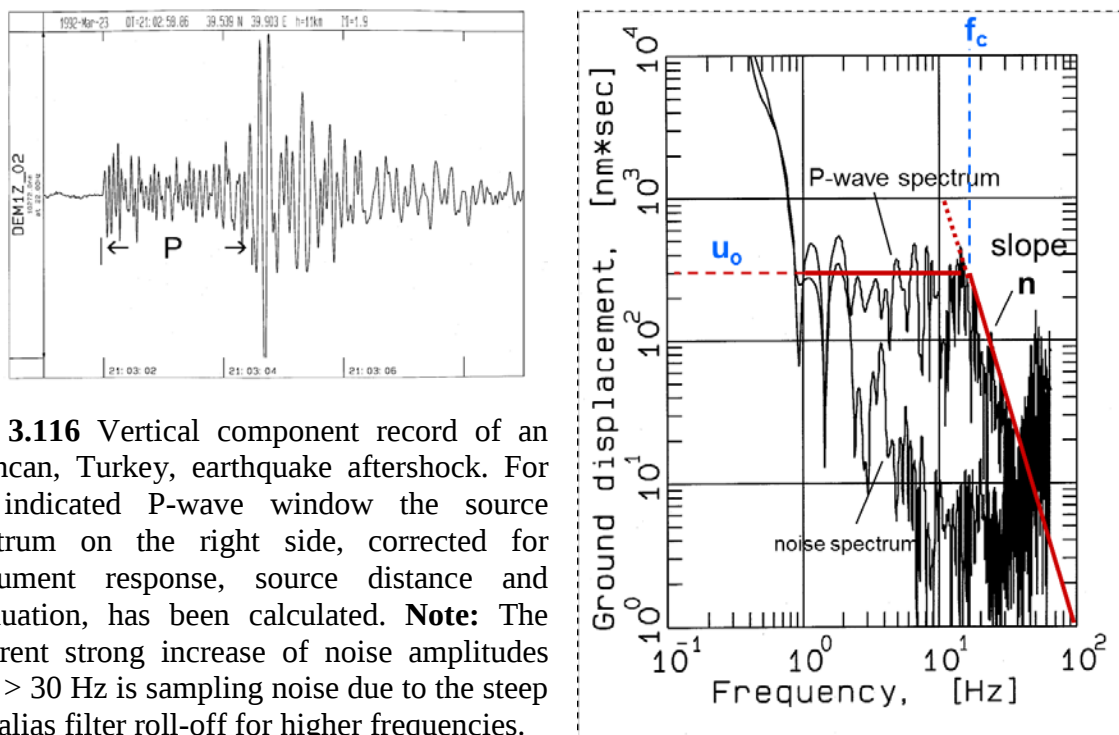
$$f_c = c_{m1} v_{p,s} / \pi R = c_{m2} v_{p,s} / (L \times W)^{1/2} \quad (3.207)$$

with  $c_{m1}$  and  $c_{m2}$  being model-dependent constants. Via (3.201) one can relate  $f_c$  to a critical wavelength  $\lambda_c$  which is in the order of the source diameter or the sqr of the product  $L \times W$ :

$$\lambda_c = v_{p,s} / f_c = c_{m3} \pi R = c_{m4} (L \times W)^{1/2}. \quad (3.208)$$

Again,  $c_{m3}$  and  $c_{m4}$  are other model-dependent parameters.





**Fig. 3.116** Vertical component record of an Erzincan, Turkey, earthquake aftershock. For the indicated P-wave window the source spectrum on the right side, corrected for instrument response, source distance and attenuation, has been calculated. **Note:** The apparent strong increase of noise amplitudes for  $f > 30$  Hz is sampling noise due to the steep anti-alias filter roll-off for higher frequencies.

With EX 3.4 we show that even for identical measurement input parameters  $u_0$  and  $f_c$ , which are both already afflicted with model and reading errors, the calculated values for  $R$ ,  $A$ , average displacement  $\bar{D}$  via Eq. (3.200), and stress drop via the formula  $\Delta\sigma = (7/16) M_0 / R^3$  according to Eshelby (1957) and Keilis-Borok (1959), may already differ by a factor of about 1.8, 3, 4 and 6, respectively, depending on whether one calculates them either according to the Brune (1970, 1971) or to the two Madariaga (1976) models, which all assume a circular rupture. This means, however, that one has to use all such model-derived parameters with great caution and accept that they cannot be accurate in absolute terms. Nevertheless, they may be useful, especially when comparing relative parameter changes in space and time in certain seismotectonic areas, or, e.g., for specific aftershock sequences or earthquake swarms. And if in future for the given area better models are available then one may have a chance also to re-calibrate earlier data with the hope to improve also their accuracy in absolute terms.

## Comments and Acknowledgments

The new Chapter 3 is a strongly revised and largely expanded version of Chapter 3 in the NMSOP-1 editions of 2002 and 2009. Two of the former co-authors (M. Baumbach and G. Bock) have regrettably passed away already at much too young an age. We have kept and acknowledged some of the figures by M. Baumbach in the current Chapter and moved others in the related exercise EX 3.2 by Bormann, Baumbach and Wendt. The former sub-chapter of G. Bock has now the status of an independent Information Sheet (IS 3.8), thus being a lasting memory to Günther's major contribution to NMSOP-1. Also, the original sub-chapter on seismic energy determination by G. Choy has become an independent IS 3.6, complemented already by IS 3.5 and awaiting another committed IS 3.10 on seismic energy determination from local and regional seismic events, in collaboration with J. L. Boatwright. Thus, the upgrading and complementation of these important and rapidly developing modern topics have been made independent on the future availability of the editor and main author of Chapter 3. In concordance with this G. Bock's IS 3.8 has also been upgraded and

complemented by a new information sheet IS 3.9 on seismic moment tensor solutions in the local, regional and teleseismic range and by a practical, EX 3.6, on seismic moment tensor inversions using the KIWI tools (with downloadable data and software). All these new or still forthcoming complementary IS should be considered as being integral parts of Chapter 3.

The current finally reviewed and revised version of Chapter 3 replaces the earlier pre-review publication on the Internet. The authors owe greatest thanks to J. W. Dewey of the USGS, co-chairman of the IASPEI/CoSOI Working Group on Magnitude Measurements, for his painstakingly careful and constructive review of the whole Chapter. We have greatly benefitted from his comments, recommendations, corrections and hints to missing items. For several more partial reviews, editorial changes and additional contributions, some made in terms of complementary figures, the editor owes great thanks to J. Dewey, G. Choy, J. Havskov, G. Nolet, A. Lomax, A. Michelini, P. Gasperini, B. Lolli and L. Ottemöller. Quite many figures have also been contributed by J. Saul of the GFZ German Research Centre for Geosciences. They are based on collaborative work and joint journal publications with P. Bormann as well as on a joint presentation with P. Bormann and S. Wendt at the 2012 General Assembly of the European Seismological Commission. All contributed figures by non-authors have been separately acknowledged in the respective sections and captions.

## **Recommended overview readings**

Aki and Richards (1980, 2002 and 2009)  
Båth (1966, 1981)  
Ben Menahem and Singh (1981 and 2000)  
Bormann (2011)  
Bormann and Saul (2009a)  
Das and Kostrov (1988)  
Di Giacomo and Bormann (2011)  
Di Giacomo et al. (2013)  
Grünthal (2011)  
Kanamori and Brodsky (2004)  
Lay and Wallace (1995)  
McGarr et al. (2002)  
Richter (1958)  
Scholz (1990 and 2002)  
Udias (1999 and 2002)

## **References**

- Abe, K. (1973). Tsunami and mechanism of great earthquakes, *Phys. Earth Planet. Int.*, **7**, 143–153.
- Abe, K. (1975). Reliable estimation of the seismic moment of large earthquakes. *J. Phys. Earth*, **23**, 381–390.
- Abe, K. (1979). Size of great earthquakes of 1837–1974 inferred from tsunami data. *J. Geophys. Res.*, **84**, 1561–1568.
- Abe, K. (1981a). Magnitudes of large shallow earthquakes from 1904 to 1980. *Phys. Earth Planet. Interiors*, **27**, 72–92.

- Abe, K. (1981b). Physical size of tsunamigenic earthquakes of the northwestern Pacific. *Phys. Earth Planet. Inter.*, **27**, 194-205.
- Abe, K. (1982). Magnitude, seismic moment and apparent stress for major deep earthquakes. *J. Phys. Earth*, **30**, 321-330.
- Abe, K. (1984). Complements to "Magnitudes of large shallow earthquakes from 1904 to 1980". *Phys. Earth Planet. Interiors*, **34**, 17-23.
- Abe, K. (1989). Quantification of tsunamigenic earthquakes by the  $M_t$  scale. *Tectonophysics*, **166**, 27-34.
- Abe, K., and Kanamori, H. (1979). Temporal variation of the activity of intermediate and deep focus earthquakes. *J. Geophys. Res.*, **84**, B7, 3589-3595.
- Abe, K., and Kanamori, H. (1980). Magnitudes of great shallow earthquakes from 1953 to 1977. *Tectonophysics*, **62**, 191-203.
- Abercrombie, R. E. (1994). Regional bias in estimates of earthquake  $M_s$  due to surface-wave path effects. *Bull. Seism. Soc. Am.*, **84**, 2, 377-382.
- Abercrombie, R. E. (1995). Earthquake source scaling relationships from -1 to 5  $M_L$  using seismograms recorded at 2.5 km depth. *J. Geophys Res.*, **100**, 24,015-24,036.
- Adams, W. M., and Allen, D. W. (1961). Seismic decoupling for explosions in spherical underground cavities. *Geophysics*, **26**(6), 772-799.
- Adams, W. M., and Swift, L. W. (1961). The effect of shotpoint medium on seismic coupling. *Geophysics*, **26**(6), 765-771.
- Aki, K. (1966). Generation and propagation of  $G$  waves from the Niigata earthquake of June 16, 1964, part 2: Estimation of earthquake moment, released energy, and stress-strain drop from the  $G$  wave spectrum. *Bull. Earthq. Res. Inst. Toyo Univ.*, **44**, 73-88.
- Aki, K. (1967). Scaling law of seismic spectrum. *J. Geophys. Res.*, **72**, 1217-1231.
- Aki, K., and Chouet, B. (1975). Origin of coda waves: Source, attenuation and scattering effects. *J. Geophys. Res.*, **80**, 3322.
- Aki, K., and Patton, H. (1978). Determination of seismic moment tensor using surface waves. *Tectonophysics*, **49**, 213-222.
- Aki, K., and Richards, P. G. (1980). *Quantitative seismology – theory and methods*, Volume II, Freeman and Company, ISBN 0-7167-1059-5(v. 2), 609-625.
- Aki, K., and Richards, P. G. (2002). *Quantitative seismology*. 2<sup>nd</sup> ed., ISBN 0-935702-96-2, University Science Books, Sausalito, CA., xvii + 700 pp.
- Aki, K., and Richards, P. G. (2009). *Quantitative seismology*. 2<sup>nd</sup> ed., corr. Print., Sausalito, Calif., Univ. Science Books, 2009, XVIII + 700 pp., ISBN 978-1-891389-63-4 (see [http://www.ldeo.columbia.edu/~richards/Aki\\_Richards.html](http://www.ldeo.columbia.edu/~richards/Aki_Richards.html)).
- Albarello, D., Berardi, A., Margottini, C., and Mucciarelli, M. (1995). Macroseismic estimates of magnitude in Italy. *Pure Appl. Geophys.*, **145**, 2, 297-312.
- Allen, R.V., and Kanamori, H. (2003). The potential for earthquake early warning in Southern California. *Science*, **300**, 786-789.
- Alsaker, A., Kvamme, L. B., Hansen, R. A., Dahle, A., and Bungum, H. (1991). The  $M_L$  scale in Norway. *Bull. Seism. Soc. Am.*, **81**, 2, 379-389.
- Ambraseys, N. N. (1975). The correlation of intensity with ground motion. *Proc. XIV General Ass. ESC 1974 in Trieste*, Published by Nationalkomitee für Geodäsie und Geophysik, Akademie der Wissenschaften der DDR, Berlin, 335-341.

- Ambraseys, N. N. (1985). Magnitude assessment of northwestern European earthquakes. *Earthq. Eng. Struct. Dynam.*, **13**, 307-320.
- Ambraseys, N. N. (1988). Magnitude-fault length relationships for earthquakes in the middle east. In: W. H. K. Lee (ed.). *Historical seismograms and earthquakes of the world*. Academic Press, 309-310.
- Ambraseys, N. N. (1990a). Uniform magnitude re-evaluation of European earthquakes associated with strong-motion records. *Earthq. Eng. Struct. Dynam.*, **19**, 1-20.
- Ambraseys, N. N., and Bommer, J. J. (1990). Uniform magnitude re-evaluation for the strong motion database of Europe and adjacent areas. *European Earthquake Engineering*, **2**, 3-16.
- Ammon, C. J. (2001). Moment tensor inversion overview. <http://eqseis.geosc.psu.edu/~cammon/> (last accessed January 2005).
- Anderson, E. M. (1951). The dynamics of faulting and dyke formation with applications to Britain. 2<sup>nd</sup> ed. rev., *Oliver & Boyd Publishers*, Edinburgh.
- Anderson, D. L., Kanamori, H., Hart, R.S., and Liu, H. (1976). The Earth as a seismic absorption band. *Science*, **196**, 1104-1106.
- Anderson, D. L., and Given, J.W. (1982). Absorption band Q model for the Earth. *J. Geophys. Res.*, **87**, 3893-3904.
- Andresen, A. (n. y.) Fault plane solutions. <http://www.learninggeoscience.net/free/00071/> (last accessed October 2013).
- Atkinson, G. M., and Hanks, Th. (1995). A high-frequency magnitude scale. *Bull. Seism. Soc. Am.*, **85**, 3, 825-833.
- Atkinson, G. M., and Beresnev, I. A. (1997). Don't call it stress drop. *Seism. Res. Lett.*, **68**, 3-4.
- Bakun, W. H., and Joyner, W. (1984). The  $M_L$  scale in Central California. *Bull. Seism. Soc. Am.*, **74**, 5, 1827-1843.
- Bakun, W. H., and Lindh, A. G. (1977). Local magnitudes, seismic moments, and coda durations for earthquakes near Oroville, California. *Bull. Seism. Soc. Am.*, **67**, 615 -629.
- Bakun, W. H., and Wentworth, C. M. (1997). Estimating earthquake location and magnitude from seismic intensity data. *Bull. Seism. Soc. Am.*, **87**, 1502-1521.
- Bates, D. M., and Watts, D. G. (1988). *Nonlinear regression analysis and its applications* Wiley, New York.
- Bates, D. M., and Chambers, J. M. (1992). Nonlinear models. In Chambers, J. M. and Hastie, T. J. (Eds). *Statistical Models in S*, Wadsworth, Pacific Grove, CA, 421-454,
- Båth, M. (1962). Seismic records of explosions – especially nuclear explosions. *Part III FOA 4 Rapport*.
- Båth, M. (1962). Earthquake energy and magnitude. *Phys. Chem.of the Earth*, **7**, 115-165.
- Båth, M. (1981). Earthquake magnitude - recent research and current trends. *Earth Science Reviews*, **17**, 315-398.
- Båth, M. (1985). Surface-wave magnitude corrections for intermediate and deep earthquakes. *Physics Earth Planet. Interiors*, **37**, 228-234.
- Båth, M, Kulháněk, O., van Eck, T., and Wahlstroem, R. (1976). Engineering analysis of ground motion in Sweden. *Seismological Institute of Uppsala, University Report no. 5-76*.
- Bates, D. M., and Watts, D. G. (1988). *Nonlinear regression analysis and its applications*. Wiley, New York.

- Bates, D. M., and Chambers, J. M. (1992). Nonlinear models. In: Chambers, J. M. and Hastie, T. J. (eds), *Statistical Models in S*, 421-454, Wadsworth, Pacific Grove, CA.
- Ben-Menahem, A., and Singh, S. J. (1981). *Seismic waves and sources*. Springer Verlag, New York, Heidelberg, Berlin, 1108 pp.
- Ben-Menahem, A., and Singh, S. J. (2000). *Seismic waves and sources*." 2<sup>nd</sup> edition, Dover Publications, New York.
- Benz, H. M., Frankel, A., and Boore, D. M. (1997). Regional Lg attenuation for the continental United States. *Bull. Seism. Soc. Am.*, **87**(3), 606-619.
- Ben-Zion, Y. (1989). The response of two joined quarter spaces to SH line sources located at the material discontinuity interface. *Geophys. J. Int.*, **98**, 213-222.
- Ben-Zion, Y. (2001). On quantification of earthquake source. *Seism. Res. Lett.*, **72** (2), 151-152.
- Ben-Zion, Y., and Zhu, L. (2002). Potency-magnitude scaling relations for southern California earthquakes with  $1.0 < M_L < 7.0$ . *Geophys. J. Int.*, **148**, F1-F5
- Berckhemer, H., and Lindenfeld, M. (1986). Determination of source energy from broadband seismograms. In: Buttkus, B. (Ed.)(1986). Ten years of the Gräfenberg Array: Defining the frontiers of broadband seismology. *Geologisches Jahrbuch Reihe E*, Heft 35, 79-83.
- Beresnev, I. A. (2001). What we can and cannot learn about earthquake sources from spectra of seismic waves. *Bull. Seism. Soc. Am.*, **91**, 397-400.
- Beresnev, I. A. (2008). The reality of the scaling law of earthquake-source spectra? *J. Seismol.*; DOI: 10.1007/s10950-008-9136-9.
- Beyreuther, M., Barsch, R., Krischer, L., Megies, T., Behr, Y., and Wassermann, J. (2010). ObsPy: A Python Toolbox for Seismology. *Seism. Res. Lett.*, **81**(3), 530-533; doi: 10.1785/gssrl.81.3.530.
- Bindi, D., Spallarossa, D., Eva, C., and Cattaneo, M. (2005). Local and duration magnitudes in Northwestern Italy, and seismic moment versus magnitude relationships. *Bull. Seism. Soc. Am.*, **95**, 2, 592- 604.
- Bisztricsany, E. (1958). *A new method for the determination of the magnitude of earthquake*. *Geofiz. Kozl.*, **7**, 69-96 (in Hungarian with English abstract).
- Boatwright, J., and Choy, G. (1986). Teleseismic estimates of the energy radiated by shallow earthquakes. *J. Geophys. Res.*, **91**, B2, 2095-2112.
- Boatwright, J., and G. L. Choy (1989). Acceleration spectra for subduction zone earthquakes. *J. Geophys. Res.*, **94**, B11, 15,541-15,553.
- Bondár, I. and D. Storchak. (2011). Improved location procedures at the International Seismological Centre. *Geophys. J. Int.*, **186**, 1220-1244, doi:10.1111/j.1365-246X.2011.05107.x
- Bondár, I., Engdahl, E. R., Villaseñor, A., Harris, J., and Storchak, D. A. (2013). ISC-GEM: Global Instrumental Earthquake Catalogue (1900-2009), II. Location and seismicity patterns. *Phys. Earth Planet. Inter.*, submitted.
- Bonner, J. L., Russel, D. R., Harkrider, D. G., Reiter, D. T., and Herrmann, R. B. (2006). Development of a time-domain, variable-period surface-wave magnitude measurement procedure for application at regional and teleseismic distances, part II: Application and Ms-mb performance. *Bull. Seism. Soc. Am.* **96**, 2, 678-696, doi: 10.1785/0120050056.
- Boore, D. M. (1977). The motion of the ground in earthquakes. *Scientific American*, December 1977, W. H. Freeman and Company, San Francisco, 1-19.

- Boore, D. M. (1983). Stochastic simulation of high-frequency ground motions based on seismological models of the radiated spectra. *Bull. Seism. Soc. Am.*, **83**, 1865-1894.
- Boore, D. M. (1989). The Richter scale: its development and use for determining earthquake source parameter. *Tectonophysics*, **166**, 1-14.
- Bormann, P. (1966). *Recording and interpretation of seismic events (principles, present state and tendencies of development)* (in German: Registrierung und Auswertung seismischer Ereignisse – Grundlagen, Stand und Entwicklungstendenzen). Publications of the Institut für Geodynamik, Jena, Akademie Verlag, Berlin, Vol. 1, 158 pp.
- Bormann, P. (2011). Earthquake magnitude. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 207-218; doi: 10.1007/978-90-481-8702-7.
- Bormann, P., and Khalturin, V. I. (1975). Relations between different kinds of magnitude determinations and their regional variations. *Proceed. XIVth General Assembly of the European Seismological Commission*, Trieste, 16-22 September 1974. Published by the Nationalkomitee Geod. Geophys., AdW der DDR, Berlin, 27-39.
- Bormann, P., and Wylegalla, K. (1975). Investigation of the correlation relationships between various kinds of magnitude determination at station Moxa depending on the type of instrument and on the source area (in German). *Public. Inst. Geophys. Polish Acad. Sci.*, **93**, 160-175.
- Bormann, P., Wylegalla, K., and Grosser, H. (1992) The strong mining event of March 13, 1989: Analysis of the multiple event. *Proceedings XXII. General Assembly of the European Seismological Commission (ESC), Barcelona, 17-22 Sept. 1990*, Proceedings and Activity Report 1988-1990, Servei Geològic de Catalunya, 153-158.
- Bormann, P., and Wendt, S. (1999). Identification and analysis of longitudinal core phases: Requirements and guidelines. In: Bormann (1999), Regional International Training Course 1999 on Seismology, Seismic Hazard Assessment and Risk Mitigation. Lecture and exercise notes. Vol. I and II, GeoForschungsZentrum Potsdam, *Scientific Technical Report STR99/13*, 346-366.
- Bormann, P., and K. Wylegalla (2005). Quick estimator of the size of great earthquakes. *EOS, Transactions, American Geophysical Union*, **86** (46), 464.
- Bormann, P., and J. Saul (2008). The new IASPEI standard broadband magnitude  $m_B$ , *Seismol. Res. Lett.* **79** (5), 699-706.
- Bormann, P., and Saul, J. (2009a). Earthquake magnitude. In: *Encyclopedia of Complexity and Systems Science*, edited by A. Meyers, Springer, Heidelberg, Vol. 3, 2473-2496.
- Bormann, P., and Saul, J. (2009b). A fast, non-saturating magnitude estimator for great earthquakes, *Seism. Res. Lett.* **80** (5), 808-816; doi: 10.1785/gssrl.80.5.808.
- Bormann, P., Liu, R., Ren, X., Gutdeutsch, R., Kaiser, D., Castellaro, S. (2007). Chinese national network magnitudes, their relation to NEIC magnitudes, and recommendations for new IASPEI magnitude standards. *Bull. Seism. Soc. Am.*, **97**, 114-127.
- Bormann, P., Liu, R., Xu, Z., Ren, K., Zhang, L., Wendt S. (2009). First application of the new IASPEI teleseismic magnitude standards to data of the China National Seismographic Network, *Bull. Seism. Soc. Am.*, **99** (3), 1868-1891; doi: 10.1785/0120080010
- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bossu, R. (1996). Etude de la sismicité intraplaque de la région de Gazli (Ouzbékistan) et localisation de la déformation sismique. *PhD Thesis*, University Joseph Fourier, Grenoble

- Bossu, R., and Grasso, J. R. (1996). Stress analysis in the intraplate area of Gazli, Uzbekistan, from different sets of earthquake focal mechanisms. *J. Geophys. Res.*, **101**, 17645-17659.
- Bouchon, M. (1982). The complete synthesis of seismic crustal phases at regional distances, *J. Geophys. Res.* **78**, 1735-1741.
- Bouchon, M., and Vallée, M. (2003). Observation of long supershear rupture during the magnitude 8.1 Kunlunshan earthquake. *Science*, **301**, 824-826.
- Braunmiller, J., Deichmann, N., Giardini, D., Wiemer, St., and the SED Magnitude Working Group (2005). Homogeneous moment-magnitude calibration in Switzerland. *Bull. Seism. Soc. Am.*, **95**, 1, 58-74.
- Brillinger, D. R., Udias, A., and Bolt, B. A. (1980). A probability model for regional focal mechanism solutions. *Bull. Seism. Soc. Am.*, **70**, 149-170.
- Brune, J. N. (1970). Tectonic stress and the spectra of shear waves from earthquakes. *J. Geophys. Res.*, **75**, 4997-5009.
- Brune, J. N. (1971). Correction. *J. Geophys. Res.*, **76**, 5002.
- Brune, J. N., and Engen, G. R. (1969). Excitation of mantle Love waves and definition of mantle wave magnitude. *Bull. Seism. Soc. Am.*, **59**, 923-933.
- Bufo, E., and Udias, A. (1984). An algorithm for focal mechanism determination using signs of first motion of P, SV, SH waves. *Rev. Geofisica*, **40**, 11-26.
- Bullen, K. E., and Bolt, B. A. (1985). An introduction to the theory of seismology. Fourth Edition, *Cambridge University Press*, 4<sup>th</sup> revised ed., xvii+499 pp., ISBN: 0-521-28389-2, 451-454, 499 pp..
- Bune, V. J., Vvedenskaya, N. A., Gorbunova, J. V., Kondorskaya, N. V., Landyrev, N. S., and Fedorova, J. V. (1970). Correlation of  $M_{LH}$  and  $M_{PV}$  by data of the network of seismic stations of USSR. *Geophys. J. R. astr. Soc.*, **19**, 533-542.
- Castellaro, S., and Bormann, P. (2007). Performance of different regression procedures on the magnitude conversion problem. *Bull. Seism. Soc. Am.*, **97**, 1167-1175.
- Castellaro, S., Mulargia, F., and Kagan, Y. Y. (2006). Regression problems for magnitudes. *Geophys. J. Int.*, **165**, 913-930.
- Castello, B., Olivieri, M., and Selvaggi, G. (2007). Local and duration magnitudes determination for the Italian earthquake catalog, 1981-2002. *Bull. Seism. Soc. Am.*, **97**, 128-139.
- Chávez, D. E., and Priestley, K. F. (1985). ML observations in the Great Basin and Ms versus ML relationships for the 1980 Mammoth Lakes, California, earthquake sequence. *Bull. Seism. Soc. Am.*, **75**, 6, 1583-1598.
- Chen, P., and Chen, H. (1989). Scaling law and its applications to earthquake statistical relations. *Tectonophysics*, **166**, 53-72.
- Chinnery, M. A. (1969). Earthquake magnitude and source parameters. *Bull. Seism. Soc. Am.*, **59**, 5, 1969-1982.
- Choy, G. L. (2012). IS 3.5: Stress conditions inferable from modern magnitudes: development of a model of fault maturity, 10 pp., DOI: 10.2312/GFZ.NMSOP-2\_IS\_3.5; In: Bormann, P. (Ed.) (2012). *New Manual of Seismological Observatory Practice (NMSOP-2)*, IASPEI, GFZ German Research Centre for Geosciences, Potsdam; <http://nmsop.gfz-potsdam.de>; DOI: 10.2312/GFZ.NMSOP-2.
- Choy, G. L., and Boatwright, J. L. (1995). Global patterns of radiated seismic energy and apparent stress. *J. Geophys. Res.*, **100**, B9, 18,205-18,228.



- Choy, G. L., and Kirby, S. H. (2004). Apparent stress, fault maturity and seismic hazard for normal-fault earthquakes at subduction zones, *Geophys. J. Int.*, **159**, 991-1012.
- Choy, G. L., and Boatwright, J. (2007). The energy radiated by the 26 December 2004 Sumatra-Andaman earthquake estimated from 10-minute P-wave windows. *Bull. Seism. Soc. Am.*, **97**, 1A, S18-S24; doi: 10.1785/0120050623.
- Choy, G. L., McGarr, A., Kirby, S. H., and Boatwright, J. (2006). An overview of the global variability in radiated energy and apparent stress; in: Abercrombie R. , McGarr, A., and Kanamori, H. (eds): Radiated energy and the physics of earthquake faulting, *AGU Geophys. Monogr. Ser.* **170**, 43-57.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1978). Homogeneous magnitude system of the Eurasian continent. *Tectonophysics*, **49**, 131-138.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1983). Homogeneous magnitude system of the Eurasian continent: S and L waves. *World Data Center A for Solid Earth*, Boulder Report SE-34.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1985). Magnitude calibration functions for a multidimensional homogeneous system of reference stations. *Tectonophysics*, **118**, 213-226.
- Christoskov, L., Kondorskaya, N. V., and Vaněk, J. (1991). Homogeneous magnitude system with unified level for usage in seismological practice. *Studia geoph. ett geod.*, **35**, 221-233.
- Cifuentes, I. L., and Silvers, P. G. (1989). Low-frequency source characteristics of the great 1960 Chilean earthquake. *J. Geophys. Res.*, **94**, No. B1, 643-663.
- Das, S., and Kostrov, B. V. (1988). Principles of earthquake source mechanics. *Cambridge University Press*.
- Das, R., Wason, H. R., and Sharma, M. L. (2011). Global regression relations for conversion of surface wave and body wave magnitudes to moment magnitude. *Nat. Hazards*, **59**, 801-810.
- Das, R., Wason, H. R., and Sharma, M. L. (2012). Magnitude conversion to unified moment magnitude using orthogonal regression relation. *J. Asian Earth Sc.*, **50**, 44-51.
- Das, R., Wason, H. R., and Sharma, M. L. (2013). General orthogonal regression relations between body-wave and moment magnitudes. *Seism. Res. Lett.*, **84**(2), 219-224; doi: 10.1785/0220120125.
- Dawers, N. H., Anders, M. H., and Scholz, C. H. (1993). Growth of normal faults: Displacement-length scaling. *Geology*, **21**, 1107-1110.
- Der, Z. A., McElfresh, Th. W., and O'Donnell, A. (1982). An investigation of the regional variations and frequency-dependence of anelastic attenuation in the mantle under the United States in the 0.5-4 Hz band. *Geophys. J. R. astr. Soc.*, **69**, 68-99.
- Dewey, J. W., Cantavella Nathal, J.V., and Wendt, S. (2011). Testing of USGS/NEIC automatic procedures for computing IASPEI standard magnitudes. Poster presented at the IUGG Meeting 2011 in Melbourne.
- Di Giacomo, D., and Bormann, P. (2011). Earthquake energy. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 233-236; doi: 10.1007/978-90-481-8702-7.
- Di Giacomo, D., Grosser, H., Parolai, S., Bormann P., and Wang, R. (2008). Rapid determination of  $M_e$  for strong to great shallow earthquakes. *Geophys. Res. Lett.*, **35**, L10308; doi:10.1929/2008GL033505.



- Di Giacomo, D., Parolai, S., Bormann P., Grosser, H., Saul, J., Wang, R., and Zschau, J. (2010a). Suitability of rapid Energy magnitude estimations for emergency response purposes. *Geophys. J. Int.*, **180**, 361-374; doi: 10.1111/j.1365-246X.2009.04416.x.
- Di Giacomo, D., Parolai, S., Bormann P., Grosser, H., Saul, J., Wang, R., and Zschau, J. (2010b). Erratum to “Suitability of rapid Energy magnitude estimations for emergency response purposes.” *Geophys. J. Int.*, **181**, 1725-1726; doi: 10.1111/j.1365-246X.2010.04610.x.
- Di Giacomo, D., Harris, J., Villaseñor, A., Storchak, D. A., Engdahl, E. R., Lee, W. H. K., and the Data Entry Team (2013a). ISC-GEM: I. Data collection from early instrumental seismological bulletins. *Phys. Earth Planet. Inter.*, submitted.
- Di Giacomo, D., Bondár, I., Storchak, D. A., Engdahl, E., R., Bormann, P., and Harris, J. (2013b). ISC-GEM: Global Instrumental Earthquake Catalogue (1900-2009), III. Re-computed  $M_S$  and  $m_b$ , proxy  $M_W$ , final magnitude composition and completeness assessment. *Phys. Earth Planet. Int.*, submitted.
- Dindi, E. J., Havskov, J., Iranga, M., Jonathan, E., Lombe, D. K., Mamo, A., and Tutyomurugyendo, G. (1995). *Bull. Seism. Soc. Am.*, **85**, 354-360.
- Doornbos, D. J. (Ed.) (1988). Seismological algorithms, Computational methods and computer programs. *Academic Press*, New York, xvii + 469 pp.
- Duda, S. (1965). Secular seismic energy release o the circum-Pacific belt. *Tectonophysics*, **2**, 409-452,
- Duda, S. J. (1986). The spectra and magnitudes of earthquakes. In: Buttkus, B. (Ed.). Ten years of the Gräfenberg Array: Defining the frontiers of broadband seismology. *Geologisches Jahrbuch Reihe E*, Heft **35**, 71-79.
- Duda, S. J., and Kaiser, D. (1989). Spectral magnitudes, magnitude spectra, and earthquake quantification: the stability issue of the corner period and of the maximum magnitude for a given earthquake. *Tectonophysics*, **166**, 205-219.
- Duda, S. J., and Yanovskaya, T. B. (1993). Spectral amplitude-distance curves for P-waves: effects of velocity and Q-distribution. *Tectonophysics*, **217**, 255-265.
- Dziewonski, A. M., and Anderson, D. L. (1981). Preliminary reference Earth model. *Phys. Earth Planet. Inter.*, **25**, 297-356.
- Dziewonski, A. M., Chou, T. A., and Woodhouse, J.H. (1981). Determination of earthquake source parameters from waveform data for studies of global and regional seismicity. *J. Geophys. Res.*, **86**, 2825-2852.
- Eaton, J. P. (1992). Determination of amplitude and duration magnitudes and site residuals from short-period seismographs in Northern California. *Bull. Seism. Soc. Am.*, **82**, 2, 533-579.
- Ebel, J. E. (1982).  $M_L$  measurements for northeastern United States earthquakes. *Bull. Seism. Soc. Am.*, **72**, 1367-1378.
- Ebel, J. E. (1994). The  $M_{Lg}(f)$  magnitude scale: A proposal for its use for northeastern North America. *Seism. Res. Lett.*, **65**, 2, 157-166.
- Ekström, G., and Dziewonski, A. M. (1988). Evidence of bias in estimations of earthquake size. *Nature*, **332**, 319-323.
- Ekström, G., Nettles, M., and Dziewonski, A.M. (2012). The global CMT project 2004–2010: Centroid-moment tensors for 13,017 earthquakes, *Phys. Earth Planet. Inter.*, 200-201, 1-9.

- Engdahl, E. R., and Gunst, R. H. (1966). Use of a high speed computer for the preliminary determination of earthquake hypocenters. *Bull. Seism. Soc. Am.*, **56**, 325-336.
- Eshelby, J. D. (1969). The elastic field of a crack extending non-uniformly under general anti-plane loading. *J. Mech. Phys. Solids*, **8**, 100-104.
- Eshelby, J. (1957). The determination of the elastic field of an ellipsoidal inclusion and related problems. *Proc. R. Soc. London A*, 241, 376–396.
- Espinoza-Aranda, H. M., Jiminez, A., Ibarrola, G., Alcantar, F., Aguilar, A. et al. (1995): Mexico City seismic alert system. *Seism. Res. Lett.*, 66, 42-53.
- Espinoza-Aranda, H. M., and Rodriguez, F. H. (2003). The seismic Alert System of Mexico City. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part B. Academic Press, Amsterdam, 1253-1259.
- Ewing, M., Jardetzky, W. S., and Press, F. (1957). *Elastic Waves in Layered Media*. McGraw-Hill, New York, 358 pp..
- Fedotov S. A. (1963). On attenuation of transverse seismic waves in the upper mantle and on energy classification of near earthquakes with intermediate depth. *Izvestiya Acad. Sci. USSR, Sect. geophys.*, No.6.
- Fedotov, S. A. (1972). Energeticheskaya klassifikatsiya kurilo-kamchatskih zemletryasenii i problema magnitud (Energy classification of Kuril-Kamchatka earthquakes and the problem of magnitudes.) Moscow, *Nauka*, 116 pp.
- Fukao, Y. (1979). Tsunami earthquakes and subduction processes near deep-sea trenches. *J. Geophys. Res.*, **84**, 2303–2314.
- Fukushima, Y. (1996). Scaling relations for strong ground motion prediction models with  $M^2$  terms. *Bull. Seism. Soc. Am.*, **86**, 2, 329-336.
- Frankel, A. (1994). Implications of felt area-magnitude relations for earthquake scaling and the average frequency of perceptible ground motion. *Bull. Seism. Soc. Am.*, **84**, 2, 462-465.
- Galanopolous, A. G. (1961). On magnitude determination by using macroseismic data. *Ann. Geofis.*, **14**, 225-253.
- Gasperini, P., and Lolli, B. (2013). Comment on ‘Magnitude conversion problem using general orthogonal regression’ by H. R. Wason, Ranjit Das and M. L. Sharma, *Geophys. J. Int.*, **190**, 1091-1096 (in press).
- Gasperini, P., Bernardini, F., Valensise, G., and Boschi, E. (1999). Defining seismogenic sources from historical earthquakes felt reports. *Bull. Seism. Soc. Am.*, **89**, 94-110.
- Gasperini, P., Manfredi, G., and Zschau, J. (2007). *Earthquake early warning systems*. Springer Berlin-Heidelberg, 349 pp.
- Gasperini, P., Lolli, B., Vannucci, G., and Boschi, E. (2012). A comparison of moment magnitude estimates for the European–Mediterranean and Italian regions. *Geophys. J. Int.*, **190**, 1733–1745; doi: 10.1111/j.1365-246X.2012.05575.x
- Gasperini, P., Lolli, B., and G. Vannucci (2013a). Empirical calibration of local magnitude data sets versus moment magnitude in Italy. *Bull. Seism. Soc. Am.*, **103**, 4, doi: 10.1785/0120120356 (in press).
- Gasperini, P., Lolli, B., and Vannucci, G. (2013b). Body wave magnitude  $m_b$  is a good proxy of moment magnitude  $M_w$  for small earthquakes ( $m_b < 4.5$ -5.0). *Seism. Res. Lett.*, (in press).

- Geller, R. J. (1976). Scaling relations for earthquake source parameters and magnitudes. *Bull. Seism. Soc. Am.*, **66**, 1501-1523.
- Geller, R. J., and Kanamori, H. (1977). Magnitudes of great shallow earthquakes from 1904 to 1952. *Bull. Seism. Soc. Am.*, **67** (3), 587-598.
- Geller, R. J., and Mueller, C. S. (1980). Four similar earthquakes in central California. *Geophys. Res. Lett.*, **7**, 821-824.
- Gephart, J. W. (1990). FMSI: Focal mechanism stress inversion package using earthquake fault mechanisms and fault/slickenside data. *Computer and Geosciences*, **16**(7), 953-989.
- Gephart, J. W., and Forsyth, D. W. (1984). An improved method for determining the regional stress tensor using earthquake focal mechanism data: application to the San Fernando earthquake sequence. *J. Geophys. Res.*, **89**, 9305-9320.
- Giardini, D. (1988). Frequency distribution and quantification of deep Earthquakes. *J. Geophys. Res.*, **93**, 2095-2105.
- Goldstein, P., and Archuleta, R. I. (1991). Deterministic frequency-wavenumber methods and direct measurement of rupture during earthquakes using a dense array-data analysis. *J. Geophys. Res.*, **96**, 6187-6198.
- Gordon, D. W. (1971). Surface-wave versus body-wave magnitude. *Earthquake Notes*, **42**, 3-4, 20-28.
- Granville, J. P., Kim, W.-Y., and Richards, P. G. (2002). An assessment of seismic body-wave magnitudes published by the Prototype International Data Centre. *Seism. Res. Lett.*, **73**(6), 893-906.
- Granville, J. P., Richards, P. G., Kim, W.-Y., and Sykes, L. R. (2005). Understanding the differences between three teleseismic  $m_b$  scales. *Bull. Seism. Soc. Am.*, **95**(5), 1809-1824.
- Grasso, J. R., and Wittlinger, G. (1990). Ten years of seismic monitoring a gas field. *Bull. Seism. Soc. Am.*, **80**, 2, 450-479.
- Grecksch, G., and Kümpel, H.-J. (1997). Statistical analysis of strong-motion accelerograms and its application to earthquake early-warning systems. *Geophys. J. Int.*, **129**, 113-123.
- Greenhalgh, S. A., and Singh, R. (1986). A revised magnitude scale for South Australian earthquakes. *Bull. Seism. Soc. Am.*, **76**, 3, 757-769.
- Greenhalgh, S. A., Denham, D., McDougall, R., and Rynn, J. M. (1989). Intensity relations for Australian earthquakes. *Tectonophysics*, **166**, 255-267.
- Griggs, D., and Press, F. (1961). Probing the earth with nuclear explosions. *J. Geophys. Res.*, **66**(1) 237-258.
- Gross, S. (1996). Magnitude distributions and slip scaling of heterogeneous seismic sources. *Bull. Seism. Soc. Am.*, **86** (2), 498-504.
- Grünthal, G. (Ed.) (1998). European Macroseismic Scale 1998, *Cahiers du Centre Européen de Gèodynamique et de Seismologie*, **15**, Conseil de l'Europe, Luxembourg, 99 pp.
- Grünthal, G. (2011). Earthquake intensity. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 237-242; doi: 10.1007/978-90-481-8702-7.,
- Grünthal, G., and Wahlström, R. (2003). An Mw based earthquake catalogue for central, northern and northwestern Europe using a hierarchy of magnitude conversions. *J. Seismol.*, **7** (4), 507-531.
- Grünthal, G., Wahlström, R., and Stromeyer, D. (2009). Harmonization check of Mw within the central, northern, and northwestern European earthquake catalogue (CENEC). *J. Seismology*, **13**(4), 613-623.

- Gupta, H. K., and Rastogi, B. K. (1976). *Dams and earthquakes*. Elsevier.
- Gusev, A. A. (1989). Multiasperity fault model and the nature of short-period subsources. *Pure Appl. Geophys.*, **130**, 635-660.
- Gusev, A. A. (1991). Intermagnitude relationships and asperity statistics. *Pure Appl. Geophys.*, **136**, 515-527.
- Gusev, A. A., and Melnikova, V. N. (1992). Relationships between magnitude scales for global and Kamchatkan earthquakes. *Volc. Seism.*, **12**, 723-733. (English translation of the original Russian publication: Гусев А. А., Мельникова, В. Н. (1990). Связи между магнитудами - среднемировые и для Камчатки. *Вулканология и сейсмология*, №6, 55-63.
- Gusev, A. A., Guseva, E. M., and Panza, G. F. (2007). Size and duration of the high-frequency radiator in the source of the 2004 December 26 Sumatra earthquake. *Geophys. J. Int.*, **170**, 1119-1128; doi: 10.1111/j.1365-246X.2007.03368.x.
- Gutdeutsch, R., Kaiser, D., and Jentzsch, G. (2002). Estimation of earthquake magnitudes from epicentral intensities and other focal parameters in central and southern Europe. *Geophys. J. Int.*, **151** (3), 824-834.
- Gutdeutsch, R., Castellaro, S., and Kaiser, D. (2011). The magnitude conversion problem: Further insights. *Bull. Seism. Soc. Am.*, **101**, 379-384; doi: 10.1785/0120090365..
- Gutenberg, B. (1945a). Amplitudes of surface waves and magnitudes of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 3-12.
- Gutenberg, B. (1945b). Amplitudes of P, PP, and S and magnitude of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 57-69.
- Gutenberg, B. (1945c). Magnitude determination of deep-focus earthquakes. *Bull. Seism. Soc. Am.*, **35**, 117-130.
- Gutenberg, B. (1956). The energy of earthquakes. *Q. J. Geol. Soc. London*, **112**, 1-14.
- Gutenberg, B., and Richter, C. F. (1934). On seismic waves (first paper). *Gerl. Beitr. Geophysik*, **43**, 56 – 133.
- Gutenberg, B., and Richter, C. F. (1936). On seismic waves (third paper). *Gerl. Beitr. Geophysik*, **47**, 73-131.
- Gutenberg, B., and Richter, C. F. (1942). Earthquake magnitude, intensity, energy and acceleration. *Bull. Seism. Soc. Am.*, **32**, 163-191.
- Gutenberg, B., and Richter, C. F. (1954). Seismicity of the Earth and associated Phenomena. 2<sup>nd</sup> edition, Princeton University Press, 310 pp.
- Gutenberg, B., and Richter, C. F. (1956a). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Gutenberg, B., and Richter, C. F. (1956b). Earthquake magnitude, intensity, energy and acceleration. *Bull. Seism. Soc. Am.*, **46**, 105-145.
- Habermann, R. E. (1995). Opinion. *Seism. Res. Lett.*, **66**, 5, 3.
- Hanks, T. C. (1977). Earthquake stress drops, ambient tectonic stress, and the stresses that drive plate motion. *Pure Appl. Geophys.*, **115**, 441-458.
- Hanks, T. C. and Kanamori, H. (1979). Moment magnitude. *J. Geophys. Res.*, **84**, 2348-2350.
- Hanks, T. C., and Johnston, A. C. (1992). Common features of the excitation and propagation of strong ground motion for North American earthquakes. *Bull. Seism. Soc. Am.*, **82**, 1-23.

- Hanks, T. C., Hileman, J. A., and Thatcher, W. (1975). Seismic moments of the larger earthquakes of the southern California region. *Soc. Bull. Geol.*, **86**, 1131-1139.
- Hansen, R. A., Ringdal, F., and Richards, P. G. (1990). The stability of RMS Lg measurements and their potential for accurate estimation of the yields of Soviet underground nuclear explosions. *Bull. Seism. Soc. Am.*, **80**(6), 2006-2126.
- Hara, T. (2007a). Measurement of the duration of high-frequency energy radiation and its application to determination of the magnitudes of large shallow earthquakes. *Earth Planets Space* **59**, 227-231.
- Hara, T. (2007b). Magnitude determination using duration of high frequency radiation and displacement amplitude: application to tsunami earthquakes. *Earth. Planet Space* **59**, 561-565.
- Hartzel, S. H., and Heaton, T. (1985). Teleseismic time functions for large, shallow subduction zone earthquakes. *Bull. Seism. Soc. Am.*, **75**(4), 965-1004.
- Haskell, N. A. (1964). Total energy and energy spectral density of elastic wave radiation from propagating faults. *Bull. Seism. Soc. Am.*, **54**, 1811-1841.
- Haskell, N. A. (1966). Total energy and energy spectral density of elastic wave radiation from propagating faults. Part II. A statistical fault model. *Bull. Seism. Soc. Am.*, **56**, 125-140.
- Hatzidimitriou, P., Papazachos, C., Kiratzi, A., and Theodulidis, N. (1993). Estimation of attenuation structure and local earthquake magnitude based on acceleration records in Greece. *Tectonophysics*, **217**, 243-253.
- Havskov, J. (Ed.) (1996). The SEISAN earthquake analysis software for the IBM PC and SUN, Version 5.2. *Institute of Solid Earth Physics, Univ. of Bergen*, August 1996.
- Havskov, J., and Macias (1983). A coda-length magnitude scale for some Mexican stations. *Geofosica Int.*, **22**, 205-213.
- Havskov, J., and Ottemöller, L. (1999). Electronic Seismologist - SeisAn Earthquake Analysis Software. *Seism. Res. Lett.*, **70**, 532-534.
- Havskov, J., and Sorensen, M. B. (2006). New coda magnitude scales for mainland Norway and the Jan Mayen region. Norwegian National Seismic Network, *Technical Report No. 19, University of Bergen*.
- Havskov, J., and Ottemöller, L. (2010). *Routine data processing in earthquake seismology*. Springer, 347 pp.
- Hayes, G. P., Rivera, L., and Kanamori, H. (2009). Source inversion of the W-phase: Real-time implementation and extension to low magnitudes. *Seism. Res. Lett.*, **80**, 5, 817-822; doi: 10.1/85/gssrl.80.5.817.
- Helfrich, G. R. (1997). How good are routinely determined focal mechanisms? Empirical statistics based on a comparison of Harvard, USGS and ERI moment tensors. *Geophys. J. Int.*, **131**, 741-750.
- Herak, M., and Herak, D. (1993). Distance dependence of  $M_S$  and calibrating function for 20 s Rayleigh waves. *Bull. Seism. Soc. Am.*, **83**, 6, 1881-1892.
- Herak, M., Panza, G., and Costa, G. (2001). Theoretical and observed depth corrections for  $M_s$ . *Pure Appl. Geophys.*, **158**, 1517-1530.
- Herrmann, R. B. (1975). The use of duration as a measure of seismic moment and magnitude. *Bull. Seism. Soc. Am.*, **65**, 899-913.
- Herrmann, R. B., and Nuttli, O. W. (1982). Magnitude: The relation between  $M_L$  and  $m_{bLg}$ . *Bull. Seism. Soc. Am.*, **72**, 389-397.

- Herrmann, R. B., and Kijko, A. (1983). Modeling some empirical Lg relations. *Bull. Seism. Soc. Am.*, **73**, 1835-1850.
- Hirshorn, B., and Weinstein, S. (2009). Earthquake source parameters, rapid estimates for tsunami warning. In *Encyclopedia of Complexity and Systems Science*, ed. R. Meyers, Springer, Heidelberg-New York: vol. 3, 2657-2675.
- Houston, H. (1999). Slow ruptures, roaring tsunamis, *Nature* **400**, 409-410
- Horton, S. (2012). Disposal of hydrofracturing waste fluid by injections into subsurface aquifers which triggered an earthquake swarm in Central Arkansas with the potential for damaging earthquake. *Seism. Res. Lett.*, **83**, 2, 250-260.
- Houston, H., and Kanamori, H. (1986). Source spectra of great earthquakes: teleseismic constraints on rupture process and strong motion. *Bull. Seism. Soc., Am.*, **76**, 1, 19-42.
- Hunter, R. N. (1972). Use of LPZ for magnitude. In: *NOAA Technical Report ERL 236-ESL21*, J. Taggart, Editor, U.S. Dept. Commerce, Boulder, Colorado.
- Husseini, M. I. (1977). Energy balance for formation along a fault. *Geophys. J. Roy. Astron. Soc.*, **49**, 699-714.
- Hutton, L. K., and Boore, D. M. (1987). The  $M_L$  scale in Southern California. *Bull. Seism. Soc. Am.*, **77**, 6, 2074-2094.
- Hutton, L. K., and Jones, L. M. (1993). Local magnitudes and apparent variations in seismicity rates in Southern California. *Bull. Seism. Soc. Am.*, **83**, 2, 313-329.
- Hwang, L. J., and Kanamori, H. (1989). Teleseismic and strong-motion source spectra from two earthquakes in western Taiwan. *Bull. Seism. Soc. Am.*, **79**, 935-944.
- Hyvernaud, O., Reymond, D., Talandier, J., and Okal, E. A. (1993). Four years of automated measurement of seismic moments at Papeete using the mantle magnitude  $M_m$ : 1987-1991. *Tectonophysics*, **217**, 175-193.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf)
- IASPEI (2013). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)
- Iida, K. (1959). Earthquake energy and earthquake fault length. *J. Earth Sci.*, Nagoya Univ., **7**, 98-107.
- Janský, J., Ruprechtova, L., and Tittel, B. (1977). Magnitude determination based on short-period core phases. *Studia geophys. et geod.*, **21**, 267-273.
- Janský, J., and Kvasnicka, M. (1992). Theoretical PKIKP magnitude determination. *Studia geophys. et geod.*, **36**, 20-25.
- Johnston, A. C. (1993). The stable continental region data base. In: *The earthquakes of stable continental regions: Assessment of large earthquake potential. EPRI report TR-102261*, Electric Power Research Institute, Palo Alto, California.
- Johnston, A. C. (1996). Seismic moment assessment of earthquakes in stable continental regions – II. Historical seismicity. *Geophys. J. Intern.*, **125**(3), 639-678.
- Kagan, Y. Y. (2002). Seismic moment distribution revisited: I. Statistical results. *Geophys. J. Int.*, **148**, 520-541.
- Kagan, Y. Y. (2003). Accuracy of modern global earthquake catalogs. *Phys. Earth Planet. Inter.*, **135**(2-3), 173-209.

- Kanamori, H. (1972). Mechanism of tsunami earthquakes. *Phys. Planet. Earth Inter.*, **6**, 346-359.
- Kanamori, H. (1977). The energy release in great earthquakes. *J. Geophys. Res.*, **82**, 2981-2987.
- Kanamori, H. (1983). Magnitude scale and quantification of earthquakes. *Tectonophysics*, **93**, 185-199.
- Kanamori, H. (1988). The importance of historical seismograms for geophysical research. In: W. H. K. Lee (ed), *Historical seismograms and earthquakes of the world*, Academic Press, New York, 16-33.
- Kanamori, H. (2001). Energy budget of earthquakes and seismic efficiency. Chapter 11 in: Teisseyre, R. and Majenski, E. (Eds): *Earthquake thermodynamics and phase transformations in the Earth Interior*, Academic Press, Internat. Geophys. Series, **76**, 293-305.
- Kanamori, H. (2005). Real-time seismology and earthquake damage mitigation. *Annu. Rev. Earth Planet. Sci.*, **33**, 195-214; doi: 10.1146/annurev.earth.33.092203.122626.
- Kanamori, H. (2006). The radiated energy of the 2004 Sumatra-Andaman earthquake; In: Abercrombie R, McGarr A, Kanamori H (eds): Radiated energy and the physics of earthquake faulting, *AGU Geophys. Monogr. Ser.*, **170**, 59-68.
- Kanamori, H., and Cipar, J. J. (1974). Focal process of the great Chilean earthquake May 22, 1960. *Phys. Earth Planet. Inter.*, **9**, 128-136.
- Kanamori, H., and Anderson, D. L. (1975). Theoretical basis of some empirical relations in seismology. *Bull. Seism. Soc. Am.*, **65**, 1073-1095.
- Kanamori, H., and Jennings, P. C. (1978). Determination of local magnitude,  $M_L$ , from strong motion accelerograms. *Bull. Seism. Soc. Am.*, **68**, 471-485.
- Kanamori, H., and Given, J. (1981). Use of long-period surface waves for rapid determination of earthquake-source parameters. *Phys. Earth Planet. Inter.*, **27**, 8-31.
- Kanamori, H., and Kikuchi, M. (1993). The 1992 Nicaragua earthquake: a slow tsunami earthquake associated with subducted sediments. *Nature*, **361**, 714-716.
- Kanamori, H., and Kikuchi, M. (1995). Source characteristics of the 1992 Nicaragua tsunami earthquake inferred from teleseismic body waves. *Pure Appl. Geophys.*, **144**, 441-453.
- Kanamori, H., and Brodsky, E. E. (2004). The physics of earthquakes. *Rep. Prog. Phys.*, **67**, 1429-1496; doi: 10.1088/0034-4885/67/8/R03.
- Kanamori, H., and Helmberger, D. (2005). Energy radiation from the Sumatra earthquake. *Nature*, **434**, p. 582.
- Kanamori, H., and Rivera, L. (2008). Source inversion of W phase: speeding up seismic tsunami warning. *Geophys. J. Int.*, **175**, 222-238.
- Kanamori, H., Mori, J., Hauksson, E., Heaton, Th. H., Hutton, L. K., and Jones, L. M. (1993). Determination of earthquake energy release and  $M_L$  using TERRASCOPE. *Bull. Seism. Soc. Am.*, **83**, 2, 330-346.
- Kanamori, H., Hauksson, E., and Heaton, T. (1997) Real-time seismology and earthquake hazard mitigation. *Nature*, **390**: 461-464
- Karnik, V. (1956). Magnitudenbestimmung europäischer Nahbeben. (In German). Travaux Inst. Géophys. Acad. Tchecosl. Sci., No. **64**.
- Karnik, V. (1969). Seismicity of the European area. Part I. *Reidel Publishing Company*, Dordrecht, 364 pp.

- Karnik, V. (1972). Differences in magnitudes. Vorträge des Soproner Symposiums der 4. Subkommission von KAPG 1970, Budapest, 69-80.
- Karnik, V., Kondorskaya, N. V., Riznichenko, Yu. V., Savarensky, Ye. F., Soloviev, S. L., Shebalin, N. V., Vaněk, J., and Zatopek, A. (1962). Standardisation of the earthquake magnitude scales. *Studia Geophysica et Geodaetica*, **6**, 41-48.
- Katsumata, A. (1964). A method to determine the magnitude of deep focus earthquakes in and near Japan. (in Japanese), *Zisin*, II, vol. 17, 158-165.
- Katsumata, A. (1996). Comparison of magnitudes estimated by the Japan Meteorological Agency with moment magnitudes for intermediate and deep earthquakes. *Bull. Seism. Soc. Am.*, **86**(3), 832-842.
- Kaverina, A. N., Lander, A. V., and Prozorov, A. G. (1996). Global creepex distribution and its relation to earthquake-source geometry and tectonic origin. *Geophys. J. Int.*, **135**, 249-265.
- Kawasumi, H. (1951). Measures of earthquake danger and expectancy of maximum intensity throughout Japan as inferred from the seismic activity in historical times. *Bull. Earthq. Res. Inst.*, **29**, 469-482.
- Keilis-Borok, V. I., (1959). On the estimation of displacement in an earthquake source and of source dimension. *Ann. Geofis.*, **12**, 205-214.
- Kennett, B. L. N. (1985). On regional S, *Bull. Seism. Soc. Am.*, **75**, 1077-1086.
- Khromovskikh, V. S. (1989). Determination of magnitudes of ancient earthquakes from dimensions of observed seismodislocations. *Tectonophysics*, **166**, 269-280.
- Kikuchi, M., and Kanamori, H. (1982). Inversion of complex body waves. *Bull. Seism. Soc. Am.*, **72**, 491-506.
- Kikuchi, M., and Kanamori, H. (1986). Inversion of complex body waves – II. *Phys. Earth Planet. Int.*, **43**, 205-222.
- Kikuchi, M., and Fukao, Y. (1987). Inversion of long-period P-waves from great earthquakes along subduction zones. *Tectonophysics*, **144**, 231-247.
- Kikuchi, M., and Fukao, Y. (1988). Seismic wave energy inferred from long-period body wave inversion. *Bull. Seism. Soc. Am.*, **78**, 5, 1707-1724.
- Kikuchi, M., and Ishida, M. (1993). Source retrieval for deep local earthquakes with broadband records. *Bull. Seism. Soc. Am.*, **83**, 6, 1855-1870.
- Kim, W.-Y. (1998). The  $M_L$  scale in Eastern North America. *Bull. Seism. Soc. Am.*, **88**, 4, 935-951.
- Kim, W.-Y., Wahlström, R., and Uski, M. (1989). Regional spectral scaling relations of source parameters for earthquakes in the Baltic Shield. *Tectonophysics*, **166**, 151-161.
- King, G. C. P. (1978). Geological faults, fractures, creep and strain. *Phil. Trans. R. Soc. London, A*, **288**, 197-212.
- Kiratzi, A. A., and Papazachos, B. C. (1985). Local Richter magnitude and total signal duration in Greece. *Ann. Geophys.*, **3**, 4, 531-537.
- Klein, F. W. (1978). Hypocenter location program HYPOINVERSE. *U.S. Geol. Surv. Open-File Report*. **78-694**.
- Klein, F. W. (1985). HYPOINVERSE, a program for VAX and professional 350 computers to solve the earthquake locations. *U.S. Geological Survey Open-File Report* **85-515**, 53 pp.
- Klein, F. W. (2002). User's guide to HYPOINVERSE-2000, a Fortran program to solve for earthquake locations and magnitudes. *Open File Report 02-171*, U.S. Geological Survey.



- Klinge, K. (1989). Duration magnitudes. In: Bormann, P. (ed.). *Monitoring and analysis of the earthquake swarm 1985/86 in the region Vogtland/Western Bohemia*. Akademie der Wissenschaften der DDR, ZIPE Veröffentlichung Nr. 110, 109-114.
- Knopoff, L., Schwab, F., and Kansel, E. (1973). Interpretation of Lg. *Geophys. J. R. Astr. Soc.*, **33**, 389-404.
- Kostrov, B.V. (1966). Unsteady propagation of longitudinal shear cracks. *J. Appl. Math. Mech.*, **30**, 1241-1248.
- Kostrov, B. (1974). Seismic moment and energy of earthquakes, and seismic flow of rock. *Izv. Acad. Sci., USSR, Phys. Solid Earth* (Engl. Transl.), **1**, 23-40.
- Kowalle, G., Tittel, B., and Bormann, P. (1983). Determination of a magnitude function using short-period readings of PKP. *Tectonophysics*, **93**, 289-294.
- Koyama, J. (1985): Earthquake source time-function from coherent and incoherent rupture.. *Tectonophysics*, **118**(3-4), 227-242.
- Koyama, J., and Zheng, S.-H. (1985). Excitation of short-period body-waves by great earthquakes. *Physics Earth Planet. Int.*, **37**, 108-123.
- Koyama, J., and Shimada, N. (1985). Physical basis of earthquake magnitudes: an extreme value of seismic amplitudes from incoherent fracture of random fault patches. *Phys. Earth Planet. Int.*, **40**, 301-308.
- Koyama, J., Takemura, M., and Suzuki, Z. (1982). A scaling model for quantification of earthquakes in and near Japan. *Tectonophysics*. **84**: 3-16.
- Krüger, F., and Ohrnberger, M. (2005a). Tracking the rupture of the  $M_w = 9.3$  Sumatra earthquake over 1,150 km at teleseismic distances. *Nature*, **435**, 937-939; doi: 10.1038/nature03696.
- Krüger, F., and M. Ohrnberger (2005b). Spatio-temporal source characteristics of the 26 December 2004 Sumatra earthquake as imaged by teleseismic broadband arrays. *Geoph. Res. Lett.*, **32**, L24312, doi:10.1029/2005GL023939.
- Kwiatek, G., Plenkens, K., Nakatani, M., Yabe, Y., Dresen, G., and JAGUARS-Group (2010). Frequency-magnitude characteristics down to magnitude -4.4 for induced seismicity recorded at Mponeng gold mine, South Africa. *Bull. Seism. Soc. Am.*, **100**(3), 1165-1173 ; doi: 10.1785/0120090277.
- Krystek, M., and Anton, M. (2007). A weighted total least-squares algorithm for fitting a straight line. *Meas. Sci. Tech.*, **22**, 3438-3442.
- Lahr, J. C. (1989). HYPOELLIPSE/Version 2.0: A computer program for determining local earthquakes hypocentral parameters, magnitude, and first motion pattern. *U.S. Geological Survey Open-File Report 89-116*, 92 pp.
- Lahr, J. C. (2003). The HYPOELLIPSE earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*. Academic Press, Amsterdam, 1617-1618.
- Langston, C. A. (1981). Source inversion of seismic waveforms: The Koyna, India, earthquakes of 13 September 1967. *Bull. Seism. Soc. Am.*, **71**, 1-24.
- Langston, C. A., Brazier, R., Nyblade, A. A., and Owens, T. J. (1998). Local magnitude scale and seismicity rate for Tanzania, East Africa. *Bull. Seism. Soc. Am.*, **88**, 3, 712-721.
- Latter, A. L., Le Levier, R. E., Martinelli, E. A., and McMillan, W. C. (1961a). Method of concealing underground nuclear explosions. *J. Geophys. Res.*, **66**, 943-946.

- Latter, A. L., Martinelli, E. A., Mathews, J., and McMillan, W. C. (1961b). The effect of plasticity on decoupling of underground explosions. *J. Geophys. Res.*, **66**, 2929-2937.
- Lay, T., and Wallace, T. C. (1995). Modern global seismology. ISBN 0-12-732870-X, Academic Press, 521 pp.
- Lay, T., and Bilek, S. (2007). Anomalous earthquake ruptures at shallow depths on subduction zone megathrusts. In: T. H. Dixon and C. Moore (Eds.), *The seismogenic Zone of subduction thrust faults*, Columbia Univ. Press, New York, 692 pp ISBN:978-0-231-13866-6.
- Lazareva, A. P., and Yanovskaya, T. B. (1975). The effect of the lateral velocity on the surface wave amplitudes. Proc. Intern. Symp. Seismology and Solid-Earth Physics, Jena, April 1-6, 1974. *Veröff. Zentralinstitut für Physik d. Erde*, **No. 31**, Vol. 2, 433-440.
- Lee, V., Trifunac, M., Herak, M., Živčić, M., and Herak, D. (1990).  $M_L^{SM}$  computed from strong motion accelerograms recorded in Yugoslavia. *Earthq. Engin. Structur. Dyn.*, **19**, 1167-1179.
- Lee, W. H. K. (1995). Realtime seismic data acquisition and processing. *IASPEI Software Vol.1*, 2<sup>nd</sup> Edition.
- Lee, W.H.K. (2013). Bibliographical search for reliable seismic soments of large earthquakes during 1900-1979 to compute direct Mw in the ISC-GEM Global Instrumental Reference Earthquake Catalogue. *Phys. Earth Planet. Inter.*, submitted.
- Lee, W. H. K., and Lahr, J. C. (1975). HYPO71 (revised): A computer program for determining hypocenter, magnitude and first motion pattern of local earthquakes. U.S. Geological Survey Open-File Report 75-311, 116 pp.
- Lee, W. H. K., Bennet, R., and Meagher, K. (1972). A method of estimating magnitude of local earthquakes from signal duration. *U.S. Geol. Surv. Open-File Rep.*, 28 pp.
- Lee, W. H. K., and Espinoza-Aranda, J. M. (2002). Earthquake earlywarning systems: current status and perspectives. In: Zschau, J., and Küppers, A. N. (Eds.), *Early Warning Systems for Natural Disaster Reduction*, Springer, Berlin, 409-423.
- Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, section with articles on earthquake geology and mechanics, 455-661.
- Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B. Academic Press, Amsterdam, section with articles on earthquake geology and mechanics.
- Lemzikov, V. K., and A. A. Gusev (1991). Coda-based energy classification of near Kamchatka earthquakes. *Volc. Seis.*, **11**, 558-578. Englisch translation of the original publication in Russian: Лемзиков В.К., Гусев А.А.(1989). Энергетическая классификация близких камчатских землетрясений по уровню кода-волн. *Вулканология и сейсмология*, №4., 83-97.
- Lienert, B. R. E. (1991). Report on modifications made to Hypocenter. *Institute of Solid Earth Physics, University of Bergen*.
- Lienert, B. R. E., Berg, E., and Frazer, L. N. (1988). HYPOCENTER: An earthquake location method using centered, scaled, and adaptively least squares. *Bull. Seism. Soc. Am.*, **76**, 771-783.
- Lienert, B. R. E., and Havskov, J. (1995). A computer program for locating earthquakes both locally and globally. *Seism. Res. Lett.*, **66**, 26-36.

- Lienkaemper, J. J. (1984). Comparison of two surface-wave magnitude scales: M of Gutenberg and Richter (1954) and  $M_s$  of "Preliminary Determination of Epicenters". *Bull. Seism. Soc. Am.*, **74**(6), 2357-2378.
- Lilwal, R. C. (1987). Empirical amplitude-distance-depth curves for short-period P waves in the distance range 20 to 180°. *AWRE Report No. O 30/86*, HMSO, London
- Liu, H.-P., Anderson, D.L., and Kanamori, H. (1976). Velocity dispersion due to anelasticity; implications for seismology and mantle composition. *Geophys. J. R. Astron. Soc.*, **47**, 41-58.
- Lolli, B., and Gasperini, P. (2012). A comparison among general orthogonal regression methods applied to earthquake magnitude conversions. *Geophys. J. Int.*, **190**, 1135–1151 doi: 10.1111/j.1365-246X.2012.05530.x
- Lomax, A. (2005). Rapid determination of earthquake size for hazard warning. *EOS*, **86**, 21, p. 202.
- Lomax, A., Michelini, A., and Piatanesi, A. (2007). An energy-duration procedure for rapid and accurate determination of earthquake magnitude and tsunamigenic potential, *Geophys. J. Int.*, **170**, 1195-1209; doi: 10.1111/j.1365-246X.2007.03469.x.
- Lomax, A., and A. Michelini (2009a).  $M_{wpd}$ : A duration-amplitude procedure for rapid determination of earthquake magnitude and tsunamigenic potential from P waveforms, *Geophys. J. Int.*, **176**, 200-214; doi: 10.1111/j.1365-246X.2008.03974.x.
- Lomax, A., and Michelini, A. (2009b). Tsunami early warning using earthquake rupture duration. *Geophys. Res. Lett.*, **36**, L09306; doi: 10.1029/2009GL037223.
- Lomax, A., and Michelini, A. (2011a). Tsunami early warning using earthquake rupture duration and P-wave dominant period: the importance of length and depth of faulting. *Geophys. J. Int.*, **185**(1), 283-291; doi: 10.1111/j.1365-246X.2010.04916.x.
- Lomax, A., and Michelini, A. (2011b). Erratum. *Geophys. J. Int.*, **186**, 1454, doi:10.1111/j.1365-246X.2011.05128.x.
- Lomax, A., and Michelini, A. (2012). Tsunami early warning within five minutes. *Pure Appl. Geophys.*, **169**, nnn-nnn; doi: 10.1007/s00024-012-0512-6
- Madariaga, R. (1976). Dynamics of an expanding circular fault. *Bull. Seism. Soc. Am.*, **66**, 639-666.
- Madariaga, R. (2011). Earthquakes, source theory. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, Vol. 1, 248-252; doi: 10.1007/978-90-481-8702-7.
- Madariaga, R., and Olsen, K. B. (2002). Earthquake dynamics. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 175-194.
- Mackey, K. G., Fujita, K., Hartse, H. E., Stead, R. J., Steck, L. K., Gunbina, L. V., Leyshuk, N., Shibaev, S. V., Koz'min, B. M., Imaev, V. S., Gordeev, E. I., Chebrov, V. N., Masal'ski, O. K., Gileva, N. A., Bormatov, V. A., Voitenok, A. A., Levin, Y. N., and Fokina, T. A. (2010). Seismicity Map of Eastern Russia, 1960–2010. *Seism. Res. Lett.*, **81**(5), 761-768; doi: 10.1785/gssrl.81.5.
- Margaris, B. N., and Papazachos, C. B. (1999). Moment-magnitude relations based on strong-motion records in Greece. *Bull. Seism. Soc. Am.*, **89**, 442-455.
- Marshall, P.D., and Basham, P.W. (1972). Discrimination between earthquakes and underground explosions employing an improved  $M_s$  scale, *Geophys. J. R. astr. Soc.* **28**, 431-458.

- Marrett, R. (1994). Scaling of intraplate earthquake recurrence interval with fault length and implications for seismic hazard assessment. *Geophys. Res. Lett.*, **21**, 24, 2637-2640.
- Mason, D. B. (1996). Earthquake magnitude potential of the intermountain seismic belt, USA, from surface-parameter scaling of Late Quaternary faults. *Bull. Seism. Soc. Am.*, **86**, 5, 1487-1506.
- Masuda, K. (1992). P-coda amplitudes as a measure of earthquake magnitude of local microearthquake. *J. Phys. Earth*, **40**, 565-572.
- Mayeda, K. (1993). Mb(LgCoda): A stable single station estimator of magnitude. *Bull. Seism. Soc. Am.*, **83**, 851-861.
- McGarr, A. (1991). On a possible connection between three major earthquakes in California and oil production. *Bull. Seism. Soc. Am.*, **81**, 948-979.
- McGarr, A., and Fletcher, J. B. (2002). Mapping apparent stress and energy radiation over fault zones of major earthquakes. *Bull. Seism. Soc. Am.*, **92**, 1633-1646.
- McGarr, A., Spottiswoode, S. M., and Gay, N. C. (1975). Relationship of mine tremors induced stresses and to rock properties in the focal zone. *Bull. Seism. Soc. Am.*, **65**, 4, 981-993.
- McGarr, A., Simpson, D., and Seeber, L. (2002). Case histories of induced and triggered seismicity. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*. Academic Press, Amsterdam, 647-661.
- Medvedev, S. V. (1962). Engineering seismology (in Russian). *Academy of Sciences, Inst. of Physics of the Earth*, Publ. house for literature on Civil Engineering, Architecture and Building Materials, Moscow.
- Meissner, R., and Wever, Th. (1988). Lithosphere rheology: Continental versus oceanic unit. *J. Petrology, Oxford University Press*, Special Lithosphere Issue, 53-61.
- Mendez, A. J., and Anderson, J. G. (1991). The temporal and spatial evolution of the 19 September 1985 Michoacan earthquake as inferred from near-source ground-motion records. *Bull. Seism. Soc. Am.*, **81**, 3, 844-861.
- Miyamura, S. (1974). Determination of body-wave magnitudes for shallow earthquakes in the New Zealand and Macquarie Loop regions using PKP phases. *Phys. Earth Planet. Int.*, **8**, 167-176.
- Mizone, M. (1977). On the magnitude calibration function for core phases with special reference to the PKP1-PKP2 caustic. *J. Phys. Earth*, **25**, 143-162.
- Montagner, J.-P., and Kennett, B. L. N. (1996). How to reconcile body-wave and normal-mode reference Earth models?. *Geophys. J. Int.*, **125**, 229-248.
- Moore, G. F., Bangs, N. L., Taira, A., Kuramoto, S., Pangborn, E., Tobin, H. J. (2007). Three-dimensional splay fault geometry and implications for tsunami generation. *Science*, **318**, 1128, doi:[10.1126/science.1147195](https://doi.org/10.1126/science.1147195).
- Muir-Wood, R. (1993). From global seismotectonics to global seismic hazard. *Annali di Geofisica*, **36**, 3-4, 153-168.
- Musson, R. M. W. (1996). Determination of parameters for historical British earthquakes, *Annali di Geofisica*, **39**(5), 1041-1048.
- Musson, R. M. W., and Cčić, I. (2002). Macroseismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*. Academic Press, Amsterdam, 807-822.

- Musson, R. M. W., and Jiménez, M.-J. (2008). Macroseismic estimation of earthquake parameters. *NERIES project report*, Module NA4, Deliverable D3, Edinburgh.
- Nabelek, J. (1984). Determination of earthquake source parameters from inversion of body waves. *Ph.D. Thesis*, Mass. Inst. Tech., Cambridge.
- Nakamura, Y. (1988). On the urgent earthquake detections and alarm system (UrEDAS). *9<sup>th</sup> World Conference on Earthquake Engineering*, Vol. VII, 673-678.
- Nakamura, Y., and Saita, J. (2007). UrEDAS: The earthquake warning system today and tomorrow. In: Gasparini, P., Manfredi, G., and Zschau, J. (2007). *Earthquake Early Warning Systems*. Springer, Berlin-Heidelberg, 249-281.
- Newman, A. V., and Okal, E. A. (1998). Teleseismic estimates of radiated seismic energy: The  $E/M_0$  discriminant for tsunami earthquakes. *J. Geophys. Res.*, **103**, 26,885-26,897.
- Neuman, A.V., Hayes, G., Wei, Y., and Convers, J. (2011). The 25 October 2010 Mentawai tsunami earthquake, from real-time discriminants, finite-fault rupture, and tsunami excitation, *Geophys. Res. Lett.*, **38**, L05302 ; doi:[10.1029/2010GL046498](https://doi.org/10.1029/2010GL046498).
- Ni, S., Kanamori, H., and Helmberger, D. (2005). Seismology – Energy radiation from the Sumatra earthquake. *Nature*, **434**, 582-582.
- Nolet, G., Krueger, S., and Clouser, R. M. (1998a). Empirical determination of depth-distance corrections for  $m_b$  and  $M_w$  from Global Seismograph Network stations. *Geophys. Res. Lett.*, **25**(9), 1451-1454.
- Nolet, G., Krueger, S., and Clouser, R. M. (1998b). Comment on “Empirical determination of depth-distance corrections for  $m_b$  and  $M_w$  from global seismograph network stations”. *Geophys. Res. Lett.*, **25**, 22, 4271-4272.
- Nowroozi, A. (1985). Empirical relations between magnitudes and fault parameters for earthquakes in Iran. *Bull. Seism. Soc. Am.*, **75**, 1327-1338.
- Nuttli, O. W. (1973). Seismic wave attenuation and magnitude relations for eastern North America. *J. Geophys. Res.*, **78**, 876-885.
- Nuttli, O. W. (1985). Average seismic source-parameter relations for plate-margin earthquakes. *Tectonophysics*, **118**, 161-174.
- Nuttli, O. W. (1986). Yield estimates of Nevada Test Site explosions obtained from seismic Lg waves, *J. Geophys. Res.*, **91**, 2137-2151.
- Nuttli, O. W. (1988). Lg magnitudes and yield estimates for underground Novaya Zemlya nuclear explosions, *Bull. Seism. Soc. Am.* **78**, 873-884.
- Nuttli, O. W., and Zollweg, J. E. (1974). Relation between felt area and magnitude for central United States earthquakes. *Bull. Seism. Soc. Am.*, **64**, 1, 73-85.
- Nuttli, O. W., Bollinger, G. A., and Griffith, D. W. (1979). On the relation between modified Mercalli intensity and body-wave magnitude. *Bull. Seism. Soc. Am.*, **69**, 3, 893-909.
- Ochozinskaya, M. V. (1974). Relationship between  $m_{PV}$  and  $M_{LH}$  depending on the source depth of earthquakes. In: Magnitude and energetic classification of earthquakes (II), (in Russian), *Institut Fiziki Zemli, Akademii Nauk*, Moskva, 203-207.
- Okal, E. A. (1992a). Use of the mantle magnitude  $M_m$  for the reassessment of the moment of historical earthquakes. I: Shallow events. *Pageoph*, **139**, 1, 17-57.
- Okal, E. A. (1992b). Use of the mantle magnitude  $M_m$  for the reassessment of the moment of historical earthquakes. II: Intermediate and deep events. *Pageoph*, **139**, 1, 59-85.
- Okal, E. A. (1989). A theoretical discussion of time domain magnitudes: The Prague formula for  $M_s$  and the mantle magnitude  $M_m$ , *J. Geophys. Res.* **94**, 4194-4204.

- Okal, E. A., and Talandier, J. (1989).  $M_m$ : A variable-period mantle magnitude. *J. Geophys. Res.*, **94**, 4169-4193.
- Okal, E. A., and Talandier, J. (1990).  $M_m$ : Extension to Love waves of the concept of a variable-period mantle magnitude. *Pure Appl. Geophys.*, **134**, 355-384.
- Okal, E. A., Alsset, P.-J., Hyvernaud, O., and Schindelé, F. (2003). The deficient T waves of tsunami earthquakes. *Geophys. J. Int.*, **152**, 416-432.
- Olson, E. L., and Allen, R. (2005). The deterministic nature of earthquake rupture. *Nature*, **438**, 212-215.
- Oth, A., Bindi, D., Parolai, S., and Di Giacomo, D. (2010). Earthquake scaling characteristics and the scale-(in)dependence of seismic energy-to-moment ratio: Insights from KiK-net data in Japan. *Geophys. Res. Lett.*, **37**, L19304; doi:10.1029/2010GL044572.
- Panza, J. F., Duda, S. J., Cernobori, L., and Herak, M. (1989). Gutenberg's surface-wave magnitude calibrating function: theoretical basis from synthetic seismograms. *Tectonophysics*, **166**, 35-43.
- Panza, J. F., Prozorov, A. G., and Pazzi, G. (1993). Extension of global creepex definition ( $M_s$ - $m_b$ ) to local studies ( $M_d$ - $M_L$ ): The case of the Italian region. *Terra Nova*, **5**, 2, 150-156.
- Parolai, S., Bindi, D., Durukal, E., Grosser, H., and Milkereit, C. (2007). Source parameter and seismic moment-magnitude scaling for northwestern Turkey, *Bull. Seism. Soc. Am.* **97**(2), 655-660.
- Pasečnik, J. P., Kogan, S. D., Sultanov, D. D., and Cibulskij, V. I. (1960). Rezul'taty sejmičeskij nabljudenij pri podzemnyh jadernych vzryvach. *Akad. Nauk SSSR, Trudy Inst. Fiz. Zemli, Ser. Geofiz.*, **6**, 3-52.
- Patton, H. J. (1998). Bias in the centroid moment tensor for central Asian earthquakes: Evidence from regional surface wave data. *J. Geophys. Res.*, **103**, 26,885-26,898.
- Patton, H. J. (2001). Regional magnitude scaling, transportability, and  $M_s$ : $m_b$  discrimination at small magnitudes, *Pure and Applied Geophysics*, **158**, 1951-2015.
- Patton, H.J., and J. Schlittenhardt (2005). A transportable  $m_b(L_g)$  scale for central Europe and implications for low-magnitude  $M_s$ - $m_b$  discrimination. *Geophys. J. Int.*, **163**, 126-140.
- Pegeler, G., and Das, S. (1996). Analysis of the relationship between seismic moment and fault length for large crustal strike-slip earthquakes between 1977-92. *Geophys. Res. Lett.*, **23**, 905-908.
- Pennington, W. D., Davis, S. D., Carlson, S. M., Dupres, J., and Ewing, T. E. (1986). The evolution of seismic barriers and asperities caused by the depressuring of fault planes in oil and gas fields of South Texas. *Bull. Seism. Am.*, **76**, 4,
- Pérez-Campos, X., and Beroza, G. C. (2001). An apparent mechanism dependence of radiated seismic energy. *J. Geophys. Res.*, **106**, B6, 11,127-11,136.
- Phillips, W.S., and Stead, R.J. (2008). Attenuation of  $L_g$  in the western US using the USArray (2008). *Geophys. Res. Lett.*, **35**, L07307, doi:10.1029/2007GL032926, 5p.
- Plenkens, K., Kwiatek, G., Nakatani, M., Dresen, D., and the JAGUARS group (2010). *Seism. Res. Lett.*, **81**, 467-479.; doi: 10.1785/gssrl.81.3.467.
- Plešinger, A., Zmeškal, M., and Zednik, J. (1995). PREPROC - Software for automated preprocessing of digital data. Ver.2.1, Ed. Bergman, E., NEIC Golden/FI, Prague.
- Plešinger, A., Zmeškal, M., and Zednik, J. (1996). Automated preprocessing of digital seismograms: Principles and software. Version 2.2, Bergman, E. (Ed.), *Prague & Golden*.

- Polet, J., and Kanamori, H. (2000). Shallow subduction zone earthquakes and their tsunamigenic potential. *Geophys. J. Int.*, **142**, 684-702.
- Polet, J., and Kanamori, H. (2009). Tsunami earthquakes; in: Meyers, R. (ed), Encyclopedia of complexity and systems science, Vol. 10, 9577-9592.
- Pomeroy, P. W., and Oliver, J. (1960). Seismic waves from high altitudes nuclear explosions. *J. Geophys. Res.*, **53**, 549-562.
- Prozorov, A. G., and Hudson, J. A. (1974). A study of the magnitude difference  $M_s - m_b$  for earthquakes. *Geophys. J. R. astr. Soc.*, **39**, 551-554.
- Prozorov, A. G., Hudson, J. A., and Shimshoni, M. (1983). The behaviour of earthquake magnitudes in space and time. *Geophys. J. R. astr. Soc.*, **73**, 1-16.
- Prozorov, A. G., and Shabina, F. J. (1984). Study of properties of seismicity of the Mexico region. *Geophys. J. R. astr. Soc.*, **76**, 317-336.
- Purcaru, G., and Berckhemer, H. (1978). A magnitude scale for very large earthquakes. *Tectonophysics*, **49**, 189-198.
- Purcaru, G., and Berckhemer, H. (1982). Quantitative relations of seismic source parameters and a classification of earthquakes. *Tectonophysics*, **84**, 57-128.
- Ranalli, G., and Murphy, D. C. (1987). Rheological stratification of the lithosphere. *Tectonophysics*, **132**, 281-295.
- Randall, M. J. (1973). The spectral theory of seismic sources. *Bull. Seism. Soc. Am.*, **63**, 1133-1144.
- Rautian, T. G. (1960). Energy of earthquakes, in Y. V. Riznichenko (editor), *Methods for the Detailed Study of Seismicity*, Moscow: Izdatel'stvo Akademii Nauk SSSR, 75-114. (in Russian)
- Rautian, T. G., and Khalturin, V. I. (1978). The use of the coda for determination of earthquake source spectrum. *Bull. Seism. Soc. Am.*, **68**, 923-948.
- Rautian, T. G., Khalturin, V. I., and Shebalin, I. S. (1979). Ogibajuschtschie seismitscheskoy kody i ozenka magnitude zemletrjazenij kavkaza (The envelop of seismic coda and magnitude estimate of earthquakes in the Caucasus). *Fiziki Zemli*, **6**, 22-31.
- Rautian, T. G., Khalturin, V. I., and Zakirov, M. S. (1981). Experimental studies of seismic coda (in Russian). *Nauka*, Moscow, 144 pp.
- Rautian, T. G., Khalturin, V. I., Dotsev, N. T., and Sarkisyan, N. M. (1989). Macroseismic magnitude. In: *Voprosy inzheneroy seismologii*, iss. **30**, 98-109.
- Rautian, T., Khalturin, V., Fujita, K., Mackey, K., and Kendall, A. (2007). Origins and methodology of the Russian energy K-class system and its relationship to magnitude scales: *Seis. Res. Lett.*, **78**, 579-590.
- Reasenber, P. A., and Oppenheimer, D. (1985). FPFIT, FPLOT and FPPAGE: Fortran computer programs for calculating and displaying earthquake fault-plane solutions. *U.S. Geological Survey Open-File Report* **85-739**, 109 pp.
- Reches, Z. (1987). Determination of the tectonic stress tensor from slip along faults that obey the Coulomb yield condition. *Tectonics*, **6**, 849-861.
- Rezapour, M. (2003). Empirical global depth-distance correction terms for  $m_b$  determination based on seismic moment. *Bull. Seism. Soc. Am.*, **93**, 1, 172-189.
- Rezapour, M., and Pearce, R. G. (1998). Bias in surface-wave magnitude  $M_s$  due to inadequate distance corrections. *Bull. Seism. Soc. Am.*, **88**, 1, 43-61.

- Richard, P. (2002). Seismological methods of monitoring compliance with the Comprehensive Nuclear Test Ban Treaty. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002).. *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 369-382.
- Richard, P. G., and Wu, Z. (2011). Seismic monitoring of nuclear explosions. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, Vol. 2, 1144-1156; doi: 10.1007/978-90-481-8702-7.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bull. Seism. Soc. Am.*, **25**, 1-32.
- Richter, C. F. (1958). Elementary seismology. *W. H. Freeman and Company*, San Francisco and London, viii + 768 pp.
- Ristau, J. (2009). Comparison of magnitude estimates for New Zealand earthquakes : moment magnitude, local magnitude, and teleseismic body-wave magnitude. *Bull. Seism. Soc. Am.*, **99**, 1841-1852.
- Ristau, J., Rogers, G. C., and Cassidy, J. F. (2005). Moment magnitude-local magnitude calibration for earthquakes in Western Canada. *Bull. Seism. Soc. Am.*, **95**, 5, 1994-2000.
- Rivera, L. (1989). Inversion du tenseur des contraintes et des mécanismes au foyer à partir des données de polarité pour une population de séismes. *Thèse de doctorat, Université Louis.Pasteur de Stasbourg*.
- Riznichenko, Yu. V. (1992). Problems of seismology. *Mir and Springer Publishers*, Moscow and Berlin-Heidelberg-New York, 445 pp. (English translation of the original Russian publication of 1985).
- Romanowicz, B. (1992). Strike-slip earthquakes on quasi-vertical transcurrent faults: inferences for general scaling relations. *Geophys. Res. Lett.*, **19**, 481-484.
- Romanowicz, B. (1994). Comment on “A reappraisal of large earthquake scaling” by C. Scholz. *Bull. Seism. Soc. Am.*, **84**, 5, 1675-1676.
- Romanowicz, B., and Rundle, J. B. (1993). On scaling relations for large earthquakes. *Bull. Seism. Soc. Am.*, **83**, 1294-1297.
- Romanowicz, B., and Rundle, J. (1994). Reply to comments on “On scaling relations for large earthquakes” by B. Romanowicz and J. Rundle from the perspective of a recent nonlinear diffusion equation linking short term deformation to long term tectonics. *Bull. Seism. Soc. Am.*, **84**, 5, 1684-1685.
- Rydelek, P., and Horiuchi, S. (2006). Is earthquake rupture deterministic? *Nature*, **444**, E5-E6.
- Sadovsky, M. A., Kedrov, O. K., and Pasechnik, I. P. (1986). On the question of energetic classification of earthquakes (in Russian). *Fizika Zemli*, Moscow, **2**, 3-10.
- Satake, K. (2002). Tsunamis. In: W.H.K. Lee, H. Kanamori, P.C. Jennings and C. Kisslinger (Eds.). *International Handbook of Earthquake and Engineering Seismology*, Academic Press, Amsterdam, Vol. 1, 437–451.
- Saul, J., and Bormann, P. (2007). Rapid estimation of earthquake size using the broadband P-wave magnitude mB. *Eos, Transactions, American Geophysical Union, AGU 2007 fall meeting*, **88**, no. 52, Suppl., Abstract S53A-1035.
- Savage, M. K., and Anderson, J. G. (1995). A local-magnitude scale for the Western Great Basin-Eastern Sierra Nevada from Wood-Anderson seismograms. *Bull. Seism. Soc. Am.*, **85**, 4, 1236-1243.



- Scholz, C. H. (1982). Scaling laws for large earthquakes: consequences for physical models. *Bull. Seism. Soc. Am.*, **72**, 1-14.
- Scholz, C. H., Aviles, C. A., and Wesnousky, S. G. (1986). Scaling differences between large interplate and intraplate earthquakes. *Bull. Seism. Soc. Am.*, **76**, 1, 65-70.
- Scholz, C. H. (1990). *The mechanics of earthquakes and faulting*. Cambridge University Press, Cambridge, 439 pp.
- Scholz, C. H. (1994a). A reappraisal of large earthquake scaling. *Bull. Seism. Soc. Am.*, **84**, 215-218.
- Scholz, C. H. (1994b). Reply to comments on "A reappraisal of large earthquake scaling" by C. Scholz. *Bull. Seism. Soc. Am.*, **84**, 5, 1677-1678.
- Scholz, C. H. (1997). Size distributions for large and small earthquakes. *Bull. Seism. Soc. Am.*, **87**, 4, 1074-1077.
- Scholz, C. H. (2002). *The Mechanics of Earthquakes and Faulting*. 2<sup>nd</sup> edition, Cambridge University Press, Cambridge, 496 pp.
- Schweitzer, J., and T. Kværna (1999). Influence of source radiation patterns on globally observed short-period magnitude estimates (mb), *Bull. Seism. Soc. Am.* **89**(2), 342-347.
- Scordilis, E. M. (2006). Empirical global relations converting  $M_S$  and  $m_b$  to moment magnitude. *J. Seismol.*, **10**, 225-236.
- Seidl, D. (1980). The simulation problem for broad-band seismograms. *J. Geophys.*, **48**, 84-93.
- Seidl, D., and Berckhemer, H. (1982). Determination of source moment and radiated seismic energy from broadband recordings. *Phys. Earth Planet. Inter.*, **30**, 209-213.
- Seidl, D., and Stammer, W. (1984). Restoration of broad-band seismograms (part I). *J. Geophys.*, **54**, 114-122.
- Seidl, D., and Hellweg, M. (1988). Restoration of broad-band seismograms (part II): Signal moment determination. *J. Geophysics*, **62**, 158-162.
- Shapira, A., and Hofstetter, A. (1993). Source parameters and scaling relationships of earthquakes in Israel. *Tectonophysics*, **217**, 217-226.
- Sieberg, A. (1912). Über die makroseismische Bestimmung der Erdbebenstärke. *Gerl. Beitr. Geophys.*, **11**, 227-239.
- Simons, F. J., Dando, B. D., Allen, R. M. (2006). Automatic detection and rapid determination of earthquake magnitude by wavelet multiscale analysis of the primary arrival. *Earth Planet. Sc. Lett.*, **250**, 214-223.
- Simpson, D. W., and Leith, W. (1985). *Bull. Seism. Soc. Am.*, **75**, 1465-1468.
- Singh, S. K., and Ordaz, M. (1994). Seismic energy release in Mexican subduction zone earthquakes. *Bull. Seism. Soc. Am.*, **84**(5), 1533-1550.
- Sipkin, S. A. (1982). Estimation of earthquake source parameters by the inversion of waveform data: synthetic waveforms. *Phys. Earth Planet. Inter.*, **30**, 242-259.
- Sipkin, S. A. (1986). Estimation of earthquake source parameters by the inversion of waveform data: global seismicity. *Bull. Seism. Soc. Am.*, **76**, 1515-1541.
- Snoke, J. A., Munsey, J. W., Teague, A. G., and Bollinger, G. A. (1984). A program for focal mechanism determination by combined use of polarity and SV-P amplitude data. *Earthquake Notes*, **55**, 3, p. 15.

- Sobolev, S.V., Babeyko, A.Y., Rongjiang Wang, R., Galas, R., Rothacher, M., Sein, D.V., Schröter, J., Lauterjung, J. and Subarya, C. (2006). Towards real-time tsunami amplitude prediction, *EOS*, 87, No 37, p. 374, 378
- Sobolev, S.V., Babeyko, A.Y., Wang, R., Galas, R., Rothacher, M., Sein, D.V., Schröter, J., Lauterjung, J. and Subarya, C. (2007). Concept for fast and reliable tsunami early warning using “GPS-Shield” arrays, *J. Geophys. Res.*, 112, B08415.
- Soloviev, S. L. (1955). Classification of earthquakes in order of energy (in Russian). *Trudy Geofiz. Inst. AN SSSR*, **30**, 157, 3-31.
- Soloviev, S. L., and Shebalin, N. V. (1957). Determination of the strength of earthquakes by means of the ground displacements of surface waves (in Russian: Opređenje intensivnosti zemletrjaseńij po smeschtscheniju potschvy v poverchnostnykh volnach). *Izv. AN SSSR, Ser. Geofiz.*, no. 7, 926.
- Somerville, P. G., Irikura, K., Graves, R., Savada, S., Wald, D., Abrahamson, N., Iwasaki, Y., Kagawa, T., Smith, N., and Kowada, A. (1999). Characterizing earthquake slip models for the prediction of strong ground motion. *Seism. Res. Lett.*, **70**, 59-90.
- Sornette, D., and Sornette, A. (1994). Comments on “On scaling relations for large earthquakes” by B. Romanovicz and J. Rundle from the perspective of a recent nonlinear diffusion equation linking short term deformation to long term tectonics. *Bull. Seism. Soc. Am.*, **84**, 5, 1679-1683.
- Spence, W. (1977). Measuring the size of an earthquake. *Earthq. Inform. Bull.*, **9**, 4, 21-23.
- Sponheuer, W. (1960). Methoden zur Herdtiefenbestimmung in der Makroseismik, *Freiberger Forschungshefte C88*, Akademie Verlag, Berlin.
- Spudich, P., and Cranswick, E. (1984). Direct observation of rupture propagation during the 1979 Imperial Valley, California, earthquake using a short baseline accelerometer array. *Bull. Seism. Soc. Am.*, **74**, 2083-2114.
- Stein, S., and Okal, E. (2005). Speed and size of the Sumatra earthquake. *Nature* 434: 581-582
- Storchak, D. A., Di Giacomo, D., Bondár, I., Engdahl, E. R., Harris, J., Lee, W. H.K., Villaseñor, A., and Bormann, P. (2013). Public release of the ISC-GEM Global Instrumental earthquake catalogue (1900-2009). *Seismol. Res. Lett.*, *accepted*.
- Stover, C. W., and Coffman, J. L. (1993). *Seismicity of the United States, 1568-1989* (Revised). United States Government Printing Office, Washington.
- Street, R. L. (1976). Scaling northeastern United States/southeastern Canadian earthquakes by their Lg waves. *Bull. Seism. Soc. Am.*, **66**, 1525-1537.
- Street, R. L., and Turcotte, F. T. (1977). A study of northeastern North American spectral moments, magnitudes, and intensities. *Bull. Seism., Soc. Am.*, **67**, 2, 599-614.
- Stromeyer, D., Grünthal, G., and Wahlström, R. (2004). Chi-square regression for seismic strength parameter relations, and their uncertainties, with application to an Mw based earthquake catalogue for central and northwestern Europe. *J. Seismol.*, **8** (1), 143-153.
- Talandier, J., and Okal, E. A. (1992). One-station estimates of seismic moments from the mantle magnitude Mm: The case of the regional field ( $1.5^{\circ} \leq \Delta \leq 15^{\circ}$ ). *Pure Appl. Geophys.*, **138**, 43-60.
- Teisseyre, R., and Majewski, E. (2002). Physics of earthquakes. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 229-235.

- Teng, T. L., Wu, L., Shin, T. C., Tsai, Y. B., and Lee, W. H. K. (1997). One minute after – strong motion map, effective epicentre, and effective magnitude. *Bull. Seism. Soc. Am.*, **87**, 1209-1219.
- Thatcher, W., and Hanks, T. C. (1973). Source parameters of southern California earthquakes. *J. Geophys. Res.*, **78**, 8547-8576.
- Tinti, S., Vittori, T., and Mulargia, F. (1987). On the macroseismic magnitudes of the largest Italian earthquakes. *Tectonophysics*, **138**, 159-178.
- Tittel, B. (1977). Zur Bestimmung von Erdbebenmagnituden aus longitudinalen Kernwellen. *Gerlands Beitr. Geophys.*, **86** (1), 79-85.
- Tocher, D. (1958). Earthquake energy and ground breakage. *Bull. Seism. Soc. Am.*, **48**, 147-153.
- Toperczer, M. (1975). Zur Definition der Seismizität. *Arch. Meteorol. Geophys. Geoklimatol.*, **5**, 377-385.
- Topozada, T. R. (1975). Earthquake magnitude as a function of intensity data in California and Western Nevada. *Bull. Seism. Soc. Am.*, **65**, 5, 1223-1238.
- Tsai, V. C., Nettles, M., Ekström, G., and Dziewonski, A. (2005). Multiple CMT source analysis of the 2004 Sumatra earthquake. *Geophysical Research Letters*, **32**; L17304, doi: 10.1029/2005GL023813.
- Tsuboi, C. (1954). Determination of the Gutenberg-Richter's magnitude of earthquakes occurring in and near Japan. *Zisin. (J. Seism. Soc. Japan)*, Ser. II, **7**, 185-193.
- Tsuboi, S., Abe, K., Takano, K., and Yamanaka, Y. (1995). Rapid determination of  $M_w$  from broadband P waveforms. *Bull. Seism. Soc. Am.*, **85**, 2, 606-613.
- Tsuboi, S., Whitmore, P. M., and Sokolowski, T. J. (1999). Application of  $M_{wp}$  to deep and teleseismic earthquakes. *Bull. Seism. Soc. Am.*, **89**, 5, 1345-1351.
- Tsumura, K. (1967). Determination of earthquake magnitude from total duration of oscillation. *Bull. Earthq. Res. Inst.*, Tokyo, **45**, 7-18.
- Tumarkin, A. G., Archuleta, R. J., and Madariaga, R. (1994). Scaling relations for composite earthquake models. *Bull. Seism. Soc. Am.*, **84**, 4, 1279-1283.
- Turcotte, D. L. (1997). *Fractals and chaos in geology and geophysics*. 2<sup>nd</sup> edition, Cambridge University Press, Cambridge.
- Turcotte, R., and Malamud, E. (2002). Earthquakes as a complex system. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 209-227.
- Udías, A., and Buforn, E. (1988). Single and joint fault-plane solutions from first-motion data. In: Doornbos (1988a).
- Udías, A. (1999). *Principles of Seismology*. Cambridge University Press, United Kingdom, 475 pp.
- Udias, A. (2002). Theoretical seismology: An introduction. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 81-102.
- Uhrhammer, R. A., and Collins, E. R. (1990). Synthesis of Wood-Anderson seismograms from broadband digital records. *Bull. Seism. Soc. Am.*, **80**, 702-716.
- Uhrhammer, R. A., Loper, S. J., and Romanovicz, B. (1996). Determination of local magnitude using BDSN broadband records. *Bull. Seism. Soc. Am.*, **86**, 1314-1330.

- Uhrhammer, R. A., Hellweg, M., Hutton, K., Lombard, P., Walters, A. W., Hauksson, E., and Oppenheimer, D. (2011). California Integrated Seismic Network (CISN) local magnitude determination in California and vicinity. *Bull. Seism. Soc. Am.*, **101**, 2685-2693.
- Ulug, A. and Berckhemer, H. (1984). Frequency dependence of Q for seismic body waves in the Earth's mantle. *J. Geophys.*, **56**, 9-19.
- Utsu, T. (2002). Relationships between magnitude scales. In: *International Handbook of Earthquake and Engineering Seismology, Part A*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 733-746.
- Vaněk, J., Zapotek, A., Karnik, V., Kondorskaya, N.V., Riznichenko, Yu.V., Savarensky, E.F., Solov'yov, S.L., and Shebalin, N.V. (1962). Standarizaciya shkaly magnitude (in Russian), *Izvestiya Akad. SSSR., Ser. Geofiz.*, **2**, 153-158 (with English translation in 1962 by D. G. Frey, published in *Izv. Geophys. Ser.*).
- Veith, K. F. (1998). Comment on "Empirical determination of depth-distance corrections for mb and Mw from global seismograph network stations". *Geophys. Res. Lett.*, **25**, 22, 4241-4242.
- Veith, K. F. (2001). Magnitude estimation from short-period body waves. Paper circulated on Oct. 9, 2002, amongst the members of the IASPEI WG on Magnitude Measurements, 7 pp.
- Veith, K. F., and Clawson, G. E. (1972). Magnitude from short-period P-wave data. *Bull. Seism. Soc. Am.*, **62**, 435-452.
- Venkataraman, A., and Kanamori, H. (2004a). Effect of directivity on estimates of radiated seismic energy. *J. Geophys. Res.*, **109**, B04301; doi:10.1029/2003JB002548.
- Venkataraman, A., and Kanamori, H. (2004b). Observational constraints on the fracture energy of subduction zone earthquakes. *J. Geophys. Res.*, **109**, B04301, doi: 10.1029/2003JB002549.
- Viret, M. (1980). Determination of duration magnitude relationship for the Virginia Tech Seismic Network. In: Bollinger, G. A. (ed.): *Central Virginia Regional Seismic Report, December 1, 1980*, US Nuclear Regulatory Commission Contract No. NRC-04-077-034, 1-8.
- von Seggern, D. (1977). Amplitude-distance relation for 20-second Rayleigh waves. *Bull. Seism. Soc. Am.*, **67**, 405-411.
- Wahlström, L. (1979). Duration magnitudes for Swedish earthquakes. Seismological Institute, Uppsala, Sweden, Report No. 5-79, 22 pp.
- Wahlström, R., and Strauch, W. (1984). A regional magnitude scale for Central Europe based on crustal wave attenuation. *Seismological Dep. Univ. of Uppsala, Report No. 3-84*, 16 pp.
- Wang, J.-H., and Ou, S.-S. (1998). On scaling of earthquake faults. *Bull. Seism. Soc. Am.*, **88**, 3, 738-766.
- Wang, R., Xia, Y., Grosser, H., Wetzel, H.-U., Kaufmann, H., and Zschau, J. (2004). The 2003 Bam (SE Iran) earthquake: precise source parameters from satellite radar interferometry. *Geophys. J. Int.*, **159**, 917-922; doi: 10.1111/j.1365-246X.2004.02476.x
- Wason, H. R., Das, R., and Sharma, M. L. (2012). Magnitude conversion problem using general orthogonal regression. *Geophys. J. Int.*, **190**, 1091-1096; doi: 10.1111/j.1365-246X.2012.05520.x.
- Weinstein, S. A., and Okal, E. A. (2005). The mantle wave magnitude  $M_m$  and the slowness parameter  $\Theta$ : Five years of real-time use in the context of tsunami warning. *Bull. Seism. Soc. Am.*, **95**, 779-799; doi: 10.1785/0120040112

- Wells, D. L., and Coppersmith, K. J. (1994). New empirical relationships among magnitude, rupture length, rupture width, rupture area, and surface displacement. *Bull. Seism. Soc. Am.*, **84**, 4, 974-1002.
- Wever, Th., Trappe, H., and Meissner, R. (1987). Possible relations between crustal reflectivity, crustal age, heat flow, and viscosity of the continents. *Annales Geophysicae*, **87**, 03 B, 255-266.
- Willis, D. E. (1963). Seismic measurements of large underwater shots. *Bull. Seism. Soc. Am.*, **53**(4), 789-809.
- Willis, D. E., and Wilson, J. T. (1962). Effects of decoupling on spectra of seismic waves. . *Seism. Soc. Am.*, **52**(1), 123-131.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.
- Wood, H. O., and Neumann, F. (1931). Modified Mercalli intensity scale of 1931. *Bull. Seism. Soc. Am.*, **21**, 277-283.
- Wu, Y.-M., and Teng, T. L. (2004). Near real-time magnitude determination for large crustal earthquakes. *Tectonophysic*, **390**, 205-216.
- Wu, Y. M., and Kanamori, H. (2005). Experiment on an onsite early warning method for the Taiwan early warning system. *Bull. Seism. Soc. Am.*, **95**, 1, 347-353
- Wu., Y. M., and Zhao, L. (2006). Magnitude estimation using the first three seconds P-wave amplitude in earthquake early warning. *Geophy. Res. Lett.*, **33**, 4pp.,L16312; doi: 10.1029/2006GL026871.
- Wu, Y. M., ,Chen, C. C., Shin, Y. B., Tsai, Y. B., Lee, W. H. K., and Teng, T. L. (1997). Taiwan Rapid Earthquake Information Release System. *Seism. Res. Lett.*, **68**, 931-943.
- Wu, Y. M., Shin, T.-C., and Tsai, Y.-B. (1998). Quick and reliable determination of magnitude for seismic early warning. *Bull. Seism. Soc. Am.*, **88**, 5, 1254-1259.
- Wu, Y. M., Shin, T.-C., and Chang, C. H. (2001). Near real-time mapping of peak ground acceleration and peak ground velocity following a strong earthquake. *Bull. Seism. Soc. Am.*, **91**, 5, 1218-1228.
- Wu, Y. M., Teng, T. L., Shin, T. C., and Hsiao, N. C. (2003). Relationship between peak ground acceleration, peak ground velocitz, and intensity in Taiwan. *Bull. Seism. Soc. Am.*, **93**, 1, 386-396.
- Wu, Y. M., Yen, H., Zhao, L., Huang, B., and Liang, W. (2006). Magnitude determination using initial P waves: A single-station approach. *Goephys. Res. Lett.*, **33**, L05306; doi: 10.1029/2005GL025395.
- Wyss, M., and Brune, J. N. (1968). Seismic moment, stress and source dimensions for earthquakes in California-Nevada region. *J. Geophys. Res.*, **73**, 4681-4694.
- Xie, X. B. and T. Lay (1994). The excitation of *Lg* waves by explosions: a finite-difference investigation. *Bull Seism. Soc. Am.*, **84**, 324-342.
- Xie, X.B., and Lay, Th. (1995). The log (rms *Lg*)-*mb* scaling law slope. *Bull. Seism. Soc. Am.*, **85**(3), 834-844.
- Yacoub, N. K. (1998). Maximum spectral energy for seismic magnitude estimation. Part I: Rayleigh-wave magnitude. *Bull. Seism. Soc. Am.*, **88**, 4, 952-962.

- Yadav, R. B. S., Bormann, P., Rastogi, B. K., and Chopra, M. C. (2009). A homogeneous and complete earthquake catalog for northeast India and the adjoining region. *Seism. Res. Lett.*, **80** (4), 598-616; doi: 10.1785/gssrl.80.4.609.
- Yanovskaya, T. B. (2012). Surface-wave tomography for upper mantle studies: methods and results. Keynote lecture at the 33<sup>rd</sup> General Assembly of the European Seismological Commission, Moscow, 19-24 Aug. 2012.
- Yomogida, K., and Nakata, T. (1994). Large slip velocity of the surface rupture associated with the 1990 Luzon earthquake. *Geophys. Res. Lett.*, **21**, 1799-1802.
- Zhao, L.-F., Xie, X.-B., Wang, W.-M., and Yao, Z.-X. (2008). Regional seismic characteristics of the 9 October 2006 North Korean nuclear test. *Bull. Seism. Soc. Am.*, **98**(6), 2571-2589.
- Zoback, M. L. (1992). First- and second-order patterns of stress in the lithosphere: The World Stress Map Project. *J. Geophys. Res.*, **97**, 11703-11728.
- Zoback, M. D., and Zoback, M. L. (2002). State of stress in the Earth's lithosphere. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*. Academic Press, Amsterdam, 559-568.
- Zollo, A., Lancieri, M., and Nielsen, S. (2006). Earthquake magnitude estimation from peak amplitudes of very early seismic signals on strong motion records. *Geophys. Res. Lett.*, **33**, L23312; doi: 10.1029/2006GL027795.

| Topic   | Magnitude calibration formulas and tables, comments on their use and complementary data                                                                                            |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Peter Bormann (formerly GFZ German Research Centre for Geosciences, Potsdam, Telegrafenberg, D-14473 Potsdam, Germany;<br>E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> ) |
| Version | January 2012; DOI: 10.2312/GFZ.NMSOP-2_DS_3.1                                                                                                                                      |

## 1 Local magnitude $M_L$

The classical formula for determining the local magnitude  $M_L$  is, according to Richter (1935),

$$M_L = \log A_{\max} - \log A_0 \quad (1)$$

with  $A_{\max}$  in mm of measured zero-to-peak trace amplitude in a Wood-Anderson seismogram. The respective corrections or calibration values  $-\log A_0$  are given in Table 1 as a function of epicentral distance  $\Delta$ .

**Table 1** The classical tabulated calibration function  $-\log A_0(\Delta)$  for local magnitudes  $M_L$  according to Richter (1958).  $A_0$  are the trace amplitudes in **mm** recorded by a Wood-Anderson Standard Torsion Seismometer from an earthquake of  $M_L = 0$ .

| $\Delta$ (km) | $-\log A_0$ | $\Delta$ (km) | $-\log A_0$ | $\Delta$ (km) | $-\log A_0$ | $\Delta$ (km) | $-\log A_0$ |
|---------------|-------------|---------------|-------------|---------------|-------------|---------------|-------------|
| 0             | 1.4         | 90            | 3.0         | 260           | 3.8         | 440           | 4.6         |
| 10            | 1.5         | 100           | 3.0         | 280           | 3.9         | 460           | 4.6         |
| 20            | 1.7         | 120           | 3.1         | 300           | 4.0         | 480           | 4.7         |
| 30            | 2.1         | 140           | 3.2         | 320           | 4.1         | 500           | 4.7         |
| 40            | 2.4         | 160           | 3.3         | 340           | 4.2         | 520           | 4.8         |
| 50            | 2.6         | 180           | 3.4         | 360           | 4.3         | 540           | 4.8         |
| 60            | 2.8         | 200           | 3.5         | 380           | 4.4         | 560           | 4.9         |
| 70            | 2.8         | 220           | 3.65        | 400           | 4.5         | 580           | 4.9         |
| 80            | 2.9         | 240           | 3.7         | 420           | 4.5         | 600           | 4.9         |

Different from the above, the *IASPEI Working Group on Magnitude Measurement* now recommends, as approved by the IASPEI Commission on Seismic Observation and Interpretation (CoSOI), the following standard formula for calculating  $M_L$  (**quote from IASPEI, 2011, which uses the nomenclature ‘ $ML$ ’ instead of ‘ $M_L$ ’**):

“For crustal earthquakes in regions with attenuative properties **similar** to those of Southern California, the proposed standard equation is

$$ML = \log_{10}(A) + 1.11 \log_{10} R + 0.00189 \cdot R - 2.09, \quad (2)$$

where  $A$  = maximum **trace** amplitude in **nm** that is measured on output from a **horizontal-component** instrument that is filtered so that the response of the seismograph/filter system replicates that of a **Wood-Anderson standard seismograph** but with a static magnification of 1;

$R$  = **hypocentral distance in km**, typically less than 1000 km.

Equation (2) is an expansion of that of Hutton and Boore (1987) (see first equation in Table 2 below). The constant term in equation (2), -2.09, is based on an experimentally determined static magnification of the Wood-Anderson of 2080, rather than the theoretical magnification of 2800 that was specified by the seismograph's manufacturer. The formulation of equation (2) reflects the intent of the *Magnitude WG* that reported *ML* amplitude data not be affected by uncertainty in the static magnification of the Wood-Anderson seismograph.

For seismographic stations containing two horizontal components, amplitudes are measured independently from each horizontal component, and each amplitude is treated as a single datum. There is no effort to measure the two observations at the same time, and there is no attempt to compute a vector average.

For crustal earthquakes in regions with attenuative properties that are **different** than those of coastal California, and for measuring magnitudes with vertical-component seismographs, the standard equation is of the form:

$$ML = \log_{10}(A) + C(R) + D, \quad (3)$$

where  $A$  and  $R$  are as defined in equation (2), except that  $A$  may be measured from a **vertical-component** instrument, and where  $C(R)$  and  $D$  have been **calibrated** to adjust for the different regional attenuation and to adjust for any systematic differences between amplitudes measured on horizontal seismographs and those measured on vertical seismographs."

Table 2 gives examples of regional *Ml* calibration functions from different countries and continents followed by explanatory comments. Many more *Ml* calibration functions, but also of other magnitude formulas used by different agencies in Europe, Tunisia, Israel, Iran, Mongolia and French Polynesia have been published in the EMSC-CSEM Newsletter of Nov. 15, 1999 (see <http://www.emsc-csem.org/Documents/?d=newsl>). It can also be downloaded via the list *Download Programs & Files* (see Overview on the NMSOP-2 front page).

**Table 2** Regional calibration functions  $-\log A_0$  for *Ml* determinations.  $\Delta$  - epicentral distance and  $R$  - hypocentral ("slant") distance with  $R = \sqrt{(\Delta^2 + h^2)}$ , both in km ;  $h$  – hypocentral depth in km,  $T$  - period in s, Com. - recording component,  $S$  – station correction.

| Region                                    | $-\log A_0$                                                                                                                      | Com.                     | Range (km)                                               | Reference                                         |
|-------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------|--------------------------|----------------------------------------------------------|---------------------------------------------------|
| Southern California                       | $1.110 \log (R/100) + 0.00189(R - 100) + 3.0$                                                                                    | horiz.                   | $10 \leq R \leq 700$                                     | Hutton&Boore (1987)<br><b>oWA</b>                 |
| California                                | $1.11 \log (R) + 0.00189(R) + 0.591 + TP(n) \times T(n,z) + S$ (see Comment 1.4)                                                 | horiz.                   | $8 \leq R \leq 500$                                      | Uhrhammer et al. (2011); <b>nWA</b> but $A$ in mm |
| Mexico, Baya Calif.<br>Mexico, Imp.Valley | $1.1319 \log(R/100) + 0.0017 (R - 100) + 3.0$<br>$1.0134 \log(R/100) + 0.0025(R - 100) + 3.0$                                    | horiz.                   | $0 < R \leq 400$                                         | Vidal&Munguía (1999)<br><b>csmWA</b>              |
| US Intermountain Belt (Utah)              | $-\log A_0(\text{Richter, 1958}) + S$                                                                                            | horiz.<br>(arithm. mean) | $\Delta \leq 600$                                        | Pechmann et al. (2007); <b>oWA</b>                |
| Eastern N-America                         | $1.55 \log \Delta - 0.22$<br>$1.45 \log \Delta + 0.11$                                                                           | horiz.<br>vertic.        | $100 \leq \Delta \leq 800$<br>$100 \leq \Delta \leq 800$ | Kim (1998); <b>oWA</b>                            |
| NW Turkey                                 | $\log(R/17) + 0.00960(R-17) + 2 - S$<br>$\log(R/62) + 0.00960(R-62) + 2.95 - S$<br>$1.58 \log (R/100) + 3.0$ ; for $ML \leq 3.7$ | horiz                    | $5 \leq R \leq 62$<br>$62 \leq R \leq 110$               | Baumbach et al.(2003)<br><b>oWA</b>               |



|                                |                                              |         |                            |                                              |
|--------------------------------|----------------------------------------------|---------|----------------------------|----------------------------------------------|
| Greece                         | $2.00 \log (R/100) + 3.0$ ; for $ML > 3.7$   | horiz.  | $100 \leq \Delta \leq 800$ | Kiratzi&Papazachos (1984); <b>oWA</b>        |
| Albania                        | $1.6627 \log \Delta + 0.0008 \Delta - 0.433$ | horiz.  | $10 \leq \Delta \leq 600$  | Muco&Minga (1991) <b>oWA</b>                 |
| Central Europe                 | $0.83 \log R + (0.0017/T) (R - 100) + 1.41$  | vertic. | $100 \leq \Delta \leq 650$ | Wahlström&Strauch (1984); <b>oWA</b>         |
| SW Germany (Baden-Württemberg) | $1.11 \lg R + 0.95 R/1000 + 0.69$            | vertic. | $10 < R < 1000$            | Stange (2006) <b>oWA</b>                     |
| Norway/Fennoskan.              | $0.91 \log R + 0.00087 R + 1.010$            | vertic. | $0 < R \leq 1500$          | Alsaker et al. (1991) <b>nWA yet A in mm</b> |
| Tanzania                       | $0.776 \log(R/17) + 0.000902 (R - 17) + 2.0$ | horiz.  | $0 < R \leq 1000$          | Langston et al. (1998) <b>?oWA?</b>          |
| South Africa                   | $1.075 \log R + 0.00061R - 1.89 + S$         | vertic. | $0 < R < 1000$             | Sauder et al. (2011) <b>nWA with A in nm</b> |
| South Australia                | $1.10 \log \Delta + 0.0013 \Delta + 0.7$     | vertic. | $40 < \Delta < 700$        | Greenhalgh&Singh (1986); <b>oWA</b>          |

**Comment 1.1:** Pre-1990 regional calibration relationships in Table 2 were derived by assuming a static magnification of 2800 for the Wood-Anderson seismometer, whereas most of the more recent calibration formulas assumed the empirically derived average value of 2080 according to Uhrhammer and Collins (1990), as does the newly proposed IASPEI standard formula (2). For the response function and poles and zeros of this revised Wood-Anderson standard response see Figure 1 and Table 1 in IS 3.3. Accordingly, one might expect that reducing the constants in the older  $M_L$  formulas by  $0.129 = \log(2800/2080)$  would yield magnitude values that are compatible with the IASPEI (2011) standard formula and most of the post-1990  $M_L$  formulas. However, the re-calibration by Uhrhammer and Collins yielded also another damping factor, namely 0.7 instead of 0.8. (see Table 1 in IS 3.3). With this, the difference between the old and new  $M_L$  relationships becomes slightly frequency-dependent, as mentioned already by Kim (1998). According to D. Bindi (personal communication 2011) this difference is minimum at the response corner frequency of 1.28 Hz, 0.07 m.u. at 1 Hz, 0.08 m.u. at 0.8 Hz and 2 Hz, 0.11 m.u. at 0.5 and 3 Hz and 0.13 m.u. at  $< 0.12$  Hz and  $> 8$  Hz.

**Formulas in Table 2** that have been derived assuming the old WA response are commented in the Reference column by **oWA**, those, which accounted only for the *constant difference* in static response by **cdWA**, but those derived by using the new WA response by **nWA**. While the difference plays no role as long as the amplitudes are measured on records of the original WA seismographs it has to be taken into account, also in its frequency dependence, when producing synthetic WA seismograms. It is, therefore, also not fully correct to follow the approaches by Kanamori et al. (1993) and Pechmann et al. (2007). They used the nominal WA gain of 2800 to construct the synthetic WA seismograms and then accounted for the (assumed constant) gain difference between the synthetic and actual WA records via the station corrections. Station corrections should not only serve a formal purpose, namely to reduce the data scatter and thus improve the precision of magnitude estimates but also have a geophysical meaning which relates to the recording site itself and not to an instrumental bias.

If the procedure used by the authors of formulas presented in Table 2 is not obvious to us, we put a question mark (?) in front. Then consultation of the original publication and/or of the author is recommended.

**Comment 1.2:** Calibration of alternative regional magnitude scales to the standard formula should be made in such a way, that at some specific distance  $\leq 100$  km identical amplitude values inserted in both the standard formula and the alternative regional  $M_L$  formula yield the same magnitude value. Fig. 3.12 in Chapter 3 shows such a mutual scaling of different scales around 50 km epicentral distance. However, for larger distances, due to differences in regional attenuation, the magnitudes resulting from identical amplitude values but using different regional calibration formulas may differ by about one magnitude unit at distances larger than 800 km when comparing, e.g., the standard formula eq. (2) for Southern California (young tectonic region with high heat flow) with the formula by Saunders et al. (2011) for South Africa (stable old continental platform area, low heat flow).

To allow a more meaningful comparison of earthquakes in regions having very different attenuation of waves already within the first 100 km, Hutton and Boore (1987) recommend even a scaling at 17 km hypocentral distance. There, in agreement with the original definition of magnitude in southern California  $M_L = 3$  should correspond to 10 mm of motion on a Wood-Anderson instrument, rather than 1 mm of motion at 100 km (as for Table 1). This scaling recommendation was taken for the near range NW Turkey formula by Baumbach et al. (2003) and by Langston et al. (1998) in Tansania (see Table 2). Yet, the majority of formulas in Table 2 have been scaled to the yield  $M_L = 3$  for measured 1 mm trace motion on a WA record at  $R = 100$  km.

**Comment 1.3:** Station corrections  $S$  for  $M_L$ , based on short-period amplitude measurements, are rather large. They typically range between about  $\pm 0.5$  m.u., are usually larger for horizontal than for vertical component recordings, negative for stations on sediments or weathered rocks and near zero or positive for station sites on hard rock (e.g., Baumbach et al., 2003; Hutton and Boore, 1987; Saunders et al. (2011); Strauch and Wylegalla, 1989; Vidal and Munguía, 1999).

**Comment 1.4:** The first part of the Uhrhammer et al. (2011) calibration function for all California is identical with the expanded Hutton and Boore (1987) relationship for Southern California. However, after several inversions, these authors found out that the best least-square fit to all California data was achieved by a linear combination of the Southern California relationship with a sixth order Chebychev polynomial  $TP(n) \times T(n,z)$ , where  $n$  is summed from 1 to 6 with different coefficients  $TP(1)$  to  $TP(6)$ ,  $T(n, z) = \cos[n \times \arccos(z)]$  and  $z$  being the scale transformation of the epicentral distances  $R$ . For the given data the relationship  $z(R) = 1.11366 \log(R) - 2.00574$  transforms  $R$  in the range  $8 \text{ km} \leq R \leq 500 \text{ km}$  to  $-1 \leq z \leq +1$ . The new  $-\log A_0(R)$  relationship by Uhrhammer et al. (2011) was found to be robust with respect to constraints on the derived stations corrections and ultimately reduced to variance of the data by 50%. It should also be noted that the authors apply a pre-filter to reduce long-period noise.

## 2 Classical and new standard surface wave magnitudes $M_s$ , $M_{s\_20}$ and $M_{s\_BB}$

Gutenberg (1945a) published the following relationship for calculating the surface-wave magnitude  $M_s$  based on measurements of the “absolute” maximum horizontal-component displacement amplitude in units of  $\mu\text{m}$  at periods “around 20 s”:

$$M_s = \log A_{Hmax} + 1.656 \log \Delta + 1.818. \quad (4)$$

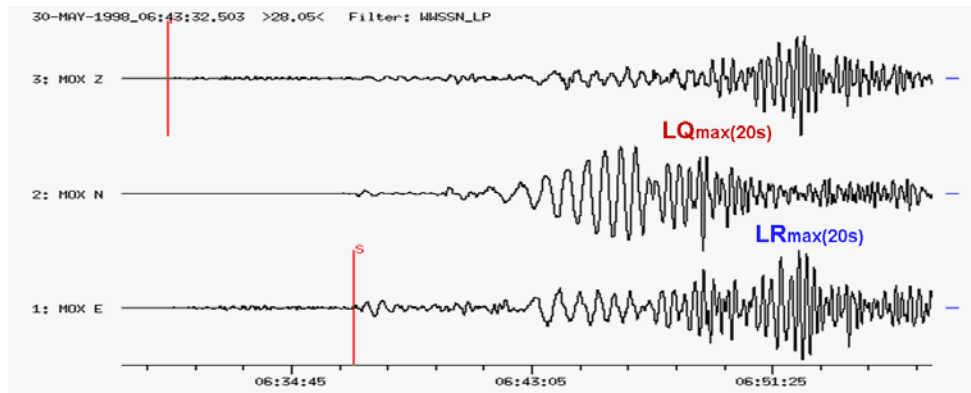
Equation (4) is applicable at epicentral distances  $\Delta$  between about  $20^\circ$  and  $130^\circ$ . Richter (1958) published tabulated Gutenberg  $M_s$  calibration values  $\sigma_s(\Delta)$ . They allow to use the Gutenberg surface-wave magnitude formula in its general form  $M_s = \log A_{Hmax}(\Delta) + \sigma_s(\Delta)$  up to  $180^\circ$  (see Table 3).

**Table 3** . Tabulated  $M_s$  calibration values  $\sigma_s(\Delta)$  according to Richter (1958) when  $A_{Hmax}$  is measured in  $\mu m$ .

| $\Delta$ (degrees) | $\sigma_s(\Delta)$ | $\Delta$ (degrees) | $\sigma_s(\Delta)$ | $\Delta$ (degrees) | $\sigma_s(\Delta)$ |
|--------------------|--------------------|--------------------|--------------------|--------------------|--------------------|
| 20                 | 4.0                | 60                 | 4.8                | 120                | 5.3                |
| 25                 | 4.1                | 70                 | 4.9                | 140                | 5.3                |
| 30                 | 4.3                | 80                 | 5.0                | 160                | 5.35               |
| 40                 | 4.5                | 90                 | 5.05               | 170                | 5.3                |
| 45                 | 4.6                | 100                | 5.1                | 180                | 5.0                |
| 50                 | 4.6                | 110                | 5.2                |                    |                    |

**Comments 2.1:** When rounded to the nearest tenth magnitude unit the values in Table 3 agree up to  $130^\circ$  with those calculated by formula (4). However, for larger distances, formula (4) yields magnitudes that are larger than those of Table 3 by 0.05 to 0.55 m.u. Reason: this simple average least-square data fitting relationship accounts only for the decay of amplitudes with distance but not for the energy focusing towards the antipodes, which is more correctly reflected by the average empirical calibration values.

**Comment 2.2:** Gutenberg calculated  $A_{Hmax}$  from the “vectorially” combined maximum amplitudes measured in the N-S and E-W component. These maxima may occur at different times and relate to different (Love and Rayleigh) surface waves. This differs from a true horizontal component vector amplitude maximum, which requires both components to be measured at the same time (within about a quarter of a period). Therefore, classical Gutenberg  $M_s$  values tends to be on average somewhat larger than  $M_s$  based on true vector sum  $A_{Hmax}$  (Figure 1 and comments by Gellert and Kanamori, 1977; Abe and Kanamori, 1980; Lienkaemper, 1984).



**Figure 1** The difference between  $M_s$  derived from these records according to the Gutenberg “vectorial” combination  $A_{Hmax} = (LR_E^2 + LQ_N^2)^{1/2}$  and the true  $A_{Hmax}$  of LR is +0.12 m.u. This is about the largest difference possible (0.15 m.u.) and depends on the BAZ (here  $85^\circ$ ).

**Comment 2.3:** Gutenberg himself did not write in his notebooks, at which period he had measured amplitude A, because he had defined his  $M_s$  scale for “periods of about 20 s”. Yet, comparison with the original bulletins also of other stations used by Gutenberg as data sources for calculating the Gutenberg-Richter  $M_{GR}$  magnitudes in “Seismicity of the Earth” (Gutenberg and Richter, 1954) showed that periods as low as 12 s and as high as 23 s were sometimes used, with values outside of the period range 18 s to 22 s not being rare (Abe, 1981; Lienkaemper, 1984).

Another surface-wave calibration function has been published by Vaněk et al. (1962) and Karnik et al. (1962). It has to be used for true (within half a period in the N-S and E-W component) vectorially combined readings of  $(A_H/T)_{\max}$  in the distance range  $1^\circ < \Delta < 160^\circ$  at periods between  $2 \text{ s} < T < 30 \text{ s}$ . This **formula, which yields in fact a broadband surface-wave magnitude estimate**, has been adopted by IASPEI in 1967 as standard formula for  $M_s$  determination. It reads:

$$M_s = \log (A_H/T)_{\max} + 1.66 \log \Delta + 3.3. \quad (5)$$

Tabulated average calibration values, to which formula (5) had been fitted, were published by Kondorskaya et al. (1981) for the distance range  $1^\circ$  to  $180^\circ$  (see Table 4).

**Table 4**  $M_s$  magnitude calibration values  $\sigma_s(\Delta)$  (according to Kondorskaya et al., 1981) for true vectorially combined horizontal component surface-wave displacement amplitudes (in  $\mu\text{m}$ ) from shallow earthquakes ( $h \leq 60 \text{ km}$ ), which relate to the largest ratio  $(A_H/T)_{\max}$  in the surface-wave group in a wide range of periods between  $2 \text{ s} < T < 30 \text{ s}$ .

| $\Delta^\circ$ | $0^\circ$ | $1^\circ$ | $2^\circ$ | $3^\circ$ | $4^\circ$ | $5^\circ$ | $6^\circ$ | $7^\circ$ | $8^\circ$ | $9^\circ$ |
|----------------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|
| $0^\circ$      |           | 3.30      | 3.80      | 4.09      | 4.30      | 4.46      | 4.59      | 4.70      | 4.80      | 4.88      |
| $10^\circ$     | 4.96      | 5.03      | 5.09      | 5.15      | 5.20      | 5.25      | 5.29      | 5.34      | 5.38      | 5.42      |
| $20^\circ$     | 5.46      | 5.50      | 5.53      | 5.56      | 5.59      | 5.62      | 5.65      | 5.68      | 5.71      | 5.73      |
| $30^\circ$     | 5.75      | 5.78      | 5.80      | 5.82      | 5.84      | 5.86      | 5.88      | 5.90      | 5.92      | 5.94      |
| $40^\circ$     | 5.96      | 5.98      | 5.99      | 6.01      | 6.03      | 6.04      | 6.06      | 6.07      | 6.09      | 6.10      |
| $50^\circ$     | 6.12      | 6.13      | 6.14      | 6.16      | 6.17      | 6.18      | 6.20      | 6.21      | 6.22      | 6.24      |
| $60^\circ$     | 6.25      | 6.26      | 6.27      | 6.28      | 6.30      | 6.31      | 6.32      | 6.33      | 6.34      | 6.35      |
| $70^\circ$     | 6.36      | 6.37      | 6.38      | 6.39      | 6.40      | 6.41      | 6.42      | 6.43      | 6.44      | 6.45      |
| $80^\circ$     | 6.46      | 6.47      | 6.48      | 6.49      | 6.49      | 6.50      | 6.51      | 6.52      | 6.53      | 6.54      |
| $90^\circ$     | 6.55      | 6.55      | 6.56      | 6.57      | 6.58      | 6.58      | 6.59      | 6.60      | 6.61      | 6.61      |
| $100^\circ$    | 6.62      | 6.63      | 6.64      | 6.64      | 6.65      | 6.66      | 6.66      | 6.67      | 6.68      | 6.69      |
| $110^\circ$    | 6.69      | 6.70      | 6.70      | 6.71      | 6.72      | 6.72      | 6.73      | 6.74      | 6.74      | 6.75      |
| $120^\circ$    | 6.75      | 6.76      | 6.76      | 6.76      | 6.77      | 6.77      | 6.78      | 6.78      | 6.78      | 6.79      |
| $130^\circ$    | 6.79      | 6.79      | 6.80      | 6.80      | 6.80      | 6.81      | 6.81      | 6.81      | 6.81      | 6.81      |
| $140^\circ$    | 6.82      | 6.82      | 6.82      | 6.82      | 6.83      | 6.83      | 6.83      | 6.83      | 6.83      | 6.83      |
| $150^\circ$    | 6.84      | 6.84      | 6.84      | 6.84      | 6.84      | 6.84      | 6.84      | 6.84      | 6.84      | 6.84      |
| $160^\circ$    | 6.84      | 6.84      | 6.83      | 6.83      | 6.83      | 6.82      | 6.82      | 6.82      | 6.82      | 6.82      |
| $170^\circ$    | 6.81      | 6.81      | 6.80      | 6.79      | 6.77      | 6.74      | 6.71      | 6.69      | 6.64      | 6.59      |
| $180^\circ$    | 6.49      |           |           |           |           |           |           |           |           |           |

**Comment 2.3:** The calibration values in Table 4 agree at epicentral distances between  $1^\circ$  and  $140^\circ$  within 0.05 magnitude units with the values calculated from the calibration term in the Prague-Moscow surface-wave magnitude formula (5). For larger distances, however, this formula overestimates the magnitude between 0.05 and 0.55 (at  $180^\circ$ ) m.u., for the same

reason as the Gutenberg (1945a) formula (4). Therefore, preference should be given to the use of the tabulated values, at least for distances  $>140^\circ$ .

**Comment 2.4:** The calibration values in Table 4 have - since the 1960s - been used at the basic seismic stations of the former Soviet Union (USSR) and its follow-up Commonwealth of Independent States (CIS) countries for  $M_s$  determination from **both horizontal and vertical component readings**. The latter usually relate to the Airy phase of Rayleigh waves ( $R_{\max}$ , see section 2.3 in Chapter 2). Table 5 gives the time difference between  $R_{\max}$  and the P wave as a function of distance. The good agreement between vertical and horizontal component  $M_s$  determinations has also been confirmed by Hunter (1972). Therefore, from May 1975 onwards, also the USGS decided to calculate their  $M_s$  exclusively from vertical component reading using formula (5)..

**Table 5** Approximate time interval ( $t_{R_{\max}} - t_P$ ) between the arrival of the maximum Rayleigh wave amplitude and the first onset of P waves as a function of  $\Delta$  according to Archangelskaya (1959) and Gorbunova and Kondorskaya (1977) (copied from Willmore, 1979).

| $\Delta^\circ$ | $t_{R_{\max}} - t_P$<br>(min) | $\Delta^\circ$ | $t_{R_{\max}} - t_P$<br>(min) | $\Delta^\circ$ | $t_{R_{\max}} - t_P$<br>(min) |
|----------------|-------------------------------|----------------|-------------------------------|----------------|-------------------------------|
| 10             | 4-5                           | 55             | 26                            | 100            | 45-46                         |
| 15             | 6-8                           | 60             | 28-29                         | 105            | 47-48                         |
| 20             | 9-10                          | 65             | 31                            | 110            | 48-50                         |
| 25             | 10-12                         | 70             | 33                            | 115            | 53                            |
| 30             | 13-14                         | 75             | 35                            | 120            | 55                            |
| 35             | 15-16                         | 80             | 37                            | 125            | 57                            |
| 40             | 18-19                         | 85             | 39-40                         | 130            | 60                            |
| 45             | 21                            | 90             | 42                            | 140            | 64                            |
| 50             | 24                            | 95             | 43                            | 150            | 70                            |

Table 6 summarizes the distance-dependent prevailing period ranges at which  $M_s$  according to formula (5) should be calculated. Note that already in Willmore (1979) it is recommended that: **“When the period differs significantly from the values in Table 3.2.2.1 (here Table 6), it may be advisable not to use the data for magnitude determination.”** This, however, is permanently done when measuring the narrowband spectral surface-wave  $M_{s20}$  (see formula (7)).

**Table 6** Distance-dependent prevailing period ranges at which broadband  $M_s$  should be measured (according to Vaněk et al. (1962), Karnik et al. (1962) and Willmore (1979). With currently available very broadband velocity seismographs  $V_{\max}$  may occasionally be observed at even longer period periods up to about 60 s.

| $\Delta^\circ$ | T in s | $\Delta^\circ$ | T in s | $\Delta^\circ$ | T in s | $\Delta^\circ$ | T in s |
|----------------|--------|----------------|--------|----------------|--------|----------------|--------|
| 1              | 3-5    | 10             | 7-10   | 40             | 12-18  | 90             | 16-22  |
| 2              | 4-6    | 15             | 8-15   | 50             | 12-20  | 100            | 16-25  |
| 4              | 5-7    | 20             | 9-14   | 60             | 14-20  | 120            | 16-25  |
| 6              | 5-8    | 25             | 9-16   | 70             | 14-22  | 140            | 18-25  |
| 8              | 6-9    | 30             | 10-16  | 80             | 16-22  | 160            | 18-25  |

**Comment 2.5:** The periods at which  $R_{\max}$  occurs depend on epicentral distance, crust and upper mantle structure along the travel path and the earthquake magnitude. Accordingly, the period of  $R_{\max}$  may vary in a wide range of periods between some 2 and <60 s (see Figures 7 and 8 in Bormann et al., 2009 and related figures in Chapter 3 of this Manual). This notwithstanding, measuring the ratio  $(A/T)_{\max}$  yields rather stable  $M_s$  determinations (Soloviev, 1955). This has been confirmed recently by Bormann et al. (2009) down to local-regional distances and periods as short as 2 to 5 s for dominantly continental travel paths. For chiefly oceanic paths this still has to be proved (or disproved), as well as the distance-dependent period-ranges at which  $R_{\max}$  is observed. This, however, might be more difficult because of the lack of regional seismic networks in ocean areas. For larger magnitudes and epicentral distances, however, the path-dependent variability of periods is much reduced (see Figures 7 and 8 in Bormann et al. (2009).

**Comment 2.6:** Because of the above, the Prague-Moscow formula (6) is now recommended, with slight modification, by IASPEI (2011) for calculating the new **standard broadband surface-wave magnitude  $M_s_{BB}$**  (see IS 3.3):

$$M_s_{BB} = \log_{10}(V_{\max}/2\pi) + 1.66 \log_{10}\Delta + 0.3, \quad (6)$$

where:  $V_{\max}$  = ground **velocity in nm/s** is associated with the maximum trace-amplitude in the surface-wave train, as recorded on a **vertical-component** seismogram that is **proportional to velocity**;

$T$ , the period of the surface-wave, should satisfy the condition **3 s <  $T$  < 60 s**;

$\Delta$  = **epicentral** distance in degrees, with (6) being applicable in the range  **$2^\circ \leq \Delta \leq 160^\circ$** ;

the focal depth  $h$  being **less than 60 km**;

and the term  $\log_{10}(V_{\max}/2\pi)$  replaces the  $\log_{10}(A/T)_{\max}$  term of the original formula (5).

The same calibration terms as in formula (6) is used for the other IASPEI (2011) **standard surface-wave magnitude  $M_s_{20}$** , which is identical with the NEIC/PDE  $M_s$ :

$$M_s_{20} = \log_{10}(A/T) + 1.66 \log_{10}\Delta + 0.3, \quad (7)$$

where:  $A$  = **vertical-component** ground displacement in **nm** measured from the maximum surface-wave trace-amplitude having a period between **18 s and 22 s** on a simulated World-Wide Standardized Seismograph Network (WWSSN) **long-period seismograph** record, with  $A$  being determined by dividing the maximum trace amplitude by the magnification of the simulated WWSSN-LP response at period  $T$ , and  $\Delta$  = **epicentral** distance in degrees **in the range  $20^\circ \leq \Delta \leq 160^\circ$** .

**Comment 2.7:** Theoretically, for periods  $T = 20$  s, the IASPEI standard formula (5) for  $M_s$  is expected to yield magnitude values that are 0.18 m.u. larger than those derived from the Gutenberg formula (4). This, however, is not confirmed by empirical testing (see Lienkaemper, 1984) and partially due to the fact that Gutenberg did not measure the true maximum vector amplitude (see *Comment 2.2*).

For more detailed discussion and figures on the relationship between  $M_s_{20}$  and  $M_s_{BB}$  see Chapter 3 and Bormann et al. (2009).

### 3 Classical and new standard body-wave magnitudes $mB$ and $mb$

Gutenberg (1945b and c) developed a magnitude formula for teleseismic body waves P, PP and S. He used it in the wide period range between about 2 s and 20 s (Abe, 1981 and 1984; Abe and Kanamori, 1980). The formula reads:

$$mB = \log (A/T)_{\max} + Q(\Delta, h). \quad (8)$$

Thus,  $mB$  is in fact a **medium-period body-wave broadband magnitude**. Gutenberg and Richter (1956) published for the three types of body waves diagrams of  $Q$  values as a function of  $\Delta$  and source depth  $h$  (Figures 1a-c).

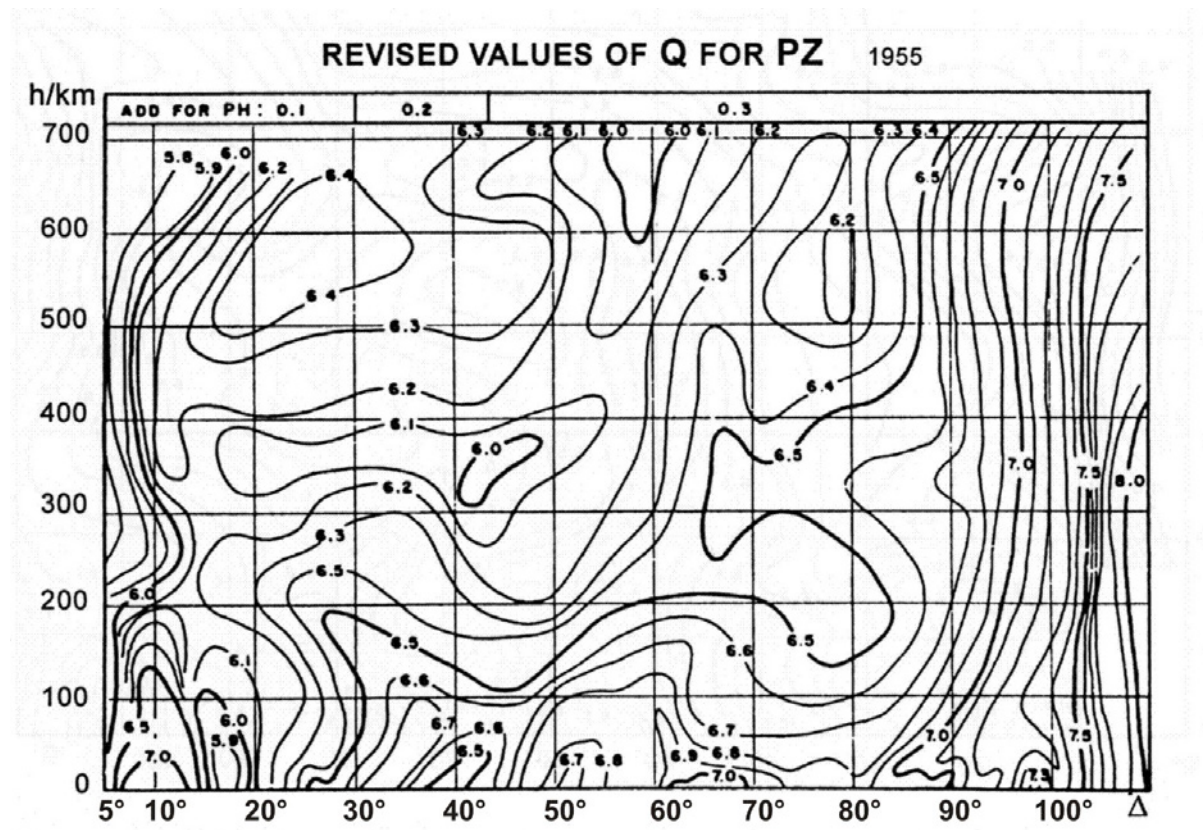


Figure 1a

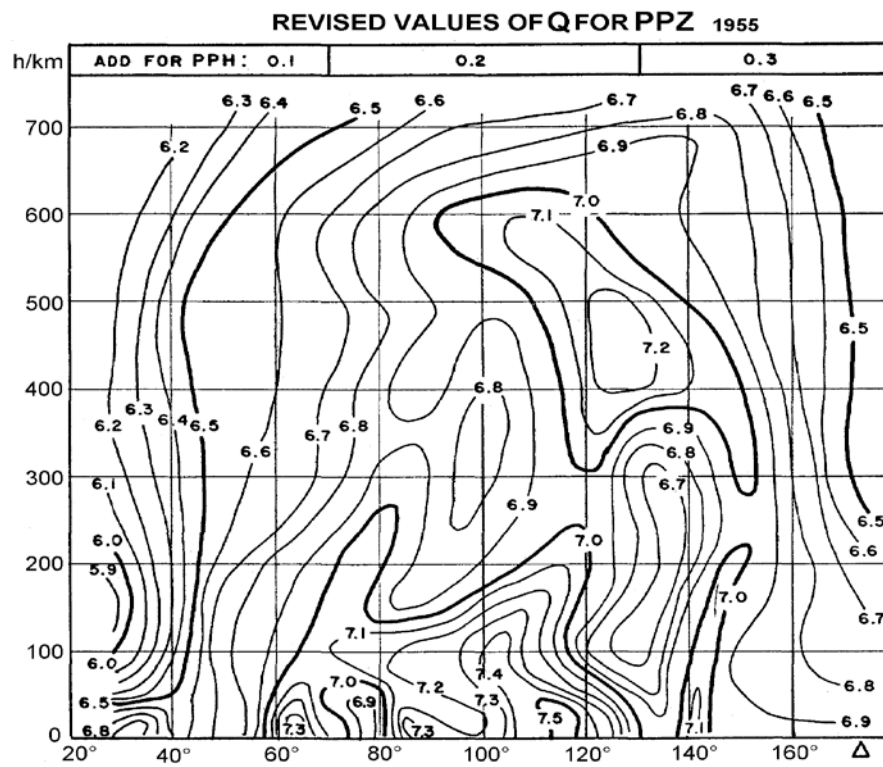


Figure 1b

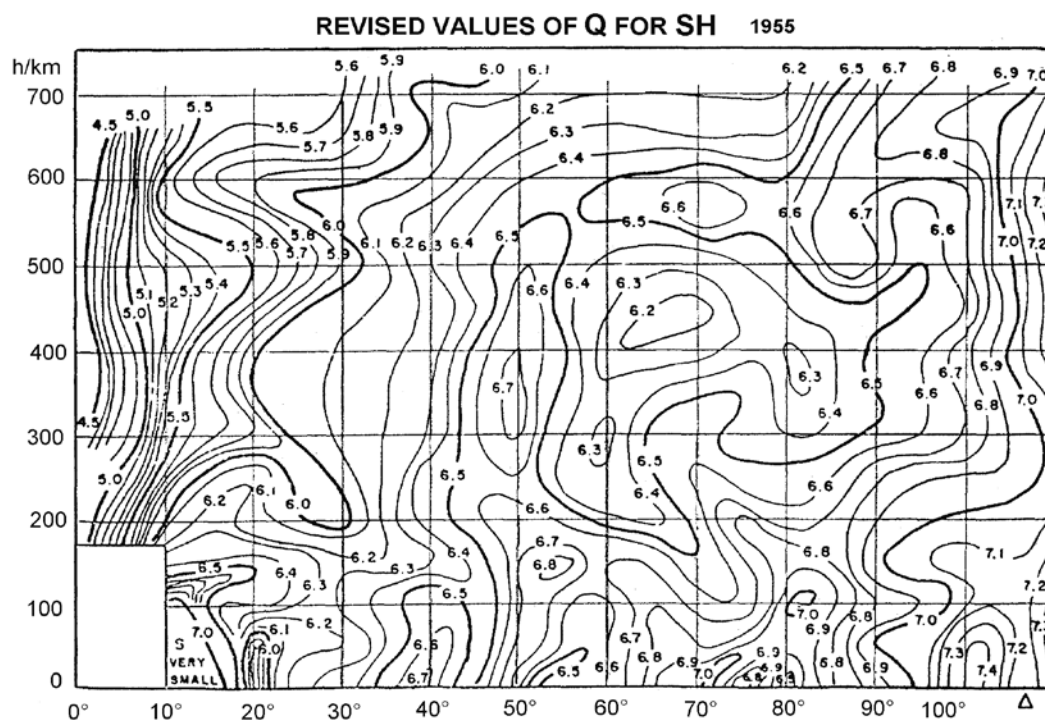


Figure 1c



Gutenberg and Richter (1956) published also a **table of  $Q(\Delta)$**  values for P, PP and S waves in vertical ( $V = Z$ ) and horizontal (H; vectorially-combined) components **for shallow earthquakes** (Table 7). All G-R  $Q$  values are valid when  $A$  is given in  $\mu\text{m}$  ( $10^{-6}$  m). If  $A$  is measured in nm ( $10^{-9}$  m), as now common in modern analysis programs for processing data with high dynamic range and resolution, the  $Q$  values have to be reduced by -3.

**Table 7** Values of  $Q(\Delta)$  for P, PP and S waves for shallow shocks ( $h \approx 15$  km) according to Gutenberg and Richter (1956) if the ground amplitude is given in  $\mu\text{m}$ .

| $\Delta^\circ$ | PV  | PH  | PPV | PPH | SH  | $\Delta^\circ$ | PV  | PH  | PPV | PPH | SH  | $\Delta^\circ$ | PV  | PH  | PPV | PPH | SH  |
|----------------|-----|-----|-----|-----|-----|----------------|-----|-----|-----|-----|-----|----------------|-----|-----|-----|-----|-----|
| 16             | 5.9 | 6.0 |     |     | 7.2 | 56             | 6.8 | 7.1 | 6.9 | 7.0 | 6.6 | 96             | 7.3 | 7.6 | 7.2 | 7.4 | 7.1 |
| 17             | 5.9 | 6.0 |     |     | 6.8 | 57             | 6.8 | 7.1 | 6.9 | 7.0 | 6.6 | 97             | 7.4 | 7.8 | 7.2 | 7.4 | 7.2 |
| 18             | 5.9 | 6.0 |     |     | 6.2 | 58             | 6.8 | 7.1 | 7.0 | 7.1 | 6.6 | 98             | 7.5 | 7.8 | 7.2 | 7.4 | 7.3 |
| 19             | 6.0 | 6.1 |     |     | 5.8 | 59             | 6.8 | 7.1 | 7.0 | 7.2 | 6.6 | 99             | 7.5 | 7.8 | 7.2 | 7.4 | 7.3 |
| 20             | 6.0 | 6.1 |     |     | 5.8 | 60             | 6.8 | 7.1 | 7.1 | 7.3 | 6.6 | 100            | 7.4 | 7.7 | 7.2 | 7.4 | 7.4 |
| 21             | 6.1 | 6.2 |     |     | 6.0 | 61             | 6.9 | 7.2 | 7.2 | 7.4 | 6.7 | 101            | 7.3 | 7.6 | 7.2 | 7.4 | 7.4 |
| 22             | 6.2 | 6.3 |     |     | 6.2 | 62             | 7.0 | 7.3 | 7.3 | 7.4 | 6.7 | 102            | 7.4 | 7.7 | 7.2 | 7.4 | 7.4 |
| 23             | 6.3 | 6.4 |     |     | 6.2 | 63             | 6.9 | 7.3 | 7.3 | 7.4 | 6.7 | 103            | 7.5 | 7.9 | 7.2 | 7.4 | 7.3 |
| 24             | 6.3 | 6.5 |     |     | 6.2 | 64             | 7.0 | 7.3 | 7.3 | 7.5 | 6.8 | 104            | 7.6 | 7.9 | 7.3 | 7.5 | 7.3 |
| 25             | 6.5 | 6.6 |     |     | 6.2 | 65             | 7.0 | 7.4 | 7.3 | 7.5 | 6.9 | 105            | 7.7 | 8.1 | 7.3 | 7.5 | 7.2 |
| 26             | 6.4 | 6.6 |     |     | 6.2 | 66             | 7.0 | 7.4 | 7.3 | 7.4 | 6.9 | 106            | 7.8 | 8.2 | 7.4 | 7.6 | 7.2 |
| 27             | 6.5 | 6.7 |     |     | 6.3 | 67             | 7.0 | 7.4 | 7.2 | 7.4 | 6.9 | 107            | 7.9 | 8.3 | 7.4 | 7.6 | 7.2 |
| 28             | 6.6 | 6.7 |     |     | 6.3 | 68             | 7.0 | 7.4 | 7.1 | 7.3 | 6.9 | 108            | 7.9 | 8.3 | 7.4 | 7.6 | 7.2 |
| 29             | 6.6 | 6.7 |     |     | 6.3 | 69             | 7.0 | 7.4 | 7.0 | 7.2 | 6.9 | 109            | 8.0 | 8.4 | 7.4 | 7.6 | 7.2 |
| 30             | 6.6 | 6.8 | 6.7 | 6.8 | 6.3 | 70             | 6.9 | 7.3 | 7.0 | 7.2 | 6.9 | 110            | 8.1 | 8.5 | 7.4 | 7.6 | 7.2 |
| 31             | 6.7 | 6.9 | 6.7 | 6.8 | 6.3 | 71             | 6.9 | 7.3 | 7.1 | 7.3 | 7.0 | 112            | 8.2 | 8.6 | 7.4 | 7.6 |     |
| 32             | 6.7 | 6.9 | 6.8 | 6.9 | 6.4 | 72             | 6.9 | 7.3 | 7.1 | 7.3 | 7.0 | 114            | 8.6 | 9.0 | 7.5 | 7.7 |     |
| 33             | 6.7 | 6.9 | 6.8 | 6.9 | 6.4 | 73             | 6.9 | 7.2 | 7.1 | 7.3 | 6.9 | 116            | 8.8 |     | 7.5 | 7.7 |     |
| 34             | 6.7 | 6.9 | 6.8 | 6.9 | 6.5 | 74             | 6.8 | 7.1 | 7.0 | 7.2 | 6.8 | 118            | 9.0 |     | 7.5 | 7.7 |     |
| 35             | 6.7 | 6.9 | 6.8 | 6.9 | 6.6 | 75             | 6.8 | 7.1 | 6.9 | 7.1 | 6.8 | 120            |     |     | 7.5 | 7.7 |     |
| 36             | 6.6 | 6.9 | 6.7 | 6.8 | 6.6 | 76             | 6.9 | 7.2 | 6.9 | 7.1 | 6.8 | 122            |     |     | 7.4 | 7.6 |     |
| 37             | 6.5 | 6.7 | 6.7 | 6.8 | 6.6 | 77             | 6.9 | 7.2 | 6.9 | 7.1 | 6.8 | 124            |     |     | 7.3 | 7.5 |     |
| 38             | 6.5 | 6.7 | 6.7 | 6.8 | 6.6 | 78             | 6.9 | 7.3 | 6.9 | 7.1 | 6.9 | 126            |     |     | 7.2 | 7.4 |     |
| 39             | 6.4 | 6.6 | 6.6 | 6.7 | 6.7 | 79             | 6.8 | 7.2 | 6.9 | 7.1 | 6.8 | 128            |     |     | 7.1 | 7.4 |     |
| 40             | 6.4 | 6.6 | 6.6 | 6.7 | 6.7 | 80             | 6.7 | 7.1 | 6.9 | 7.1 | 6.7 | 130            |     |     | 7.0 | 7.3 |     |
| 41             | 6.5 | 6.7 | 6.5 | 6.6 | 6.6 | 81             | 6.8 | 7.2 | 7.0 | 7.2 | 6.8 | 132            |     |     | 7.0 | 7.3 |     |
| 42             | 6.5 | 6.7 | 6.5 | 6.6 | 6.5 | 82             | 6.9 | 7.2 | 7.1 | 7.3 | 6.9 | 134            |     |     | 6.9 | 7.2 |     |
| 43             | 6.5 | 6.7 | 6.6 | 6.7 | 6.5 | 83             | 7.0 | 7.4 | 7.2 | 7.4 | 6.9 | 136            |     |     | 6.9 | 7.2 |     |
| 44             | 6.5 | 6.7 | 6.7 | 6.8 | 6.5 | 84             | 7.0 | 7.4 | 7.3 | 7.5 | 6.9 | 138            |     |     | 7.0 | 7.3 |     |
| 45             | 6.7 | 6.9 | 6.7 | 6.8 | 6.5 | 85             | 7.0 | 7.4 | 7.3 | 7.5 | 6.8 | 140            |     |     | 7.1 | 7.4 |     |
| 46             | 6.8 | 7.1 | 6.7 | 6.8 | 6.6 | 86             | 6.9 | 7.3 | 7.3 | 7.5 | 6.7 | 142            |     |     | 7.1 | 7.4 |     |
| 47             | 6.9 | 7.2 | 6.7 | 6.8 | 6.6 | 87             | 7.0 | 7.3 | 7.2 | 7.4 | 6.8 | 144            |     |     | 7.0 | 7.3 |     |
| 48             | 6.9 | 7.2 | 6.7 | 6.8 | 6.7 | 88             | 7.1 | 7.5 | 7.2 | 7.4 | 6.8 | 146            |     |     | 6.9 | 7.2 |     |
| 49             | 6.8 | 7.1 | 6.7 | 6.8 | 6.7 | 89             | 7.0 | 7.4 | 7.2 | 7.4 | 6.8 | 148            |     |     | 6.9 | 7.2 |     |
| 50             | 6.7 | 7.0 | 6.7 | 6.8 | 6.6 | 90             | 7.0 | 7.3 | 7.2 | 7.4 | 6.8 | 150            |     |     | 6.9 | 7.2 |     |
| 51             | 6.7 | 7.0 | 6.7 | 6.8 | 6.5 | 91             | 7.1 | 7.5 | 7.2 | 7.4 | 6.9 | 152            |     |     | 6.9 | 7.2 |     |
| 52             | 6.7 | 7.0 | 6.7 | 6.8 | 6.5 | 92             | 7.1 | 7.4 | 7.2 | 7.4 | 6.9 | 154            |     |     | 6.9 | 7.2 |     |
| 53             | 6.7 | 7.0 | 6.7 | 6.8 | 6.6 | 93             | 7.2 | 7.5 | 7.2 | 7.4 | 6.9 | 156            |     |     | 6.9 | 7.2 |     |
| 54             | 6.8 | 7.1 | 6.8 | 6.9 | 6.6 | 94             | 7.1 | 7.4 | 7.2 | 7.4 | 7.0 | 158            |     |     | 6.9 | 7.2 |     |
| 55             | 6.8 | 7.1 | 6.9 | 7.0 | 6.6 | 95             | 7.2 | 7.6 | 7.2 | 7.4 | 7.0 | 160            |     |     | 6.9 | 7.2 |     |
|                |     |     |     |     |     |                |     |     |     |     |     | 170            |     |     | 6.9 | 7.2 |     |

Besides these original tabulated Gutenberg-Richter Q values for shallow crustal earthquake there exist also tables with values resulting from scanning Figure 1a for vertical component P-wave readings in discrete intervals of source distance and depth. The respective table, produced by the USGS/NEIC, is reproduced in PD 3.1 of this Manual, together with the source code of the program used for interpolation between these tabulated values.

**Comment 3.1:** Until now, this table has been used by the USGS/NEIC exclusively for scaling **short-period narrowband P-wave amplitude readings** at periods around 1 s ( $T < 3s$ ) and **calculating mb by essentially the same formula (8)**. However, instead of looking for the true  $(A/T)_{\max}$  it has become common to measure the maximum trace amplitude in the used short-period narrow-band record, calculate from it the related ground motion amplitude by dividing the trace amplitude by the record magnification at the related period, and then dividing this calculated “ground motion amplitude” by the related period.

**Comment 3.2:** Note that for periods below 4 s differences in frequency-dependent attenuation, which are not accounted for by formula (8), may become significant. For more detailed discussion and figures on this issue see Chapter 3 of this Manual as well as Bormann et al. (2009).

**Comment 3.3:** At the IASPEI General Assembly in Zürich (1967) the Committee on Magnitudes recommended that stations should report the magnitude for all waves for which calibration functions are available, to publish amplitude and period values separately and to correct body-wave magnitudes that were exclusively determined from short-period records. However, aiming at simplifying routine analysis and saving time, it has become wide-spread practice to measure only vertical-component P-wave amplitudes, and in western countries only on short-period narrowband recordings.

**The Gutenberg and Richter (1956)  $Q(\Delta, h)_{PZ}$  values are still the accepted standard calibration values for both mB and mb.**

The now recommended IASPEI (2011) standard formula for broadband mB\_BB reads:

$$mB\_BB = \log_{10}(V_{\max}/2\pi) + Q(\Delta, h) - 3.0 \quad (9)$$

where (quote): “ $V_{\max}$  = ground **velocity in nm/s** associated with the maximum trace-amplitude in **the entire P-phase train** (time spanned by P, pP, sP, and possibly PcP and their codas, but ending **preferably** before PP, as recorded on a vertical-component seismogram that is **proportional to velocity**, where the period of the measured phase, T, should satisfy the condition  $0.2 \text{ s} < T < 30 \text{ s}$ , and where T should be preserved together with  $V_{\max}$  in. bulletin data-bases”. Standard mB\_BB is calculated in the **distance range  $20^\circ \leq \Delta \leq 100^\circ$  and for source depths down to 700 km**. However, equation (9) differs from the Gutenberg and Richter (1956) equation (8) for mB by virtue of the  $\log_{10}(V_{\max}/2\pi)$  term, which replaces the classical  $\log_{10}(A/T)_{\max}$  term.

The now recommended IASPEI (2011) standard formulas for mb determination reads:

$$mb = \log_{10}(A/T) + Q(\Delta, h) - 3.0, \quad (10)$$

where (quote): “A = P-wave ground amplitude in **nm** calculated from the maximum trace-amplitude in the **entire P-phase train** (time spanned by P, pP, sP, and possibly PcP and their codas, and ending preferably before PP)”;

T = period in seconds of the maximum P-wave trace amplitude with  $T < 3$  s. The distance and depth range of measurement as well as the  $Q(\Delta, h)$  values used are the same as for mb\_BB.

Both T and the maximum trace amplitude are measured **on simulated vertical component WWSSN-SP records**. A is determined by dividing the maximum trace amplitude by the magnification of the simulated WWSSN-SP response at period T. For the WWSSN-SP response parameters, the tabulated Q-values, and the algorithm used by the USGS/NEIC for mb determination see IS 3.3, respectively PD 3.1 of this Manual.

There is still **another IASPEI (2011) recommended short-period standard magnitude** which is used in the local and regional distance range and has been scaled to be compatible with teleseismic mb. It is also measured on simulated vertical component WWSSN-SP records **in a narrow period range around 1 s** and termed **mb\_Lg**. Its formula reads:

$$\text{mb\_Lg} = \log_{10}(A) + 0.833\log_{10}[r] + 0.4343\gamma(r - 10) - 0.87, \quad (11)$$

where (**quote**): “A = “sustained ground-motion amplitude” in **nm**, defined as the third largest amplitude in the time window corresponding to group velocities of 3.6 to 3.2 km/s, in the period (T) range 0.7 s to 1.3 s, r = **epicentral distance** in **km** and  $\gamma$  = coefficient of attenuation in  $\text{km}^{-1}$ .”

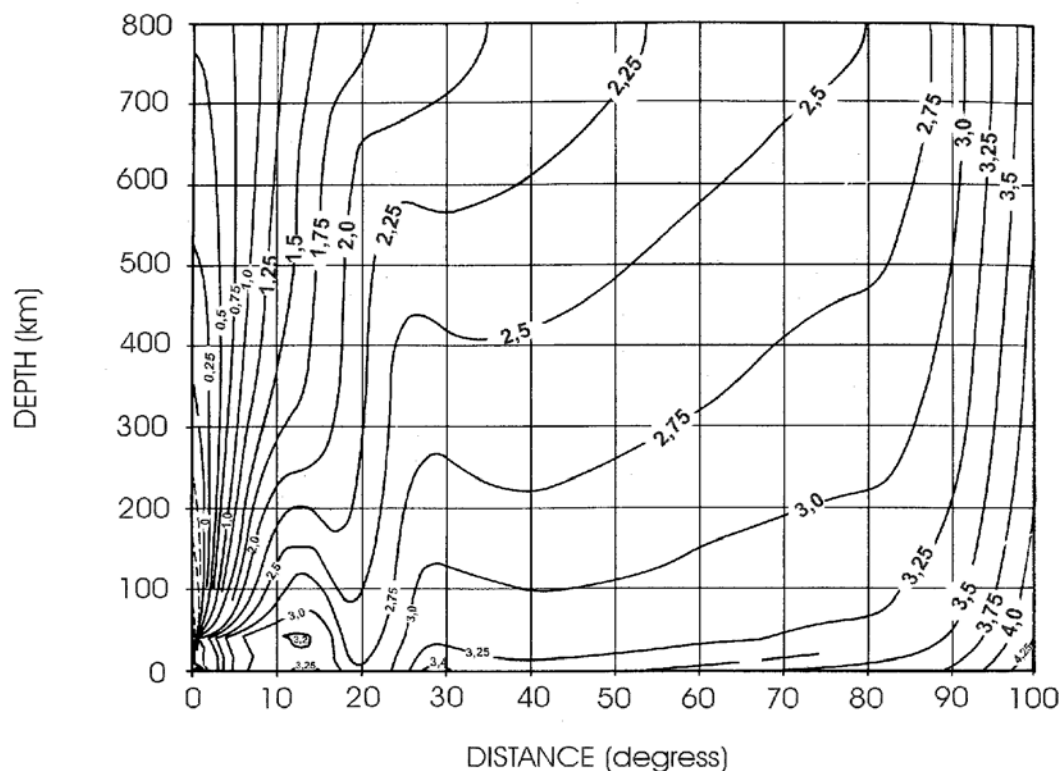
$\gamma$  is related to the quality factor Q through the equation  $\gamma = \pi/(Q \cdot U \cdot T)$ , where U is group velocity and T is the wave period of the  $L_g$  wave.  $\gamma$  is a strong function of crustal structure and should be determined specifically for the region in which the mb\_Lg is to be used.

Arrival times with respect to the origin of the seismic disturbance are used, along with epicentral distance, to compute group velocity U.”

## 4 Body-wave and surface-wave magnitude scales used by the IDC

Another calibration function  $P(\Delta, h)$  for short-period mb determination at periods around 1 s has been elaborated by Veith and Clawson (1972) (see Figure 2). It is based on large sets of vertical-component P-wave recordings of WWSSN-SP seismographs from large explosions at 19 different sites. The  $P(\Delta, h)$  were specifically derived from short-period data and scaled to the level of the Gutenberg-Richter  $Q(\Delta)_{PZ}$  for surface focus.

The  $P(\Delta, h)$  calibration curves are currently used by the International Data Center (IDC) under the Comprehensive Test-Ban Treaty Organisation (CTBTO). However, the short-period filter applied at the IDC to the original broadband records prior to mb measurement does not simulate the standard WWSSN-SP response, which was used to derive the Veith-Clawson calibration curves. Rather, at the IDC one uses a more high-frequency response that peaks at 3.4 Hz (see Figure 16 in IS 3.3). This contributes partially to the earlier saturation of mb(IDC) as compared to mb(PDE) and mb(IASPEI). The other reason is the very short and fixed IDC measurement time-window for P amplitudes within the first 5.5 s after the P onset (see Chapter 17, p. 26-27) which may cause for great earthquakes with magnitudes above 8 differences between mb(IASPEI) and mb(IDC) of more than 1 m.u (see related discussion in section 4.3 of IS 3.3).



**Figure 2** Calibration functions  $P(\Delta, h)$  for mb determination from narrow-band short-period vertical-component records with peak displacement magnification around 1 Hz (WWSSN-SP characteristic) according to Veith and Clawson (1972). **Note:** The  $P(\Delta, h)$  values have to be used in conjunction with maximum P-wave peak-to-trough ( $2A!$ ) amplitudes in units of nanometers ( $1 \text{ nm} = 10^{-9} \text{ m}$ ) (modified from Veith and Clawson (1972). Magnitude from short-period P-wave data, BSSA, **62**(2), p. 446, © Seismological Society of America).

When comparing Figure 1a and Figure 2 it is obvious that the  $P(\Delta, h)$  curves look much smoother than the  $Q(\Delta, h)_{PZ}$  curves. This agrees with the also much smoother preliminary revision of the G-R  $Q(\Delta)$  curve for mB calibration of shallow focus earthquakes by Saul and Bormann (2007) (see Figure 22 in IS 3.3). However, for deep events the  $P(\Delta, h)$  curves have been scaled according to upper mantle attenuation relationships available some 40 years ago, which have not yet been well constrained and still are largely model-dependent. Granville et al. (2005) confirmed that because of the large differences between the Gutenberg-Richter and the Veith-Clawson calibration functions for very deep earthquakes individual event magnitudes  $mb(PDE)$  and  $mb(IDC)$  may differ by several tenths of m.u. Since, however, the main focus of the CTBTO is on the location, magnitude determination and discrimination of shallow earthquakes and underground nuclear explosions (UNEs) these biases in  $P(\Delta, h)$  for deep earthquakes are not so relevant for the IDC assignment. But earthquake seismologists using IDC data should be aware of it and account for systematic  $mb(IDC)$  biases.

The local magnitude scale used by the IDC also differs from the classical  $M_L$  procedure and has been scaled so as to yield magnitude values that are compatible with the teleseismic  $mb(IDC)$ . For details see Chapter 17, Section 7.4.2.1.

More appropriate for global seismology is the use at the IDC of an  $M_s$  calibration relationship that has been specifically derived for the use of 20 s surface-wave amplitude readings. The IDC uses the difference between its  $m_b$  and  $M_s$  as the main criterion for discriminating between natural earthquakes and strong sub-surface explosions. However, Evernden (1971), von Seggern (1977), Herak and Herak (1993) and Rezapour and Pearce (1998) have shown that using the IASPEI standard formula (5) for  $M_s$  based on 20 s surface-wave readings only results in systematic distance-dependent errors up to about 0.6 m.u. On the one hand this is due to the fact that the simple relationship (5) cannot account for the energy focusing of surface-waves towards the antipodes and on the other hand to the fact that for distances  $<60^\circ$  the average periods of Rayleigh surface-wave maxima tend to be smaller than 18 s, down to about 8 s only around  $10^\circ$  (see Table 6 in this Data Sheet, Figure 11 in IS 3.3 and Bormann et al., 2009). Yet the more elaborate Rezapour and Pearce (1998)  $M_s$  relationship for 20 s surface waves compensates for it. It reads, when amplitudes are measured in nm:

$$M_s = \log_{10}(A/T)_{\max} + 1/3 \log_{10}(\Delta) + 1/2 \log_{10}(\sin \Delta) + 0.0046\Delta + 2.370, \quad (12)$$

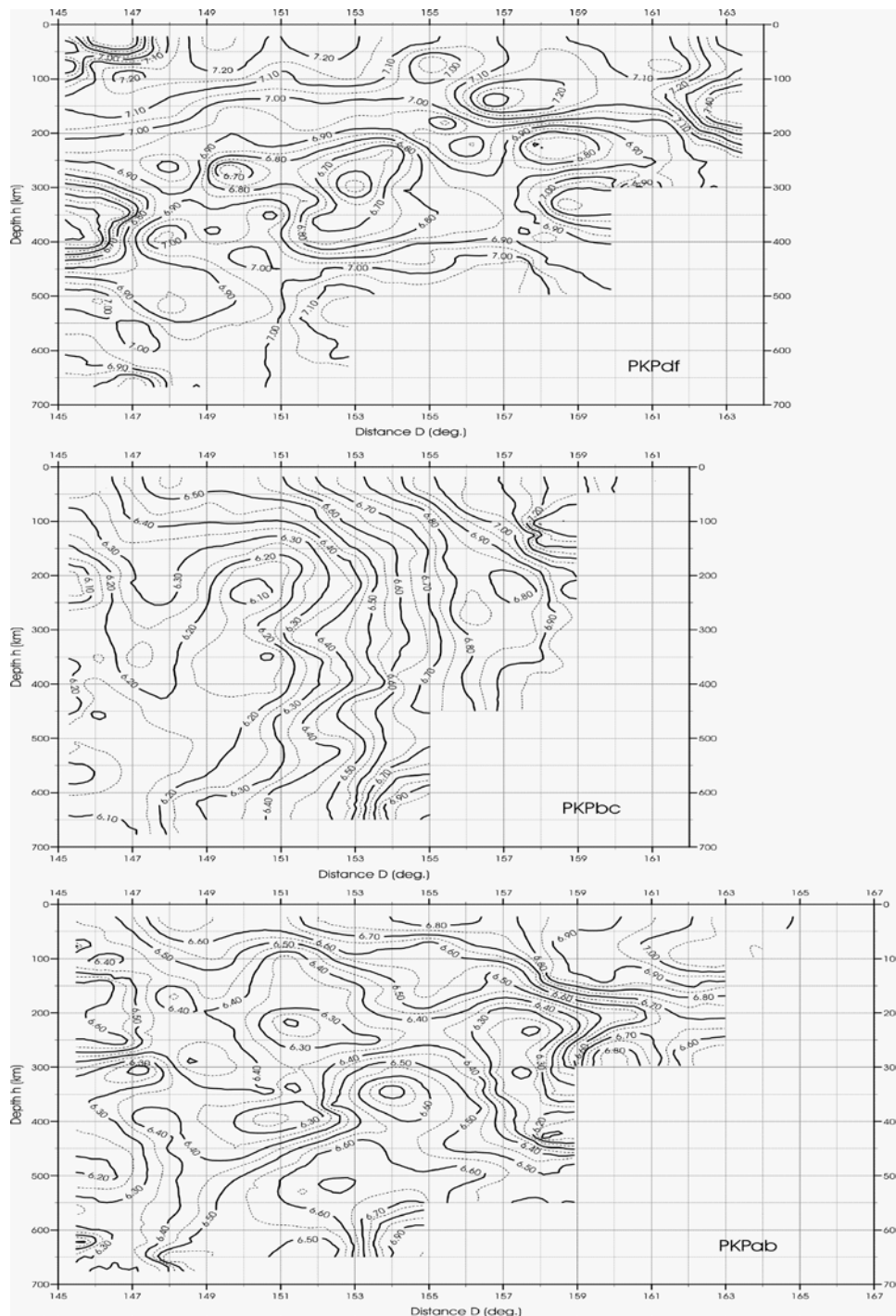
where the term 0.0046 is  $\gamma$  times  $\log_{10}(e)$  ( $= 0.4343$ ), and  $\gamma$ , the attenuation coefficient.  $\gamma$  was determined empirically by Rezapour and Pearce (1998) to be 0.0105 using a very large data set. Surely, this relationship would also be more appropriate than formula (7) for calculating standard  $M_{s\_20}$ , however, recommendations about new calibration relationships have been beyond the mandate of the current IASPEI Working Group on Magnitude Measurements. Maybe, this matter will be taken up again after several years of strict standard  $M_{s\_20}$  measurements with reduced procedure-dependent errors.

## 5 An experimental calibration function for PKP waves

An experimental calibration function for magnitude determinations, based on short-period vertical-component readings of various PKP phases in the distance range  $145^\circ$  to  $164^\circ$ , has been developed by Wendt (see Bormann and Wendt, 1999). The calibration curves are presented in Figure 3 and  $m_b$  is calculated with the following formula (with  $A$  in  $\mu m$ ):

$$m_b(\text{PKP}) = \log_{10} (A/T) + Q(\Delta, h)_{\text{PKPab,bc,df}}. \quad (13)$$

Extensive use of this relationship at station CLL proved that  $m_b$  determinations from core phases are possible with a standard deviation of less than  $\pm 0.2$  magnitude units as compared to P-wave  $m_b$  determinations by NEIC and ISC. If more than one PKP phase can be identified and  $A$  and  $T$  been measured then the average value from all individual magnitude determinations provides a more stable estimate. The applicability of these calibration functions should be tested with data from other stations of the world-wide network.



**Figure 3** Calibration functions according to S. Wendt for the determination of mb(PKP) for PKPdf, PKPbc and PKPab (see Bormann and Wendt, 1999).

## 6 Other magnitude scales

Besides the above mentioned magnitude scales there exist many other ones. Yet, most of them are less (or no longer) widely used, often only within local networks (such as coda or duration magnitudes), or based on specific instrumentation (e.g., strong-motion magnitudes), or aimed at assessing the “size” of specific earthquake related phenomena (such as tsunami magnitudes). Others are of much younger origin and still in the experimental stage, or highly

based on theory and requiring much more complicated algorithms and procedures than the above described “classical” magnitudes. But in any event, all such alternative or complementary magnitudes have to be scaled in one or the other way to the pioneering magnitudes described above, first and foremost to  $M_L$ ,  $M_s$ ,  $mB$  or  $mb$ . For more details see Chapter 3 of NMSOP-2 and encyclopedia review papers by Bormann (2011) and Bormann and Saul (2009).

Most important amongst these alternative magnitudes are the (supposedly) non-saturating and more physically based moment-magnitude scale  $M_w$  and the complementary energy magnitude scale  $M_e$ . The former has even become – not fully justified when taking the advantages of  $M_e$  into account – the preferred magnitude scale for “unifying” earthquake catalogs as well as for seismic hazard assessment. Yet, the procedures for determining  $M_w$  and  $M_e$  are not yet standardized and too sophisticated to be routinely applicable by ordinary analysts at seismological stations or smaller analysis centres. However, software is already offered which calculates these magnitudes in near real-time based on the online-retrieval of wave-form data from regional and global virtual seismic networks (see Chapter 8 and IS 8.3). But these are black-box systems with usually no more analysis personnel involved. In Chapter 3 of this Manual the procedures for calculating the seismic moment tensor, and  $M_w$  as part of it, as well as the released seismic wave energy and related  $M_e$ , are described in detail, while Bormann and Di Giacomo (2011) outline in detail the common roots, differences and benefits of complementary use of  $M_w$  and  $M_e$ .

## References

- Abe, K. (1981). Magnitudes of large shallow earthquakes from 1904 to 1980. *Phys. Earth Planet. Interiors*, **27**, 72-92.
- Abe, K. (1984). Complements to “Magnitudes of large shallow earthquakes from 1904 to 1980).
- Abe, K., and Kanamori, H. (1980). Magnitudes of great shallow earthquakes from 1953-1977. *Tectonophysics*, **62**, 191-203.
- Alsaker, A., Kvamme, L. B., Hansen, R. A., Dahle, A., and Bungum, H. (1991). The  $M_L$  scale in Norway. *Bull. Seism. Soc. Am.*, **81**, 2, 379-389.
- Archangelskaya, V. M. (1959). The dispersion of surface waves in the earth’s crust. *Izv. Akad. Nauk SSSR, Seriya Geofiz.*, **9** (in Russian).
- Bakun, W. H., and Joyner, W. (1984). The  $M_L$  scale in Central California. *Bull. Seism. Soc. Am.*, **74**, 5, 1827-1843.
- Baumbach, M., Bindi, D., Grosser, H., Milkereit, C., Parolai, S., Wang, R., Karkisa, S., Zünbul, S. and Zschau, J. (2003). Calibration of an  $M_L$  scale in northwestern Turkey from 1999 Izmit aftershocks. *Bull. Seism. Soc. Am.*, **93**(5), 2289-2295.
- Bormann, P. (1999). Regional International Training Course 1999 on Seismology, Seismic Hazard Assessment and Risk Mitigation. Lecture and exercise notes. Vol. I and II, GeoForschungsZentrum Potsdam, *Scientific Technical Report STR99/13*, 675 pp.
- Bormann, P. (2011). Earthquake magnitude. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 207-218; doi: 10.1007/978-90-481-8702-7.
- Bormann, P., and Wendt, S. (1999). Identification and analysis of longitudinal core phases: Requirements and guidelines. In: Bormann (1999), 346-366.

- Bormann, P. (2011). Earthquake magnitude. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 207-218; doi: 10.1007/978-90-481-8702-7.
- Bormann, P., and Saul, J. (2009). Earthquake magnitude. In: *Encyclopedia of Complexity and Systems Science*, edited by A. Meyers, Springer, Heidelberg, Vol. 3, 2473-2496.
- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bormann, P., Liu, R., Xu, Z., Ren, K., Zhang, L., Wendt S. (2009). First application of the new IASPEI teleseismic magnitude standards to data of the China National Seismographic Network, *Bull. Seism. Soc. Am.* **99** (3), 1868-1891; doi: 10.1785/0120080010
- Chávez, E., and Priestley, K. F. (1985).  $M_L$  observations in the great basin and  $M_o$  versus  $M_L$  relationships for the 1980 Mammoth Lakes, California, earthquake sequence. *Bull. Seism. Soc. Am.*, **75**, 6, 1583-1598.
- Evernden, J. F. (1971). Variation of Rayleigh-wave amplitude with distance, *Bull. Seism. Soc. Am.*, **61**, 2, 231-240.
- Gellert, R. J., and Kanamori, H. (1977). Magnitudes of great shallow earthquakes from 1904 to 1952). *Bull. Seism. Soc. Am.*, **67**(3), 587-598.
- Gorbunova, I. V., and Kondorskaya, N. V. (1977). Magnitudes in the seismological practice of the USSR. *Izv. Akad. Nauk SSSR, ser Fizika Zemli*, No. 2, Moscow (in Russian).
- Granville, J. P., Richards, . G., Kim, W.-Y., and Sykes, L. R. (2005). Understanding the differences between three teleseismic  $m_b$  scales. *Bull. Seism. Soc. Am.*, **95**(5), 1809-1824.
- Greenhalgh, S. A., and Singh, R. (1986). A revised magnitude scale for South Australian earthquakes. *Bull. Seism. Soc. Am.*, **76**, 3, 757-769.
- Gutenberg, B. (1945a). Amplitudes of surface waves and magnitudes of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 3-12.
- Gutenberg, B. (1945b). Amplitudes of P, PP, and S and magnitude of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 57-69.
- Gutenberg, B. (1945c). Magnitude determination of deep-focus earthquakes. *Bull. Seism. Soc. Am.*, **35**, 117-130.
- Gutenberg, B., and Richter, C. F. (1954). Seismicity of the Earth and associated Phenomena. 2<sup>nd</sup> edition, Princeton University Press, 310 pp.
- Gutenberg, B., and Richter, C. F. (1956). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Herak, M., and Herak. D. (1993). Distance dependence of  $M_S$  and calibrating function for 20 s Rayleigh waves. *Bull. Seism. Soc. Am.*, **83**, 6, 1881-1892.
- Hunter, R. N. (1972). Use of LPZ for magnitude. In: *NOAA Technical Report ERL 236-ESL21*, J. Taggart, Editor, U.S. Dept. Commerce, Boulder, Colorado.
- Hutton, L. K., and Boore, D. M. (1987). The  $M_L$  scale in Southern California. *Bull. Seism. Soc. Am.*, **77**, 6, 2074-2094.
- IASPEI (2011).  
[http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf), last accessed December 2011.
- Kiratzí, A. A., and Papazachos, B. C. (1984). Magnitude scales for earthquakes in Greece. *Bull. Seism. Soc. Am.*, **74**, 3, 969-985.



- Langston, C. A., Brazier, R., Nyblade, A. A., and Owens, T. J. (1998). Local magnitude scale and seismicity rate for Tanzania, East Africa. *Bull. Seism. Soc. Am.*, **88**, 3, 712-721.
- Lienkaemper, J. J. (1984). Comparison of two surface-wave magnitude scales: M of Gutenberg and Richter (1954) and Ms of "Preliminary Determination of Epicenters". *Bull. Seism. Soc. Am.*, **74**(6), 2357-2378.
- Muco, B., and Minga, P. (1991). Magnitude determination of near earthquakes for the Albanian network. *Bolletino di Geofisica Teorica ed Applicata.*, **XXXIII**, 129, 17-24.
- Pechmann, J.C., Nava, S.J., Terra, F.M., and Bernier, J.C. (2007). Local magnitude determinations for Intermountain Seismic Belt earthquakes from broadband digital data, *Bull. Seism. Soc. Am.*, **97**, 557-574.
- Rezapour, M., and Pearce, R. G. (1998). Bias in surface-wave magnitude  $M_S$  due to inadequate distance corrections. *Bull. Seism. Soc. Am.*, **88**, 1, 43-61.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bull. Seism. Soc. Am.*, **25**, 1-32.
- Richter, C. F. (1958). Elementary seismology. *W. H. Freeman and Company*, San Francisco and London, viii + 768 pp.
- Saunders, I., Ottemöller, L., Brandt, M. B. C., and Fourie, C. J. S. (2011). Calibration of an ML scale for Sout Africa using tectonic earthquake data recorded by the South African National Seismograph Network: 2006 to 2009.
- Soloviev, S. L. (1955). Classification of earthquakes in order of energy (in Russian). *Trudy Geofiz. Inst. AN SSSR*, **30**, 157, 3-31.
- Stange, St. (2006).  $M_L$  determination for local and regional events using a sparse network in Southwestern Germany. *J. Seismol.*, **10**, 247-257.
- Strauch, W., and Wylegalla, K. (1989). Potsdam network magnitudes of the 1985/86 Vogtland earthquakes and determination of Q from Sg-amplitudes and coda waves. In: Bormann, P. (Ed.), Monitoring and analysis of the earthquake swarm 1985/86 in the region Vogtland/Western Bohemia, Akad. Wiss. DDR, Zentralinstitut für Physik der Erde, Veröffentl. No. 110, 101-108.
- Uhrhammer, R.A., Helweg, M., Hutton, K., Lombard, P., Walters, A.W., Hauksson, E., and Oppenheimer, D. (2011). California Integrated Seismic Network (CISN) Local Magnitude Determination in California and Vicinity, *Bull. Seism. Soc. Am.*, **101**, 2685-2693.
- Vaněk, J., Zapotek, A., Karnik, V., Kondorskaya, N.V., Riznichenko, Yu.V., Savarensky, E.F., Solov'yov, S.L., and Shebalin, N.V. (1962). Standardization of magnitude scales. *Izvestiya Akad. SSSR., Ser. Geofiz.*, **2**, 153-158.
- Veith, K. F., and Clawson, G. E. (1972). Magnitude from short-period P-wave data. *Bull. Seism. Soc. Am.*, **62**, 435-452.
- Vidal, A., and Munguía, L. (1999). The ML scale in Northern Baja California, México. *Bull. Seism. Soc. Am.*, **89**(3), 750-763.
- Von Seggern, D. (1977). Amplitude-distance relation for 20-second Rayleigh waves. *Bull. Seism. Soc. Am.*, **67**, p. 405-411.
- Wahlström, R., and Strauch, W. (1984). A regional magnitude scale for Central Europe based on crustal wave attenuation. *Seismological Dep. Univ. of Uppsala, Report No. 3-84*, 16 pp.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.

| Topic   | Magnitude determinations                                                                                                                                   |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version | October 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_EX_3.1">10.2312/GFZ.NMSOP-2_EX_3.1</a>                                                     |

## 1 Aim

The exercises aim at making you familiar with the measurement of seismic amplitudes and periods in analog and digital records and the determination of related **classical magnitude** values for local and teleseismic events by using the related procedures and relationships outlined in Chapter 3, and the relevant classical magnitude calibration functions given in DS 3.1. For international measurement standards, recently adopted by IASPEI (2011) for local, regional and teleseismic magnitudes, see IS 3.3 and [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf). Guidelines and waveform data for an interactive training to measure these standard amplitude, period and magnitudes parameters by using the analysis program SEISAN are given in IS 3.4. This IS will be updated in future when other analysis programs suitable for measuring the standard data become available.

## 2 Procedures

The general relationship 
$$M = \log (A/T) + \sigma(\Delta, h) \quad (1)$$

is used for magnitude determination, with A – maximum “ground motion” amplitude in  $\mu\text{m}$  ( $10^{-6}$  m) or nm ( $10^{-9}$  m), respectively, measured for the considered wave group, T – period of that maximum amplitude in seconds. Examples, how to measure the related trace amplitudes B and period T in seismic records are depicted in Fig. 3.9. Trace amplitudes B have to be divided by the respective magnification  $\text{Mag}(T)$  of the seismograph at the considered period T in order to get the “ground motion” amplitude in either  $\mu\text{m}$  or nm, i.e.,  $A = B/\text{Mag}(T)$ .

$\sigma(\Delta, h) = -\log A_0$  is the magnitude calibration function where  $\Delta$  = epicentral distance and h = source depth. For teleseismic body waves the calibration function is usually termed  $Q(\Delta, h)$ , according to Gutenberg and Richter (1956), who published respective calibration values for medium-period vertical as well as horizontal component amplitude readings of P and PP waves and for horizontal component S-wave amplitudes. Another calibration function for short-period vertical component P-wave amplitude readings, termed  $P(\Delta, h)$ , has been published by Veith and Clawson (1972). For *teleseismic events* (distance >1000-2000 km)  $\Delta$  is generally given in degree ( $1^\circ = 111.19$  km), for *local/near regional events* (distance usually <1000 km), however, in km. For local events often the “slant range” or hypocentral distance R (in km) is used instead of  $\Delta$ . All calibration functions used in the exercise are given in DS 3.1.

**Note:** According to the original definition of the local magnitude scale  $M_l$  by Richter (1935) only the *maximum trace amplitude* B in mm as recorded in standard records of a Wood-Anderson seismometer is measured (see Fig. 3.11 and section 3.2.4), i.e.,

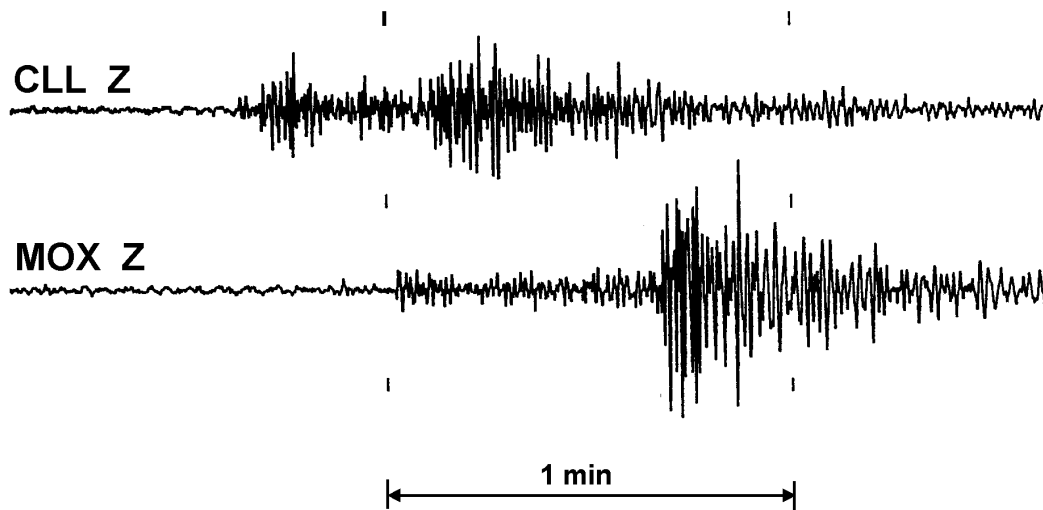
$$M_l = \log B(\text{WA}) - \log A_0(\Delta). \quad (2)$$

Accordingly, no period  $T$  is measured, and no conversion to “ground motion” amplitude is made. However, when applying  $M_L$  calibration functions to trace amplitudes  $B$  measured (in mm too) in records of another seismograph (SM) with a frequency-magnification curve  $\text{Mag}(T)$  different from that of the Wood-Anderson seismograph (WA) then this frequency-dependent difference in magnification has to be corrected. Equation (2) then becomes

$$(3) \quad M_L = \log B(\text{SM}) + \log \text{Mag}(\text{WA}) - \log \text{Mag}(\text{SM}) - \log A_0.$$

### 3 Data

The data used in the exercise are given in the following figures and tables.

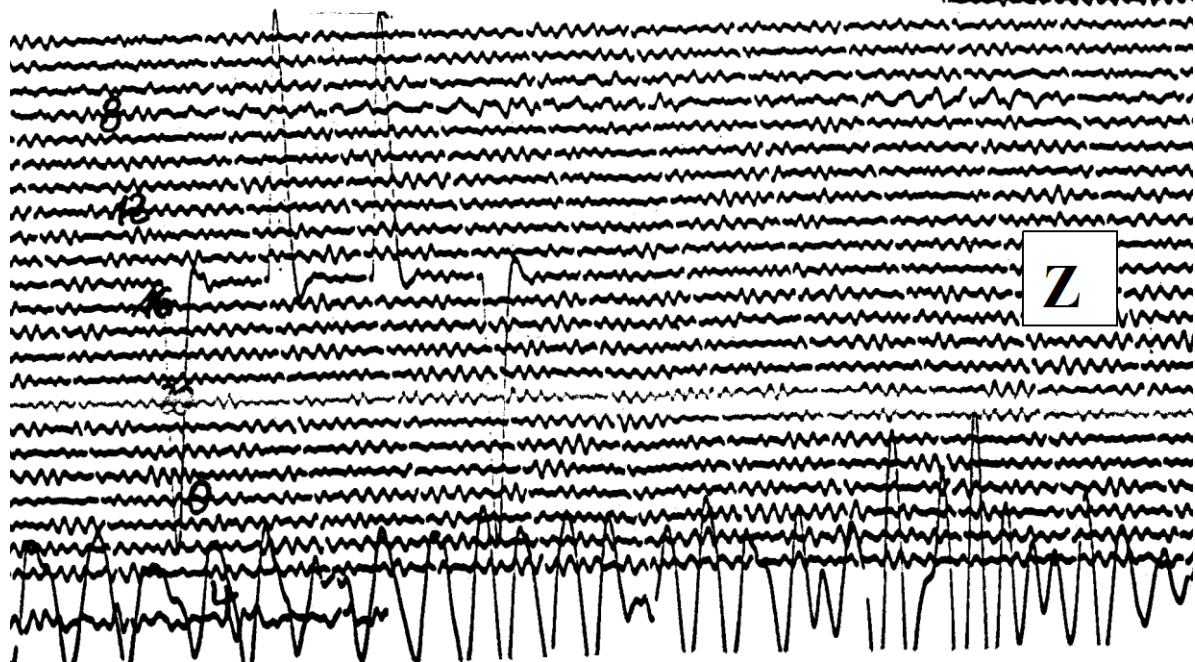
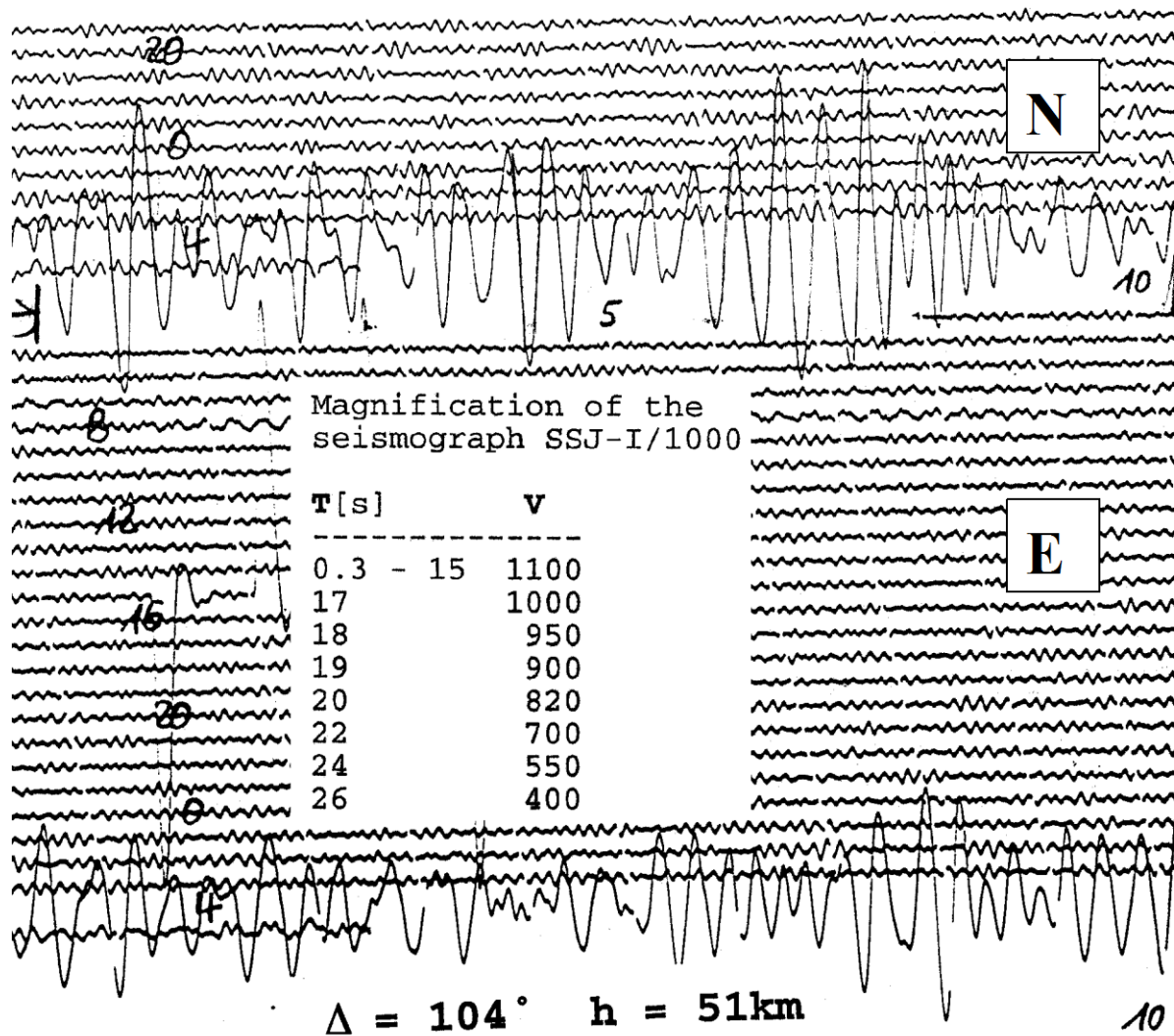


**Figure 1** Vertical component records of a local seismic event in Poland at the stations CLL and MOX in Germany at scale 1:1. The magnification values  $\text{Mag}(\text{SM})$  as a function of period  $T$  for this short-period seismograph are given in Table 1 below.

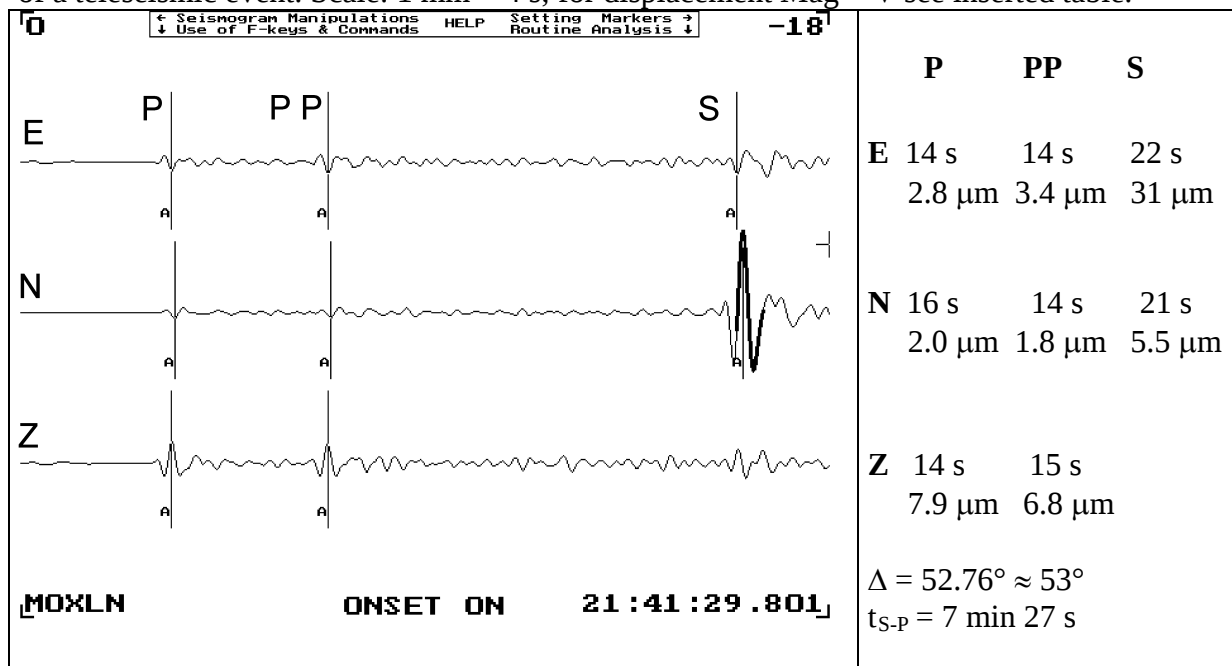
**Table 1** Magnification values  $\text{Mag}(\text{SM})$  as a function of period  $T$  in s for the short-period seismograph used for the records in Figure 1 together with the respective values  $\text{Mag}(\text{WA})$  for the Wood-Anderson standard seismograph for  $M_L$  determinations.

| $T$ (in s) | $\text{Mag}(\text{SM})$ | $\text{Mag}(\text{WA})$ | $T$ (in s) | $\text{Mag}(\text{SM})$ | $\text{Mag}(\text{WA})$ |
|------------|-------------------------|-------------------------|------------|-------------------------|-------------------------|
| 0.1        | 35,000                  | 2,800                   | 1.1        | 190,000                 | 1,100                   |
| 0.2        | 92,000                  | 2,700                   | 1.2        | 180,000                 | 950                     |
| 0.3        | 125,000                 | 2,600                   | 1.3        | 170,000                 | 850                     |
| 0.4        | 150,000                 | 2,400                   | 1.4        | 155,000                 | 750                     |
| 0.5        | 170,000                 | 2,200                   | 1.5        | 140,000                 | 700                     |
| 0.6        | 190,000                 | 2,000                   | 1.6        | 120,000                 |                         |
| 0.7        | 200,000                 | 1,800                   | 1.7        | 90,000                  |                         |
| 0.8        | 201,000                 | 1,600                   | 1.8        | 80,000                  |                         |
| 0.9        | 201,000                 | 1,400                   | 1.9        | 70,000                  |                         |

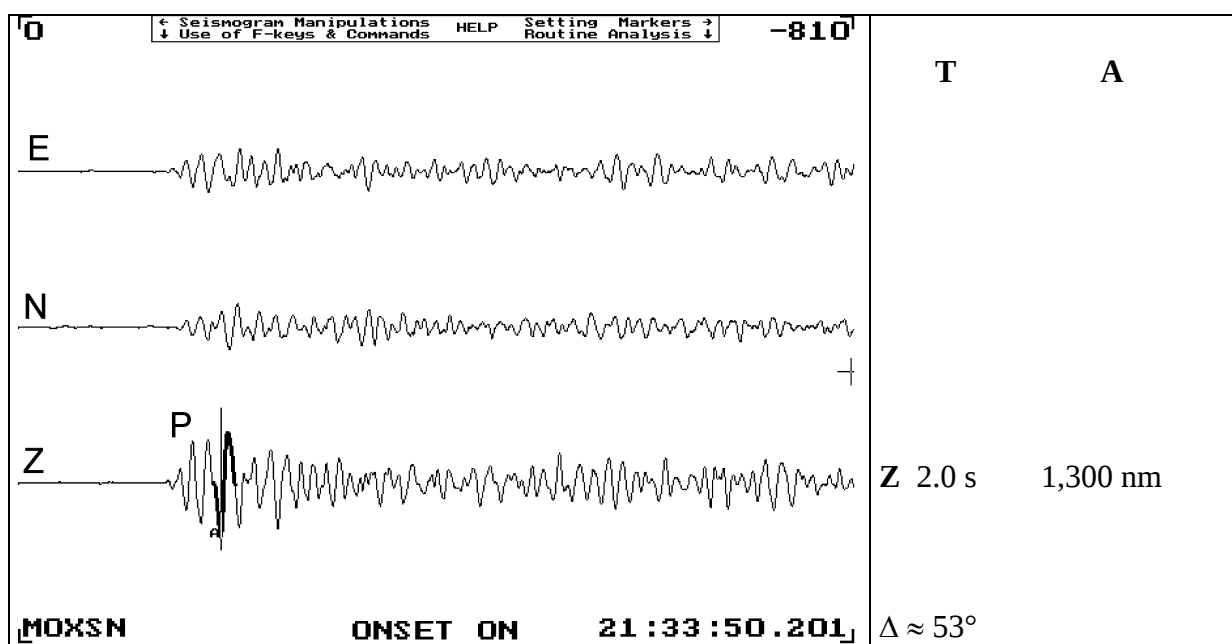
|     |         |       |     |        |  |
|-----|---------|-------|-----|--------|--|
| 1.0 | 200,000 | 1,200 | 2.0 | 60,000 |  |
|-----|---------|-------|-----|--------|--|



**Figure 2** Analog record at scale 1:1 of a Kirnos-type seismograph from a surface-wave group of a teleseismic event. Scale: 1 mm = 4 s; for displacement  $\text{Mag} = V$  see inserted table.



**Figure 3** Display of the long-period (10 to 30 s) filtered section of a broadband 3-component record of the Uttarkasi earthquake in India (19 Oct. 1991;  $h = 10 \text{ km}$ ) at station MOX in Germany. The record traces are, from top to bottom: E, N, Z. Marked are the positions, from where the computer program has determined automatically the ground displacement amplitudes  $A$  and related periods  $T$  for the onsets (from left to right) of P, PP and S. For S the respective cycle is shown as a bold trace. The respective values of  $A$  and  $T$  for all these phases are saved component-wise in the data-pick file. They are reproduced in the box on the right together with the computer picked onset-time difference  $S - P$  and the epicentral distance  $\Delta$  as published for station MOX by the ISC.



**Figure 4** As for Figure 3, however short-period (0.5 to 3 Hz) filtered record section of the P-wave group only. Note that the amplitudes are given here in nm ( $10^{-9}$  m).

## 4 Tasks

### Task 1:

- 1.1 Identify and mark in the records of Figure 1 the onsets Pn, Pg and Sg. Note that Pn amplitudes are for distances  $< 400$  km usually smaller than that of Pg! Then determine the hypocentral (“slant”) distance  $R$  in km from the rule of thumb  $t(\text{Sg-Pg})[\text{s}] \times 8 = R [\text{km}]$  for both station CLL and MOX. Note, that for this shallow event the epicentral distance  $\Delta$  and the hypocentral distance  $R$  are practically the same.
- 1.2 Determine for the stations CLL and MOX the max. trace amplitude  $B(\text{SM})$  and related period  $T$  and then, according to Equation (3), the equivalent log trace amplitude when recorded with a Wood-Anderson seismometer, i.e.,  $\log B(\text{WA}) = \log B(\text{SM}) + \log \text{Mag}(\text{WA}) - \log \text{Mag}(\text{SM})$ .
- 1.3 Use Equation (2) above, and the values determined under task 1.2, for determination of the local magnitude  $M_L$  for the event in Figure 1 for both station CLL and MOX using the calibration functions  $-\log A_0$ :
  - a) by Richter (1958) for California (see Table 1 in DS 3.1);
  - b) by Kim (1998) for Eastern North America (vertical comp.; see Table 2 in DS 3.1);
  - c) by Alsaker et al. (1991) for Norway (vertical comp., see Table 2 in DS 3.1).
- 1.4 Discuss the differences in terms of:
  - a) differences in regional attenuation in the three regions from which  $M_L$  calibrations functions were used;
  - b) amplitude differences within a seismic network;
  - c) uncertainties of period reading in analog records with low time resolution and thus uncertainties in the calculation of the equivalent Wood-Anderson trace amplitude  $B(\text{WA})$ .

### Task 2:

- 2.1 Measure the *maximum* horizontal and vertical trace amplitudes  $B$  in mm and related periods in s from the 3-component surface-wave records in Figure 2. Note, that the maximum horizontal component has to be calculated by combining vectorially  $B_N$  and  $B_E$ , measured at the same record time, i.e.,  $B_H = \sqrt{B_N^2 + B_E^2}$ .
- 2.2 Calculate the respective maximum ground amplitudes  $A_H$  and  $A_V$  (in  $\mu\text{m}$ ; vertical  $V = Z$ ) by taking into account the period-dependent magnification of the seismograph (see table inserted in Figure 2)
- 2.3 Calculate the respective surface-wave magnitudes  $M_s$  according to the calibration function
  - a)  $\sigma(\Delta)$  as published by Richter (1958) (see Table 3 in DS 3.1, for horizontal component  $H$  only);
  - b)  $\sigma(\Delta)$  as given by the Prague-Moscow formula  $M_s = \log(A/T)_{\max} + 1.66 \log \Delta + 3.3$  (Vaněk et al., 1962), which has been accepted by IASPEI in 1967 as the standard

formula for surface-wave magnitude determinations. Apply it to both the horizontal and the vertical component amplitude and period readings.

**Note:** Differentiate between surface-wave magnitudes from horizontal and vertical component records by annotating them unambiguously as MLH and MLV, respectively.

### Task 3:

Use the computer determined periods and amplitudes given in the right boxes of Figures 3 and 4 for the body-wave phases P, PP and S recorded from the shallow ( $h = 10$  km) teleseismic earthquake in India in order to determine the respective body-wave magnitudes according to the general relationship (1):

3.1 Compare the epicentral distance calculated by the ISC for MOX ( $\Delta = 52.76^\circ$ ) with your own quick determination of  $\Delta$  using the “rule of thumb”  $\Delta$  (in  $^\circ$ ) =  $[t_{S-P}(\text{in min}) - 2] \times 10$ .

3.2 Compare the differences in  $Q(\Delta)$  according to Table 7 in DS 3.1 when using the “exact” distance given by the ISC with your quick “rule of thumb” estimation of  $\Delta$ . Assess the influence of the distance error on the magnitude estimate and draw conclusions.

3.3 Calculate MPV, MPH; MPPV, MPPH and MSH using the calibration functions  $Q(\Delta)$  given in Table 7 of DS 3.1, the amplitude-period values given in Figure 3 and  $\Delta = 53^\circ$ . Discuss the degree of agreement/disagreement and possible reasons.

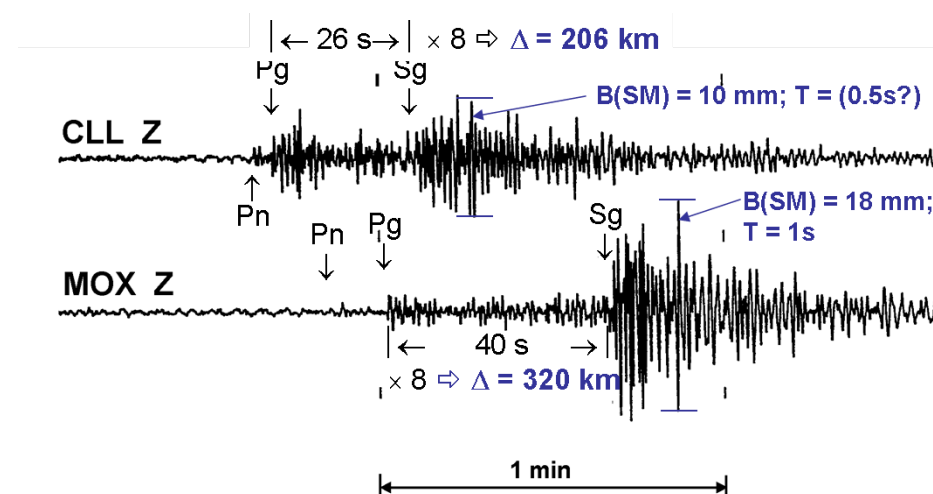
3.4 Calculate  $m_b$  for the short-period P-wave recording in Figure 4 using

- $Q(\Delta, h)$  as depicted in Figure 1a of DS 3.1 for the vertical component of P and
- $P(\Delta, h)$  as depicted in Figure 2 of DS 3.1.
- Discuss the difference between a) and b).

## 5 Solutions

**Note:** Your individual readings of times, periods and amplitudes should not deviate more than 10 % and your magnitude estimates should be within about  $\pm 0.2$  units of the values given below.

### Task 1:



- 1.1 CLL  $t(\text{Sg-Pg}) = 26 \text{ s}$   $R = 208 \text{ km}$   
 MOX  $t(\text{Sg-Pg}) = 40 \text{ s}$   $R = 320 \text{ km}$
- 1.2 CLL:  $B(\text{MS}) = 10 \text{ mm}$   $T = (0.5\text{s}?)$   $\log B(\text{WA}) = -0.888$   
 MOX:  $B(\text{MS}) = 18 \text{ mm}$   $T = 1 \text{ s}$   $\log B(\text{WA}) = -0.967$
- 1.3 CLL:  $M_l(\text{Richter}) = 2.7$   $M_l(\text{Kim}) = 2.6$   $M_l(\text{Alsaker}) = 2.4$   
 MOX:  $M_l(\text{Richter}) = 3.1$   $M_l(\text{Kim}) = 2.8$   $M_l(\text{Alsaker}) = 2.6$

1.5 California is a tectonically younger region and with higher heat flow than Eastern North America and Scandinavia. Accordingly, seismic waves are more strongly attenuated with distance. This has to be compensated by larger magnitude calibration values –  $\log A_0$  for California. But even within a seismic network amplitude variations may be, depending on different conditions in local underground and azimuth dependent wave propagation, in the order of a factor 2 to 3 in amplitude, thus accounting for magnitude differences in the order of up to about  $\pm 0.5$  magnitude units between the various stations. This scatter can be reduced by determining station corrections for different source regions. Note also, that the period reading is rather uncertain for CLL. If we assume, as for MOX, also  $T = 1 \text{ s}$  then  $\log B(\text{WA}) = -1.222$ , i.e., the magnitudes values for CLL would be even smaller by 0.3 units.

### Task 2:

- 2.1  $B_N = 20.5 \text{ mm}$ ,  $T_N = 22 \text{ s}$ ;  $B_E = 12 \text{ mm}$ ,  $T_N = 20 \text{ s}$   $\rightarrow B_H = 23.8 \text{ mm}$ ,  $\bar{T} = 21 \text{ s}$   
 $B_Z = 23 \text{ mm}$ ,  $T_Z = 18 \text{ s}$
- 2.2  $A_H = 31.3 \mu\text{m}$  for  $T = 21 \text{ s}$   $A_Z = A_V = 24.2 \mu\text{m}$  for  $T = 18 \text{ s}$
- 2.3 a)  $M_s = \text{MLH}(\text{Richter}) = \log A_{H\text{max}} + \sigma(\Delta)_{\text{Richter}} = 1.5 + 5.15 = 6.65$   
 b)  $M_s = \log(A/T)_{\text{max}} + 1.66 \log \Delta + 3.3 \rightarrow M_s = \text{MLH} = 6.82$  and  
 $M_s = \text{MLV} = 6.78$

**Discussion:** For the given record example  $M_s(\text{Richter}, 1958)$  yields somewhat smaller values than  $M_s(\text{Vaněk et al., 1962})$ . The latter formula, although derived for horizontal component surface-wave amplitude readings, yields comparable magnitude values also when applied to vertical component readings (here difference  $< 0.05$  magnitude units).

### Task 3:

- 3.1 The “rule of thumb” yields  $\Delta = 54.5^\circ$ . This is only  $1.7^\circ$  off the ISC determination. Generally, the rule-of-thumbs allows to estimate  $\Delta$  in the range  $25^\circ < \Delta < 100^\circ$  with an error not larger than  $\pm 2.5^\circ$ .
- 3.2 The deviations in  $Q(\Delta)$  and thus between magnitude estimates based on either  $\Delta$  values from NEIC/ISC calculations or quick S-P determinations at the individual stations and using the “rule of thumb” are generally less than 0.2 units. They are even smaller, when correct travel-time (difference) curves are available. This permits sufficiently accurate quick teleseismic magnitude estimates at individual stations even with very modest tools



and without prior knowledge of the event locations and distance determinations of the world data centers.

3.3 MPV = 6.45, MPH = 6.36, MPPV = 6.36, MPPH = 6.24; MSH = 6.75

The magnitude values for vertical and horizontal component amplitude readings of P and PP, respectively, agree within 0.1 magnitude units. This speaks of a good scaling of the respective  $Q(\Delta)$  calibration functions. MSH is in the given case significantly larger. This is obviously related to the different azimuthal radiation pattern for P and S waves and was one of the reasons, why Gutenberg strongly recommended the determination of the body-wave magnitudes for all these phases and averaging them to a unified body-wave magnitude value  $m$ . The latter provides more stable and less azimuth-dependent individual magnitude estimates.

3.4 a)  $Q_{PZ}(53^\circ, 10 \text{ km}) = 7.0$ ,  $\log(A/T) = -0.19$  (with  $A$  in  $\mu\text{m}$ !)  $\rightarrow$  MPV = mb = 6.8

b)  $P_Z(53^\circ; 10 \text{ km}) = 3.4$ ,  $\log(A/T) = 3.1$  (**with 2A in nm!**)  $\rightarrow$  MPV = mb = 6.5

c)  $P_Z(\Delta, h)$  yields, for the same ratio  $\log(A/T)$ , slightly lower magnitude values as compared to  $Q_{PZ}(\Delta, h)$ , for deep events, in particular. This also applies for other distance ranges (see Figures 1a and 2 in DS 3.1). Note that  $P_Z$ , although specifically developed for the calibration of short-period P-wave amplitudes, is not yet a recommended standard calibration functions.

## References

- Alsaker, A., Kvamme, L. B., Hansen, R. A., Dahle, A., and Bungum, H. (1991). The  $M_L$  scale in Norway. *Bull. Seism. Soc. Am.*, **81**, 2, 379-389.
- IASPEI (2011). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. Preliminary version October 2005, updated version September 2011; [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf).
- Gutenberg, B., and Richter, C. F. (1956). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Kim, W.-Y. (1998). The  $M_L$  scale in Eastern North America. *Bull. Seism. Soc. Am.*, **88**, 4, 935-951.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bull. Seism. Soc. Am.*, **25**, 1-32.
- Richter, C. F. (1958). Elementary seismology. W. H. Freeman and Company, San Francisco and London, viii + 768 pp.
- Vaněk, J., Zapotek, A., Karnik, V., Kondorskaya, N.V., Riznichenko, Yu.V., Savarensky, E.F., Solov'yov, S.L., and Shebalin, N.V. (1962). Standardization of magnitude scales. *Izvestiya Akad. SSSR, Ser. Geofiz.*, **2**, 153-158.
- Veith, K. F., and Clawson, G. E. (1972). Magnitude from short-period P-wave data. *Bull. Seism. Soc. Am.*, **62**, 435-452.



| Topic   | Determination of fault-plane solutions by hand                                                                                                                                                                                                                                                                                                                                                                              |
|---------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <b>Peter Bormann</b> and <b>Michael Baumbach</b> (✉)(formerly GeoForschungsZentrum Potsdam, Department 2: Physics of the Earth, Telegrafenberg, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> ; <b>Siegfried Wendt</b> , Geophysical Observatory Collm, University of Leipzig, D-04779 Wermsdorf, Germany, E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a> |
| Version | June 2012; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_EX_3.2">10.2312/GFZ.NMSOP-2_EX_3.2</a>                                                                                                                                                                                                                                                                                                                         |

## 1 Aim

Fault-plane solutions are nowadays routinely calculated and plotted by a diversity of software programs. They may result from using different types of seismic waves, recorded in different frequency bands and ranges of epicentral distances and by measuring and analyzing different record parameters. Accordingly, such solutions, may be based on more sophisticated optimal waveform fitting between real and synthesized records, on considering azimuthal variations in amplitude ratios between P and S waves, or on simple “classical” P-wave first-motion polarity readings only. The latter is still the most common procedure for masses of data, especially when recorded in the local distance and low magnitude range.

However, the under-laying assumptions, sequential procedural steps and ranges of possible errors of fault-plane solutions are often not sufficiently well known to the common user of such data and programs. Therefore, it is recommendable that at least once the step-by-step determination of a fault-plane solution by hand, based on first-motion polarity readings, should be part of the educational program of any future seismologist or geoscientist, as an eye-opening and awareness creating exercise. For this we have selected a realistic yet not too simple and data-wise not too ideally constrained example based on recordings of local aftershocks.

The exercise aims at:

- understanding how fault slip affects the polarities of P waves;
- understanding, how the applied crustal model in the source location process may influence the calculated take-off angles as a function of epicentral distance;
- understanding the presentation of P-wave polarities in an equal angle Wulff net or an equal area projection (Lambert-Schmidt net) of the focal sphere;
- constructing a fault-plane solution and determining the related parameters (**P** and **T** axes and displacement vector) for a real earthquake;
- relating the directions of the fault-plane solutions to the tectonic setting and complementary information on the regional stress pattern.

## 2 Data and procedures

Before a fault-plane solution for a teleseismic event can be constructed, the following steps must be completed:

- a) Interpretation of P-wave first-motion polarities from seismograms at several stations;

- b) Calculation of epicentral distances and source-to-station azimuths for these stations;
- c) Calculation of the take-off angles  $\Theta$  for the seismic P-wave rays leaving the hypocenter towards these stations.

Both b) and c) require the knowledge of the focal depth and of the P-wave velocity-depth function along the ray path (see EX 3.3). Commonly, standard Earth velocity models are used (e.g., Kennett, 1991 or AK135, Kennett et al., 1995; see DS 2.1), especially, when teleseismic recordings are analyzed. Most location programs provide both the source-station azimuths and take-off angles in their output files. In the case of local events, however, it is necessary to determine beforehand, which P-wave travel-time branch is arriving first at what distance. This is strongly influenced by the velocity model of the local crust. Therefore, the events should be located, if possible, with a layered velocity model that represents reasonably well the Earth crust of that region. If it deviates significantly from the global model, significantly biased take-off angles from the source may be calculated for the first arriving P-wave rays. This may result in not separable polarity patterns and thus inconsistent or biased fault-plane solutions.

The exercise below is based on related general definitions, relationships and diagrams presented in Chapter 3 of this Manual in the section “Determination of fault-plane solutions.” The starting data for the exercise are summarized in Table 1. They relate to a locally recorded aftershock of the  $M_s = 6.8$  Erzincan earthquake in Turkey (**Main shock**: 13.03.1992, 39.72°N, 39.63°E; **aftershock**: 12.04.1992,  $M_l = 2.8$ , 39.519° N, 39.874° E, source depth  $h = 3$  km). The distances of the recording stations range from 3.7 km to 49.6 km. They were calculated according to steps a)-c) by using the program HYPO71 (Lee and Lahr, 1975).

**Note:** In the special case of a near surface source in a homogeneous or a plane layered velocity model the take-off angles  $\Theta$  from the source agree with the *angle of incidence*  $A_{IN}$  calculated with HYPO71 for the respective seismic ray arriving at a given station. Figure 1 defines  $\Theta$ , which depends on source distance, source depth, velocity model and the recorded first arriving type of P wave. Further, for an average single layer crustal model of 30 to 40 km thickness, all P-wave first arrivals within a distance of about 120 km to < 200 km are  $P_g$  and up-going. That is, they emerge only from the upper half of the focal hemisphere. Also, when using HYPO71 with the average global two-layer crust according to the velocity model IASP91 (Kenneth 1991) only upper hemisphere take-off angles would have been calculated for the first P-wave arrivals up to distances of 50 km. But in the epicentral area under consideration a significant velocity increase in the upper crust was already found at 4 km depth (increase of  $v_p = 5.3$  km/s to 6.0 km/s). Therefore, a single-layer model above half-space for the upper crust had to be used in the calculations. Accordingly, only stations up to 17 km distance were reached by first arriving rays that left through the upper focal sphere. At larger distances the first arriving P waves had left the lower focal sphere. They were refracted back at this upper crustal discontinuity under the critical angle of incidence  $i_{cr}$ , which is in our layer model 62° according to the velocity contrast  $v_1/v_2 = 5.3/6.0 = \sin i_{cr} = \sin \Theta_{cr}$ . These critically refracted P-waves travel with a constant apparent horizontal velocity of 6.0 km/s and therefore arrive at stations beyond 17 km epicentral distance also with a constant incidence angle  $A_{IN_{cr}} = 62^\circ$  (Figure 2).

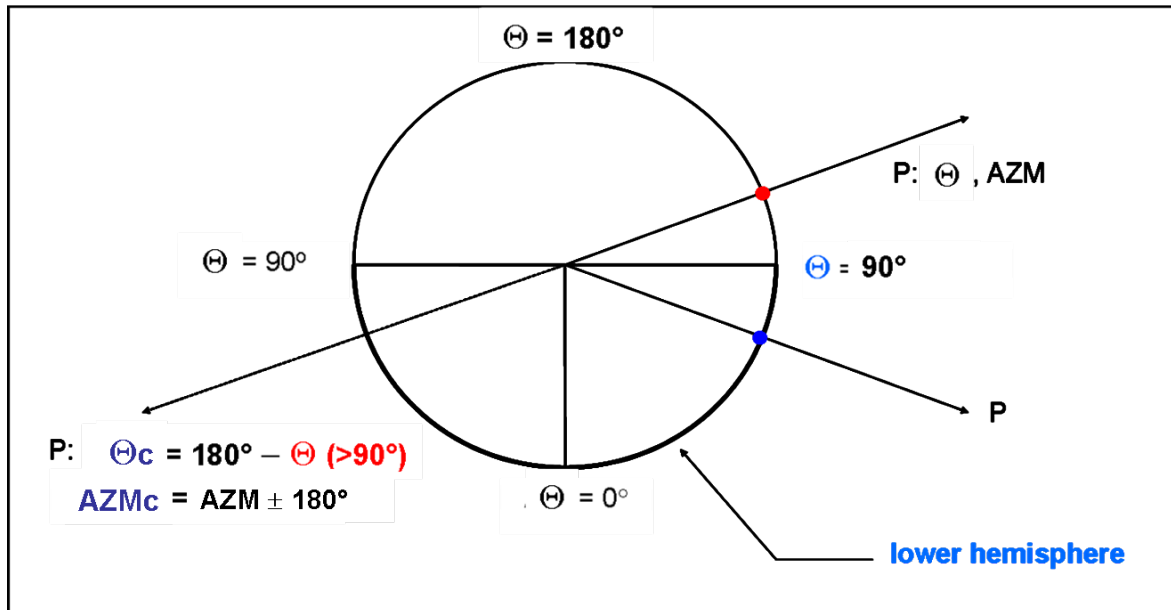
Fault-plane solutions are derived by remapping the penetration points of seismic rays through the upper and lower hemisphere on a suitable horizontal plane net projection and separating from each other sectors of opposite polarity by pairs of orthogonal great circles. However, as

obvious from Figures 1 and 2, the polarities of rays leaving the upper and lower hemisphere with the same azimuth are mirrored. Mapping them both with their originally calculated values  $\Theta$  and azimuth AZM on the same horizontal net projection would result in incompatible polarity plots. Therefore, one has to make up one's mind, whether one uses only upper or lower hemisphere polarities and has then to mirror the other hemisphere data so that they become compatible with the reference hemisphere data.

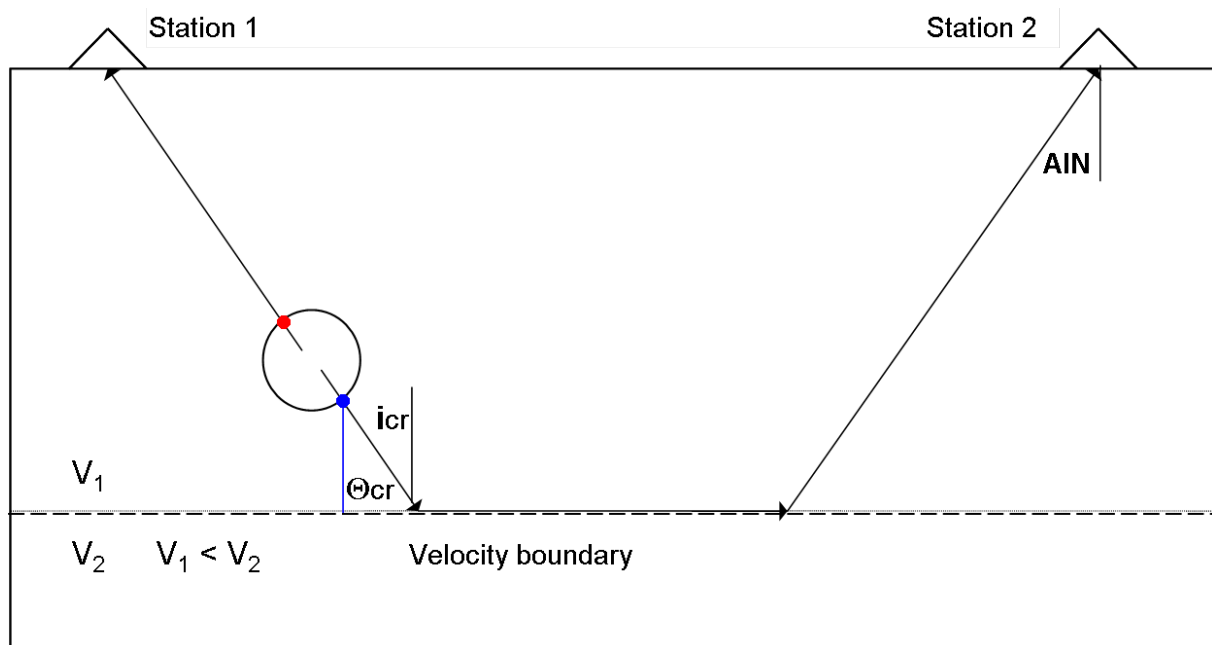
For the exercise we use only lower hemisphere data, as generally used when analyzing teleseismic polarity readings. Thus, for all upper hemisphere polarities with  $90^\circ < \Theta \leq 180^\circ$  corrected values  $\Theta_c$  and  $AZM_c$  have to be calculated by using the relationships given in Figure 1. Only these corrected upper hemisphere polarities can be plotted jointly with the original lower hemisphere polarities ( $0^\circ \leq \Theta \leq 90^\circ$ ) on the same net projection.

**Table 1** Original and corrected (not yet calculated) values of ray azimuth (AZM and  $AZM_c$ ) and take-off angles ( $\Theta$  and  $\Theta_c$ , respectively) towards stations of a temporary network at epicentral distance DIS which recorded the Erzincan aftershock of April 12, 1994. POL - polarity of P-wave first motions. D = **down** = negative (-) first half-wave and U = **up** = positive (+) first half-wave in vertical component records.

| STA | DIST<br>(km) | AZM<br>(degree) | $\Theta$<br>(degree) | POL | $AZM_c$<br>(degree) | $\Theta_c$<br>(degree) |
|-----|--------------|-----------------|----------------------|-----|---------------------|------------------------|
| ALI | 3.7          | 40              | 130                  | D   |                     |                        |
| ME2 | 7.1          | 134             | 114                  | D   |                     |                        |
| KAN | 7.7          | 197             | 112                  | D   |                     |                        |
| YAR | 8.2          | 48              | 111                  | D   |                     |                        |
| ERD | 13.6         | 313             | 103                  | D   |                     |                        |
| DEM | 14.3         | 330             | 102                  | D   |                     |                        |
| GIR | 14.4         | 301             | 102                  | U   |                     |                        |
| UNK | 16.2         | 336             | 101                  | D   |                     |                        |
| SAN | 17.2         | 76              | 62                   | U   |                     |                        |
| PEL | 18.6         | 327             | 62                   | D   |                     |                        |
| GUN | 22.6         | 290             | 62                   | U   |                     |                        |
| ESK | 22.7         | 312             | 62                   | U   |                     |                        |
| SOT | 23.8         | 318             | 62                   | D   |                     |                        |
| BA2 | 27.3         | 79              | 62                   | U   |                     |                        |
| MOL | 28.8         | 297             | 62                   | U   |                     |                        |
| YUL | 29.3         | 67              | 62                   | U   |                     |                        |
| ALT | 31.5         | 59              | 62                   | D   |                     |                        |
| GUM | 32.0         | 320             | 62                   | U   |                     |                        |
| GU2 | 32.0         | 320             | 62                   | D   |                     |                        |
| BAS | 35.6         | 308             | 62                   | D   |                     |                        |
| BIN | 36.4         | 295             | 62                   | U   |                     |                        |
| HAR | 38.7         | 24              | 62                   | D   |                     |                        |
| KIZ | 42.2         | 311             | 62                   | U   |                     |                        |
| AKS | 48.7         | 284             | 62                   | D   |                     |                        |
| SUT | 49.6         | 295             | 62                   | U   |                     |                        |



**Figure 1** Transformation of a ray leaving the focal sphere upwards with a take-off angle  $\Theta$  between  $90^\circ < \Theta \leq 180^\circ$  into an equivalent downward ray with same polarity but corrected take-off angle  $\Theta_c$  and azimuth  $AZM_c$ .



**Figure 2** Two rays, leaving the focal sphere in opposite directions, reach - because of the symmetry of radiation pattern - the stations 1 and 2 with the same polarity. The crossing point of the up-going ray with the focal sphere can, therefore, be remapped according to the formulas given in Figure 1 into a crossing point with the lower hemisphere which coincides with the ray crossing-point for station 2.

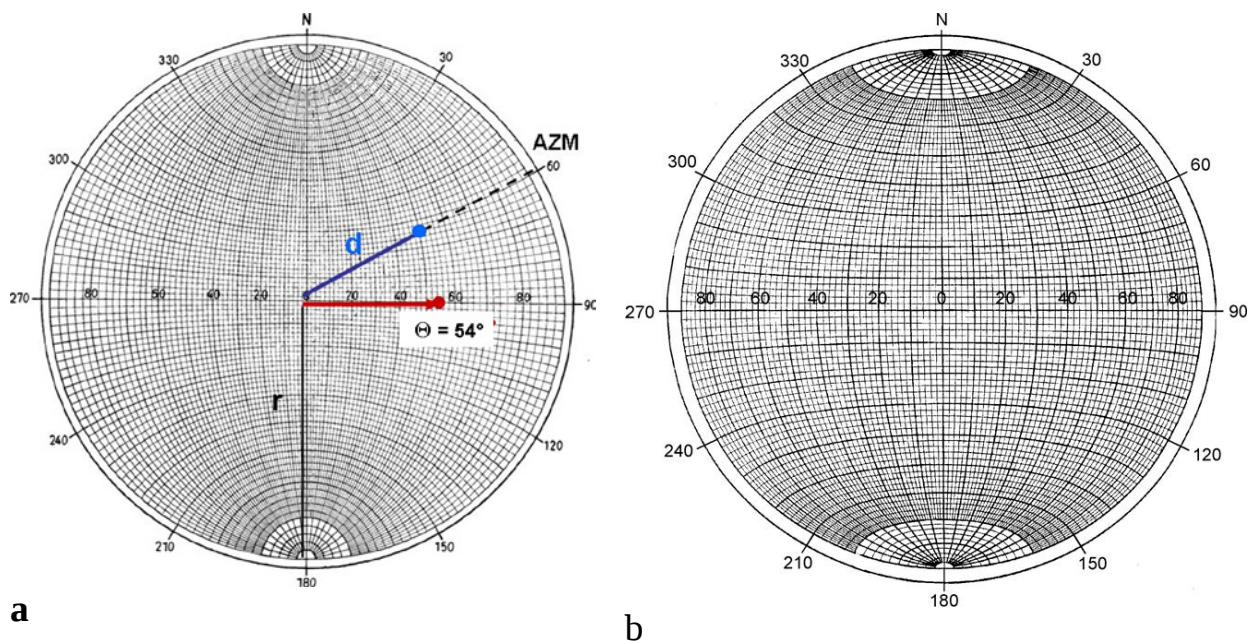
**Note:**  $\Theta$  calculations based on strongly biased velocity models might result in inconsistent fault-plane solutions or not permit a proper separation of opposite polarity readings into quadrants (as in Figure 4) at all.

The primary data in Table 1 were taken from the output file of the program HYPO71 with which the event was located. The first five columns of this file contain, as an example for the two stations ALI and ESK in Table 1, the following data:

| STN | DIST | AZM | AIN | PRMK |
|-----|------|-----|-----|------|
| ALI | 3.7  | 40  | 130 | IPD0 |
| ESK | 22.7 | 312 | 62  | IPU1 |

with STN - station code; DIST - epicentral distance in km; AZM - azimuth towards the station clockwise in degree from north; AIN (**replaced by us by  $\Theta$** ) - take-off angle of the ray leaving the source towards the station, measured as in Figure 1, and calculated for the given structure-velocity model; PRMK - P-wave reading remarks. In the column PRMK P stands for P-wave onset, I for impulsive (sharp) or E for emergent (less clear) onset, D for clear (or – for poor) downward first motion, U for clear (or + for poor) upward first motion as read at the station. The last character may range between 0 and 4 and is a measure of the quality (clarity) of the onset and thus of the weight given to the reading in the calculation procedure, e.g., 4 for zero weight and 0 for full weight. In case of the above two stations the values for ALI would need to be corrected to get the respective values for the equivalent lower hemisphere ray, i.e.,  $\Theta_c = 180^\circ - 130^\circ = 50^\circ$  and  $AZMc = 180^\circ + 40^\circ = 220^\circ$  while the values for ESK can be taken unchanged from the HYPO71 output file.

The P-wave polarities are commonly plotted according to the take-off angles and azimuths of their rays on either the Wulff or the Lambert-Schmidt net (Figures 3a and b; Aki and Richards, 1980, Vol. 1, p. 109-110). The latter yields less cluttered data plots when  $\Theta < 45^\circ$ , but in principle the fault planes are constructed in the same manner (see EX 3.3).



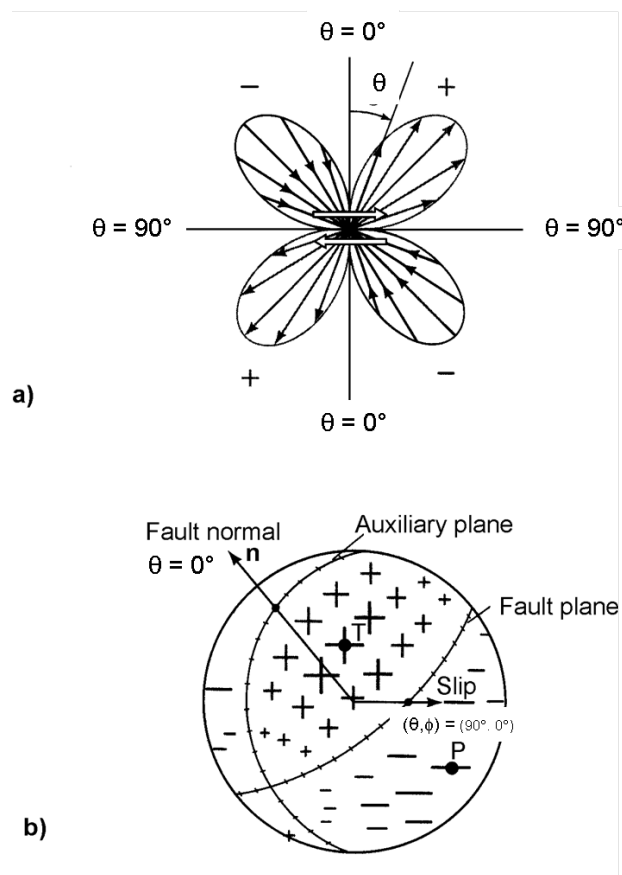
**Figure 3** a) the equal-angle Wulff net and b) the equal-area Lambert-Schmidt net.

The following relationships hold between the radius  $r$  (in cm) of the respective nets and the distance  $d$  (in cm) from the net center at which the take-off angle  $\Theta$  has to be mapped in the direction of the azimuth AZM:

Wulff net:  $d = r \cdot \tan(\Theta/2);$

Lambert-Schmidt net:  $d = r \cdot \sin(\Theta/2).$

**Note:** In both projections only the equator and the meridians **are great circles**. **Also:** All cut traces of the (possible) fault plane(s) through the focal sphere, which separate the segments of opposite polarity observations, have to be great circles. The amplitude radiation pattern of point source foci is commonly assumed to be spherically symmetric (Figure 4). For deviations see Chapter 3.



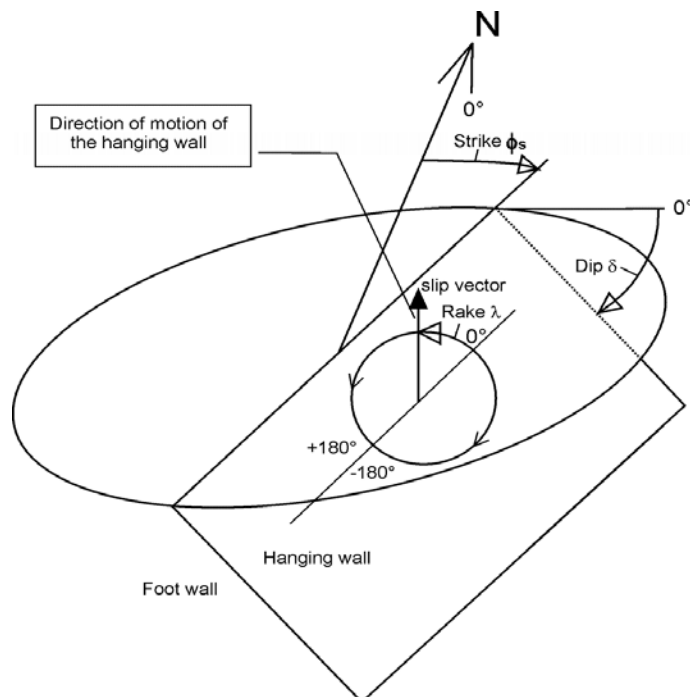
**Figure 4** Radiation pattern of the radial displacement component (P wave) due to a double-couple source: a) for a plane of constant azimuth (with lobe amplitudes proportional to  $\sin 2\theta$ ) and b) over a sphere centered on the origin. Plus and minus signs of various sizes denote amplitude variation (with the spherical coordinates  $\theta$  and  $\phi$ ) of outward (+; extensional quadrant) and inward (-; compressional quadrant) directed motions. The fault plane and auxiliary plane are nodal lines on which  $\cos \phi \sin 2\theta = 0$ . The pair of open arrows in a) at the center denotes the shear dislocation and **P** and **T** in b) mark the penetration points of the pressure and tension axes, respectively, through the focal sphere. Note the alternating quadrants of inward (D, -) and outward directed motions (U, +); (modified from Aki and Richards 1980; with kind permission of the authors).



According to the above, basically three steps are required for obtaining a fault-plane solution:

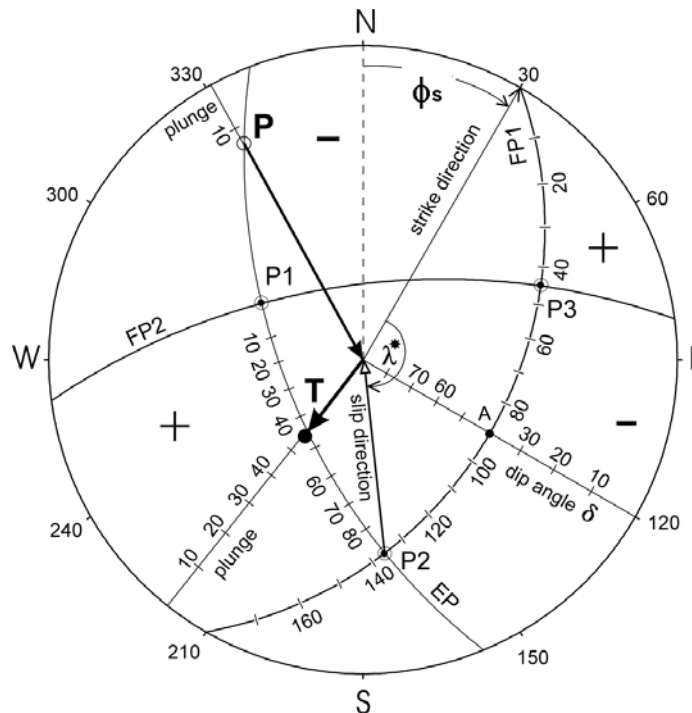
- (1) Calculating the positions of the penetration points of the seismic rays through the focal sphere. They are defined by the azimuth AZM and the take-off angle  $\Theta$  at which the ray leaves the source.
- (2) Marking these penetration points through the upper or lower hemisphere in a horizontal projection of that sphere by using different symbols for upward U (+) and downward D (–) directed first motions of P waves in vertical component records. Usually, lower hemisphere projections are used. Rays which have left the upper hemisphere have to be transformed into their equivalent lower hemisphere rays. This is possible because of spherical symmetry of the radiation pattern (see Figure 4).
- (3) Partitioning the projection of the lower focal sphere by two perpendicular great circles which separate all (or at least most) of the U (+) and D (–) arrivals in different quadrants.

Figure 5 shows the angles which describe the orientation and motion in space of a fault plane. The fault **strike angle**  $\phi_s$  is measured clockwise against North ( $0^\circ \leq \phi_s \leq 360^\circ$ ). To resolve the  $180^\circ$  ambiguity, it is assumed that when standing in the center of the net and looking into the strike direction of the fault it dips to the right hand side (i.e., its fault-trace projection is towards the right of the net center). The **dip angle**  $\delta$  describes the inclination of the hanging wall against the horizontal ( $0^\circ \leq \delta \leq 90^\circ$ ). The **rake angle**  $\lambda$  describes the displacement of the hanging wall relative to the foot wall ( $-180^\circ \leq \lambda \leq 180^\circ$ ).  $\lambda = 0$  corresponds to slip in strike direction,  $\lambda > 0$  means upward motion of the hanging wall (i.e., *reverse or thrust faulting component*) and  $\lambda < 0$  downward motion (i.e., *normal faulting component*). In other words: If the net center is located in the tension quadrant (+) or in the pressure quadrant (–) the fault has a thrust component or a normal component, respectively. The normal or thrust character of the fault-plane solution does not depend on which of the two nodal planes is the active fault plane.



**Figure 5** Angles describing the orientation and motion of faults (see text).

Figure 6 illustrates the measurement of these angles and of additional fault parameters in the net projections. P1, P2 and P3 mark the positions of the poles of the planes FP1 (fault plane), FP2 (auxiliary plane) and EP (equatorial plane) in their net projections. All three planes are perpendicular to each other (i.e.,  $90^\circ$  apart) and intersect in the poles of the respective third plane, i.e., FP1 and FP2 in P3, FP1 and EP in P2 etc.



**Figure 6** Determination of the fault plane parameters  $\phi_s$ ,  $\delta$  and  $\lambda$  in the net diagrams. The polarity distribution, slip direction and projection of FP1 shown here correspond to the faulting case depicted qualitatively in Figure 4. **Note:**  $\lambda^* = 180^\circ - \lambda$  when the center of the net lies in the tension (+) quadrant (i.e., event with thrust component) or  $\lambda^* = -\lambda$  when the center of the net lies in the pressure quadrant (i.e., event with normal faulting component). P1, P2 and P3 are the poles (i.e.,  $90^\circ$  off) of FP1, FP2 and EP, respectively. **P** and **T** are the penetration points (poles) of the pressure and tension axes, respectively, through the focal sphere. + and – signs mark the quadrants with upward and downward P-wave first motions in vertical component records, respectively.

**Note:** On the basis of polarity readings alone it can not be decided whether FP1 or FP2 was the active fault. Discrimination from seismological data alone is still possible but requires additional study of the *directivity effects* such as azimuthal variation of frequency (*Doppler effect*), amplitude ratios and/or waveforms (see Chapter 3). For sufficiently large shocks these effects can more easily be studied in low-frequency teleseismic recordings while in the local distance range, high-frequency waveforms and amplitudes may be strongly influenced by resonance effects due to low-velocity near-surface layers. Seismotectonic considerations or field evidence from surface rupture in case of strong shallow earthquakes as well as aftershock distribution may allow to resolve this ambiguity, too. For “cartoons” of several basic types of earthquake faulting and their related fault-plane solutions in so-called “beach-ball” presentations of the net projections see Chapter 3.

### 3 Tasks

#### Task 1:

If in Table 1  $\Theta > 90^\circ$ , then correct the take-off angles and azimuths in Table 1 for lower hemisphere projection:  $\Theta_c = 180^\circ - \Theta$ ,  $AZMc = AZM(<180^\circ) + 180^\circ$  or  $AZM(\geq 180^\circ) - 180^\circ$ . In case of  $\Theta < 90^\circ$  the original values remain unchanged.

#### Task 2:

Reproduce the Wulff net or the Lambert-Schmidt net projection in Figure 3, best on A4 paper with a radius  $r = 8$  cm. Place a tracing paper or a transparency sheet over the net. Mark on it the center and perimeter of the net as well as the N, E, S and W directions. Pin the marked center of the transparency sheet with a needle to the center of the net copy on paper.

#### Task 3:

Mark the azimuth of the station on the perimeter of the transparency and rotate the latter until the tick mark is aligned along an azimuth of  $0^\circ$ ,  $90^\circ$ ,  $180^\circ$  or  $270^\circ$ . Measure the take-off angle  $\Theta$  from the center of the net along this azimuth. This gives the intersection point of the particular P-wave ray with the lower hemisphere. Mark on this position the P-wave polarity with a **distinct** + for compression or o for dilatation (U or D in Table 1) using *different colors* for better distinction of closely spaced polarities of different sign. **Note:** For calculating the proper distance  $d$  of the polarity entry from the center of the respective net you may also use the respective formulas given above for the Wulff net and the Lambert-Schmidt net as a function of  $r$  and  $\Theta$ . In case that rays left the source through the upper hemisphere ( $\Theta > 90^\circ$ )  $\Theta_c$  and  $AZMc$  for lower hemisphere projection have to be plotted.

#### Task 4:

Rotate the transparent sheet with the plotted data over the net and try to find a **great circle** (**meridian** on the net copy) which separates as good as possible the expected quadrants with different first motion signs. This great circle represents the intersection trace of **one of the possible fault** (or nodal) **planes** (FP1) with the lower half of the focal sphere.

**Note:** Inconsistent polarities that are close to each other may be due to uncertainty in reading correctly the first motion direction of P-wave arrivals with low signal-to-noise ratio or to an incorrect velocity model for calculating the take-off angle. Yet, small P-wave amplitudes are to be expected for take-off angles near to the nodal (fault) planes. Thus, clusters of inconsistent polarities may guide your “eye fit” in finding the best separating great circle. Also computer programs for finding best fitting fault-plane solutions try to minimize both the number of misfitting polarities in the quadrants as well as their distance from the nodal plane. However, be aware that isolated inconsistent polarities may also be due to false polarity switching of the seismometer output or simply by reading errors on the seismic record.

#### Task 5:

Mark point A at the middle of the FP1 projection and find, perpendicular to FP1 and  $90^\circ$  apart, the pole P1 of FP1 (see Figure 6). All great circles, passing this pole are perpendicular to the FP1. Since the second possible fault (auxiliary) plane (FP2) must be perpendicular to the FP1, it has to pass P1. Find, accordingly, a great circle FP2 which again has to separate areas of different polarity.

**Task 6:**

Find the pole P2 for FP2 (which lies on FP1) and delineate the equatorial plane EP. The latter is perpendicular to both FP1 and FP2, i.e., it is a great circle through the poles P1 and P2. The intersection point of FP1 and FP2, P3, is the pole of the equatorial plane.

**Task 7:**

Mark the position of the poles of the pressure (**P**) and tension axes (**T**) on the equatorial plane and determine the direction of these axes towards (for **P**) and away from the center (for **T**) of the used net (see Figure 6). The poles for **P** and **T** lie on the equatorial plane in the center of the respective quadrants of downward (– or **o**) and upward (+) P-wave first motions, i.e., 45° away from the intersection points of the two fault planes with the equatorial plane. **Note:**

**All angles in the net projections have to be measured along great circles!**

**Task 8:**

Mark the slip vectors, connecting the intersection points of the fault planes with the equatorial plane, with the center of the considered net. If the center lies in a tension quadrant, then the slip vector points either from P2 on FP1 or from P1 on FP2 to the net center (see Figure 6). If it lies in a pressure quadrant, then the slip vectors points from the center in the opposite directions. The **slip vector** shows the direction of displacement of the **hanging wall** (see Figure 5).

**Task 9:**

Determine the azimuth (fault strike direction  $\phi_s$ ) of both FP1 and FP2. It is the angle measured clockwise against North between the directional vector connecting the center of the net with the end point of the respective projected fault trace lying towards the right of the net center (i.e., with the fault plane dipping towards the right; see Figure 6).

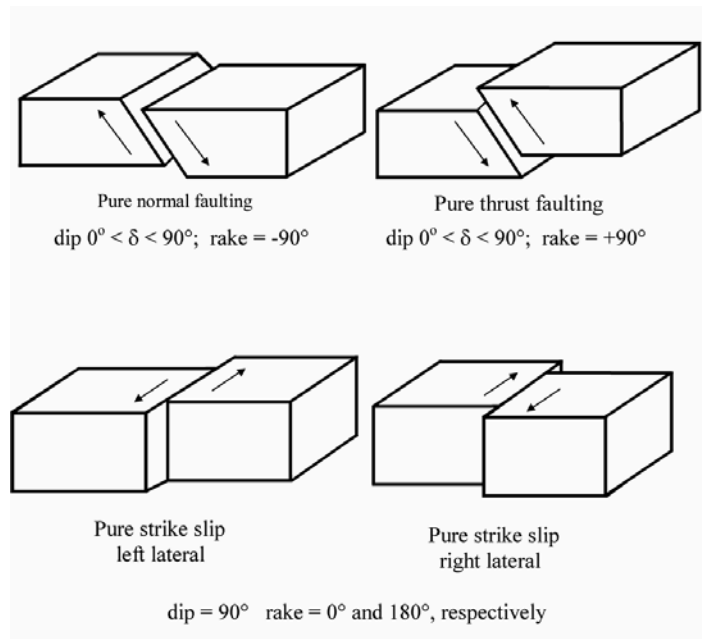
**Task 10:**

Determine the dip angle  $\delta$  (measured from horizontal) for both FP1 and FP2 by putting their projected traces on a great circle. Measure  $\delta$  as the difference angle from the outermost great circle (perimeter of the net) towards the considered fault-plane trace.

**Task 11:**

Determine the slip direction (i.e., the sense of motion along the two possible fault planes. It is obtained by drawing one vector each from the center of the net to the poles P1 and P2 of the nodal planes (or vice versa from the poles to the center, depending on the sign of the rake angle  $\lambda$ ). The vector from (or to) the center towards (or away from) P1 (P2) shows the slip direction along either FP2 or FP1. The **rake angle**  $\lambda$  is **negative** when it lies in the pressure (–) quadrant (event **with a normal faulting component**) and **positive** in case the center of the net lies in the tension (+) quadrant (i.e., an **event with a thrust component**) (Figure 7 top). In the first case holds  $\lambda = -\lambda^*$  and in the second case  $\lambda = 180^\circ - \lambda^*$ .  $\lambda^*$  has to be measured on the great circle of the respective fault plane between its crossing point with the equatorial plain and the respective azimuth direction of the considered fault plane (see Figure 6). For a pure strike slip motion ( $\delta = 90^\circ$ )  $\lambda = 0$  defines a left lateral strike-slip and  $\lambda = 180^\circ$  defines a right-

lateral strike-slip which can also be interpreted as a left-lateral (counter-clockwise) or right-lateral (clockwise) “rotational” movement of the two blocks against each other (Figure 7 below). **Note:** If the slip on the first fault plane is left-lateral the slip on the second one is right-lateral and vice versa.



**Figure 7** Basic types of earthquake faulting for some selected dip and rake angles. Mixed types of faulting occur when  $\lambda \neq 0, 180^\circ$  or  $\pm 90^\circ$ , e.g., normal faulting with strike-slip component or strike-slip with thrust component. Also, dip angles may vary between  $0^\circ < \delta \leq 90^\circ$ . For fault-plane traces and polarity distributions of mixed faulting types in their "beach-ball presentation" see also Chapter 3, section Fault-plane solutions.

#### Task 12:

The azimuth of the pressure and the tension axes, respectively, is equal to the azimuth of the line connecting the center of the net through the poles of **P** and **T** with the perimeter of the net. Their plunge is the dip angle of these vectors against the horizontal (to be measured as for  $\delta$ ) (see Figure 6).

#### Task 13:

Insert your results of estimated fault-plane parameters as well as pressure and tension axes for the Erzincan aftershock into Table 2.

**Table 2**

|                      | strike | dip | rake |
|----------------------|--------|-----|------|
| <b>Fault plane 1</b> |        |     |      |
| <b>Fault plane 2</b> |        |     |      |

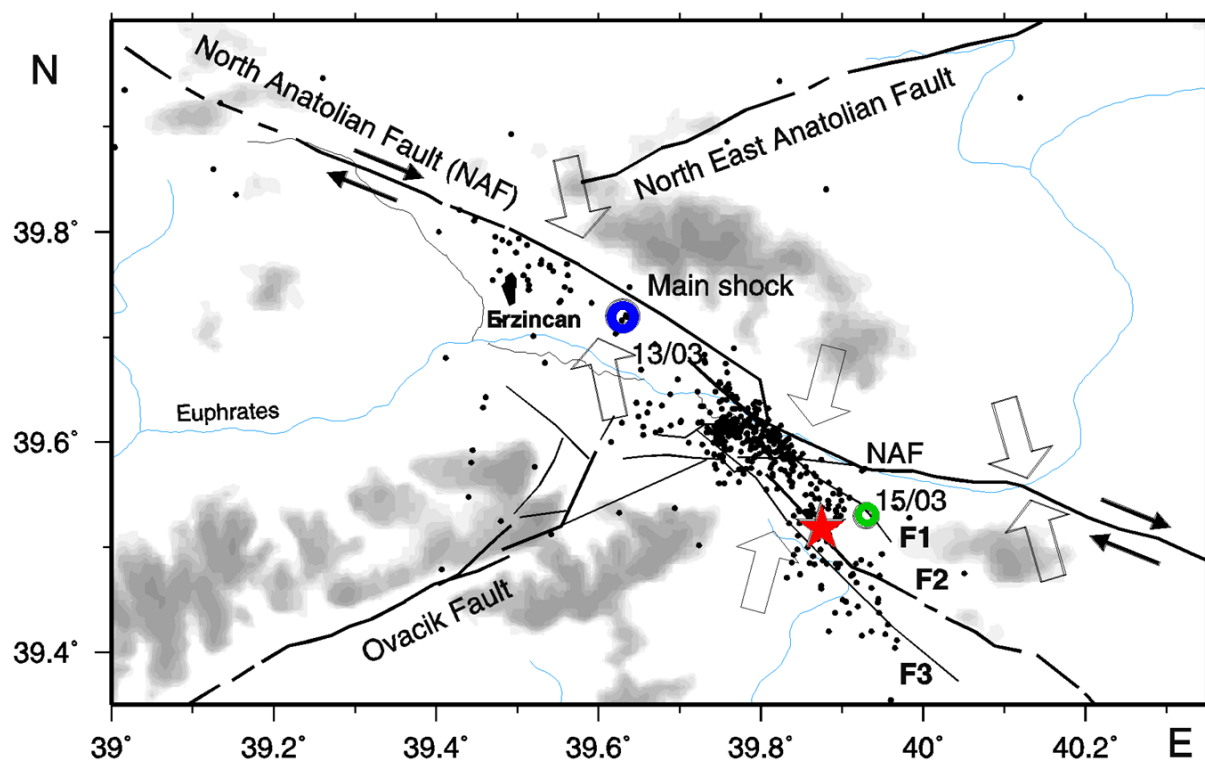
|               | azimuth | plunge |
|---------------|---------|--------|
| Pressure axis |         |        |
| Tension axis  |         |        |

**Note:** The angles may range between:

$$\begin{aligned}
 0^\circ &< \text{strike} < 360^\circ \\
 0^\circ &< \text{azimuth} < 360^\circ \\
 0^\circ &< \text{dip} < 90^\circ \\
 0^\circ &< \text{plunge} < 90^\circ \\
 -180^\circ &< \text{rake} < 180^\circ
 \end{aligned}$$

#### Task 14:

The question which of the two nodal planes was the active fault plane, and hence the other the auxiliary plane, cannot be answered on the basis of the fault-plane solution alone. Considering the event in its seismotectonic context may give an answer. Therefore, we have marked the epicenter of the event in Figure 8 with an open star at the secondary fault F2 to the North Anatolian Fault (NAF).



**Figure 7** Epicenters of aftershocks between March 21 and June 16, 1992 of the March 13, 1992 Erzincan earthquake, Turkey. The open circles represent the main shock (blue) and its strongest aftershock on March 15 (green), and the red star the analyzed aftershock. **F1, F2 and F3** are secondary faults to the North Anatolian Fault (NAF). **Black arrows** - directions of relative plate motion, **open arrows** - direction of maximum horizontal compression as derived from centroid moment-tensor solutions of stronger earthquakes (modified Figure from Grosser et al., 1998).

- Decide which was the likely fault plane (FP1 or FP2)?
- What was the type of faulting?
- What was the direction of slip? and
- Is your solution compatible with the general sense of plate motion in the area as well with the orientation of the acting fault and the orientation of stress/deformation in the area?

## 4 Solutions and discussion

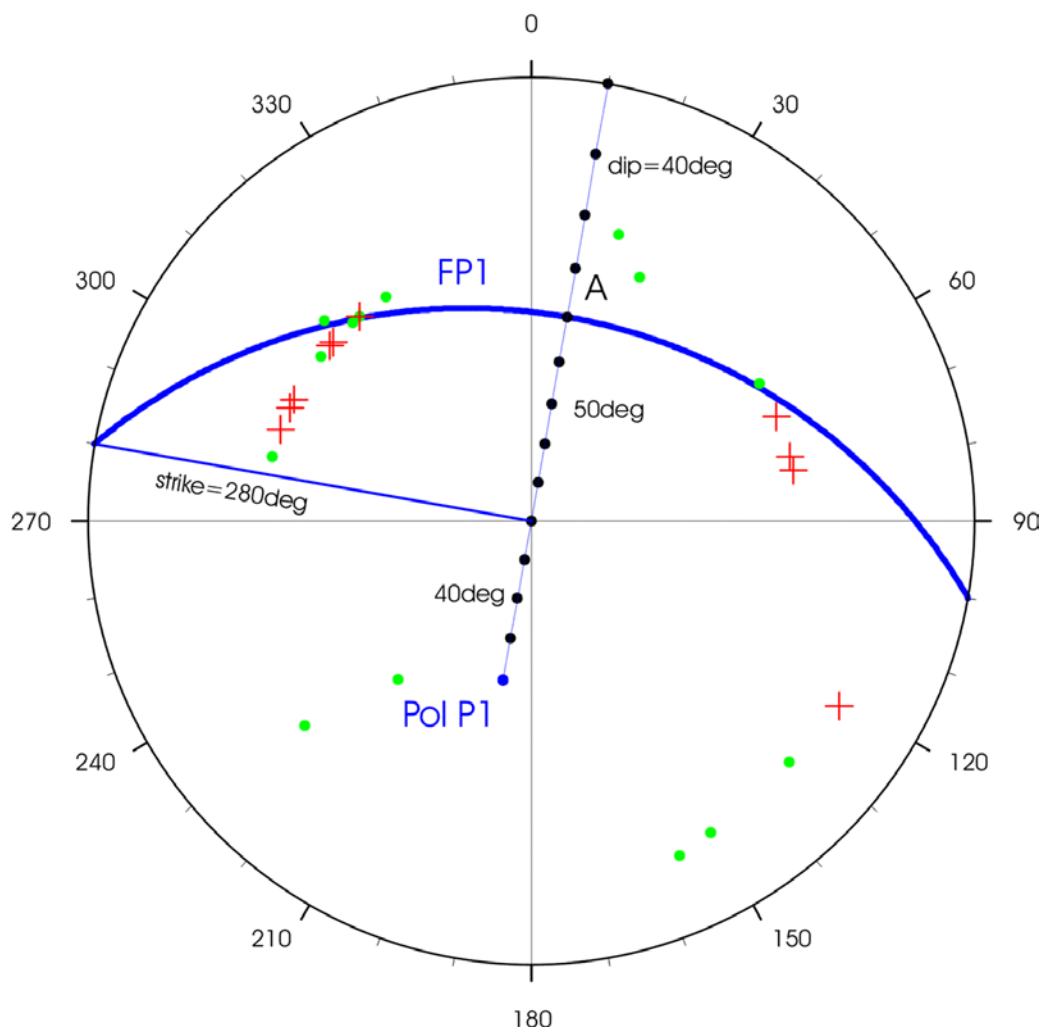
Taking it that you have carried out the exercise step-by-step yourself according to the above data, tasks and explanations, you may now compare your results, again step-by-step, with those which the authors of this exercise derived themselves by hand and or plotted them by using a computer program. The PC derived values are given in the summary parameter table in brackets. Fault planes determined by eye-fit to the polarity data may be uncertain by about  $\pm 10^\circ$ . This is acceptable. Even computer-assisted best fits to the data will produce different acceptable solutions within about the same error range with only slightly different standard deviations (e.g., Figure 1 in EX 3.3, comparing NEIC and HRVD solutions, respectively). But poor distribution of seismological stations (resulting in insufficient polarity data in several quadrants of the net diagram), erroneous polarity readings and/or differences in the velocity models used may result in still larger deviations between calculated and actual fault planes. One should also be aware that the assumed constant angular ( $45^\circ$ ) relationship between the fault plane on the one hand and the pressure and tension axis on the other hand is true in fact only in the case of a fresh rupture in a homogeneous isotropic medium. It may not be correct in the stress environment of real tectonic situations and for near surface sources in particular (i.e.,  $P$  and  $T \neq \sigma_1$  and  $-\sigma_3$ , respectively; see discussion in section 3.1.2.6 of Chapter 3). Only the parameters of the fault plane are relatively reliable, provided that this applies also to the used velocity model. If, however, your polarity patterns and/or manually determined parameter values differ by more than about  $20^\circ$  from the solutions presented below or yield even a different type of faulting mechanism then you should critically check and correct your data entries and/or fault-plane fits.

### Solution 1, Task 1:

| STA | AZM<br>(degree) | $\Theta$<br>(degree) | POL | AZMc<br>(degree) | $\Theta_c$<br>(degree) |
|-----|-----------------|----------------------|-----|------------------|------------------------|
| ALI | 40              | 130                  | D   | 220              | 50                     |
| ME2 | 134             | 114                  | D   | 314              | 66                     |
| KAN | 197             | 112                  | D   | 17               | 68                     |
| YAR | 48              | 111                  | D   | 228              | 69                     |
| ERD | 313             | 103                  | D   | 133              | 77                     |
| DEM | 330             | 102                  | D   | 150              | 78                     |
| GIR | 301             | 102                  | U   | 121              | 78                     |
| UNK | 336             | 101                  | D   | 156              | 79                     |
| SAN | 76              | 62                   | U   |                  |                        |
| PEL | 327             | 62                   | D   |                  |                        |
| GUN | 290             | 62                   | U   |                  |                        |
| ESK | 312             | 62                   | U   |                  |                        |

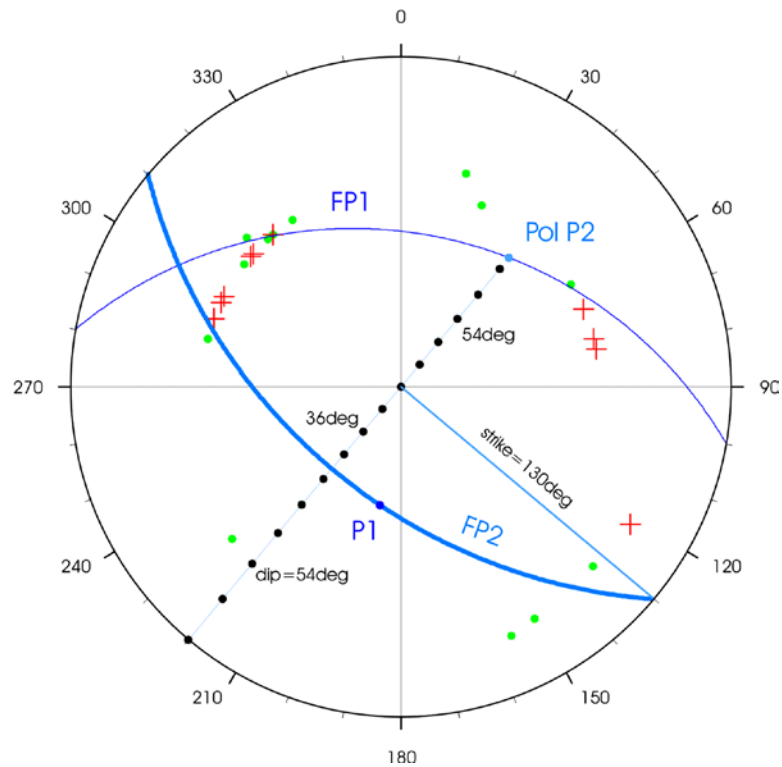
|     |     |    |   |  |  |
|-----|-----|----|---|--|--|
| SOT | 318 | 62 | D |  |  |
| BA2 | 79  | 62 | U |  |  |
| MOL | 297 | 62 | U |  |  |
| YUL | 67  | 62 | U |  |  |
| ALT | 59  | 62 | D |  |  |
| GUM | 320 | 62 | U |  |  |
| GU2 | 320 | 62 | D |  |  |
| BAS | 308 | 62 | D |  |  |
| BIN | 295 | 62 | U |  |  |
| HAR | 24  | 62 | D |  |  |
| KIZ | 311 | 62 | U |  |  |
| AKS | 284 | 62 | D |  |  |
| SUT | 295 | 62 | U |  |  |

**Solutions 2, Tasks 3-5, 9 and 10: Plotting polarities, finding FP1 and determining its strike and dip.** Note that FP1 goes through a cluster of nearby conflicting polarities but separates well most of the + and - (•) signs.

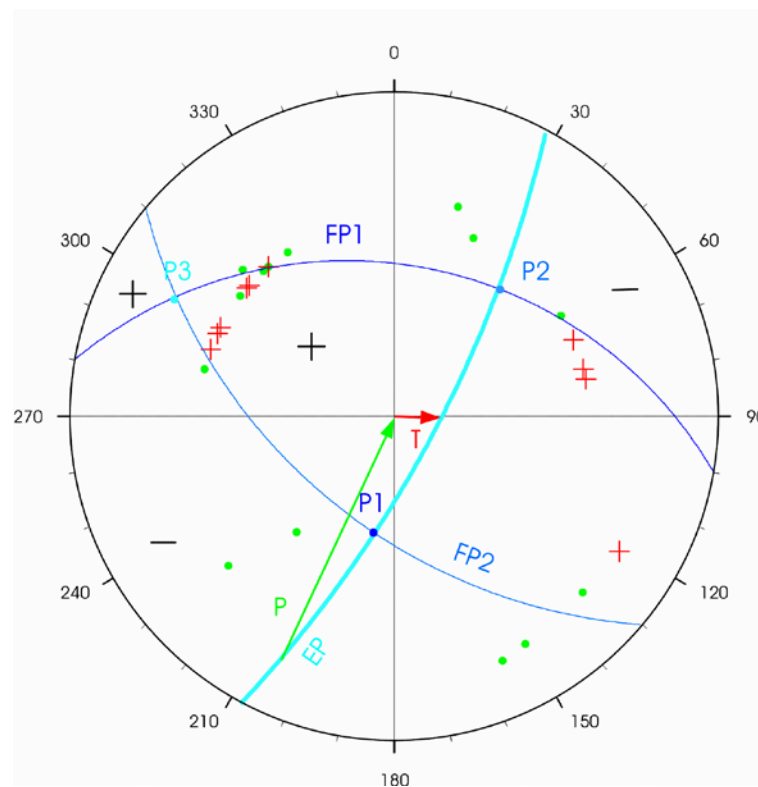




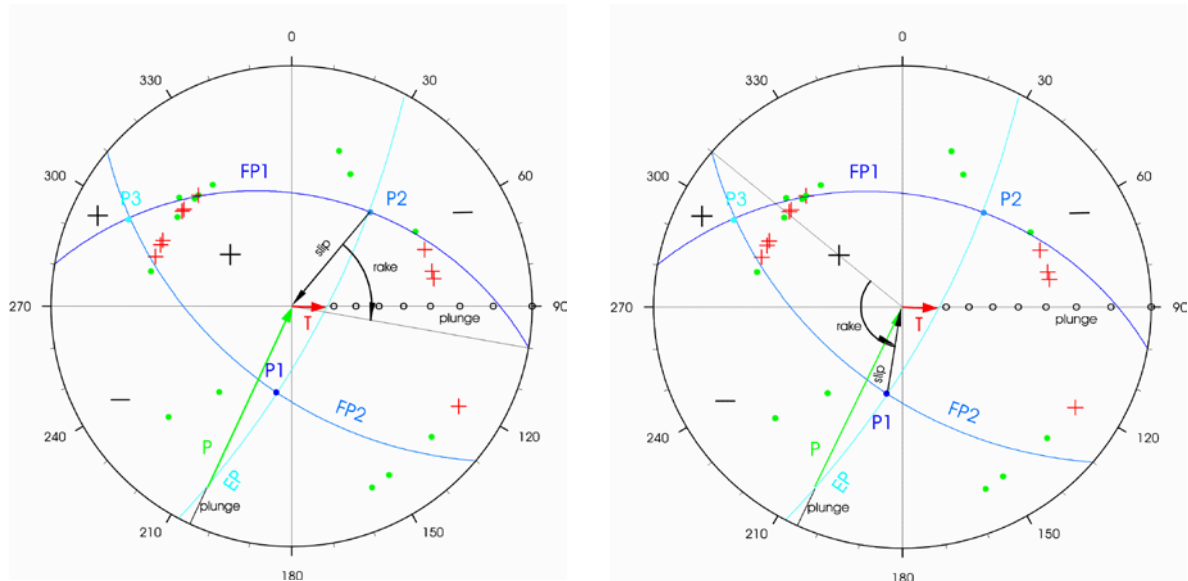
**Solution 3, Task 6, 9 and 10: Finding FP2 and determining its strike and dip.**



**Solution 3, Task 7: Position of the poles on the equatorial plane EP of the pressure axis P (green) and tension axis T (red):**



**Solutions 4, Tasks 8, 11 and 12: Drawing the slip vectors of FP1 and FP2, determining the rake angle and measuring the azimuth and plunge of the P and T axes.**



### Summary solutions of all tasks:

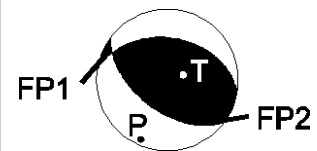
Table 3 summarizes both the angles measured by hand as well as the ones (in brackets) that result from best fitting computer analysis of the polarity plots. The agreement is astonishingly good.

**Table 3**

|                            | strike        | dip         | rake          |
|----------------------------|---------------|-------------|---------------|
| <b>Fault plane1 (FP1)</b>  | 280° (278.5°) | 40° (39.9°) | 68° (67.4°)   |
| <b>Fault plane 2 (FP2)</b> | 130° (127.0°) | 54° (53.7°) | 108° (107.8°) |

|                      | azimuth       | plunge      |
|----------------------|---------------|-------------|
| <b>Pressure axis</b> | 205° (204.4°) | 7° (7.1°)   |
| <b>Tension axis</b>  | 90° (88.6°)   | 73° (74.0°) |



### Solution 5, Task 14:

- FP2 was more likely the active fault.
- The aftershock was a thrust event with a very small right-lateral strike-slip component.

- c) The slip direction is here strike - rake azimuth, i.e., for FP2  $130^\circ - 108^\circ = 12^\circ$  from north. This is close to the direction of maximum horizontal compression ( $15^\circ$ ) in the nearby area as derived from centroid moment-tensor solutions of stronger events.
- d) The strike of FP2 for this event agrees quite well with the strike direction of mapped local faults F1, F2, and F3 in the horse tail formed aftershock cloud southeast of the Erzincan basin. The strongest aftershock ( $M_s = 5.8$ ) on March 15, 1992 (small double circles) took place in the horse tail part, too. Therefore, it is probable that FP2 was the acting fault. The North Anatolian Fault (NAF) which reflects the general tectonics confines the northern margin of the aftershock zone and it is continuing towards the triple point of the NAF with the East Anatolian Fault.

## Reference

Grosser, H., Baumbach, M., Berckhemer, H., Baier, B., Karahan, A., Schelle, H., Krüger, F., Paulat, A., Michel, G., Demirtas, R., Gencoglu, S., and Yilmaz, R. (1998). The Erzincan (Turkey) Earthquake ( $M_s$  6.8) of March 13, 1992 and its Aftershock Sequence. *Pure appl. Geophys.*, **152**, 465–505.

## Acknowledgment

The authors are grateful to Dr. Helmut Grosser for his careful review and useful comments which helped to improve the manuscript and for providing us with a figure of Grosser et al. (1998) to be amended for the purpose of this exercise.

| Topic   | Take-off angle calculations for fault-plane solutions and reconstruction of nodal planes from the parameters of fault-plane solutions                      |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version | September 1999; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_EX_3.3">10.2312/GFZ.NMSOP-2_EX_3.3</a>                                                   |

## 1 Aim

The exercise aims at making familiar with the calculation of the take-off angles AIN of seismic P-wave rays leaving the seismic source towards the seismic station. These angles are required for determining fault-plane solutions (FPS) from first-motion polarity readings (see EX 3.2). Take-off angles depend on the velocity model of the Earth, the source depth  $h$  and the epicentral distance  $\Delta$ , at which the considered rays arrive at the Earth's surface. The AIN calculated in this exercise for a given event and a number of seismic stations at different  $\Delta$  will then be checked whether they are consistent with the reported polarity readings and FPS calculated for this event by international agencies. For this you will reconstruct on a Lambert-Schmidt net projection the fault-plane traces from the reported nodal-plane parameters.

## 2 Data, models and procedure

When localizing near events by using HYPO71 or similar programs the values for both the azimuth (AZM) and for the take-off angles (AIN) of the rays leaving the source towards the considered stations are given in the localization output file. One can use them, together with the first motion polarity readings, straight forward for the determination of fault-plane solutions (see EX 3.2). When one intends to determine the fault-plane solution for seismic events published in the bulletins of the International Seismological Centre (ISC) one finds therein, besides data for polarity readings from the reporting stations ( $\uparrow$  or  $c$  for up and  $\downarrow$  or  $d$  for down in short- or long-period instruments, respectively), only values for the azimuth (AZM) but not for the respective take-off angle (AIN). Figure 1 shows a typical portion of event-stations report from the ISC. Its header also gives the seismic moment tensor and fault-plane solutions calculated by various international data centers or agencies using different (sometimes automated) procedures. Values for AIN can be calculated by using the relationship

$$\sin \text{AIN} = (180/\pi) \times (v_p/r_h) \times p(\Delta, h). \quad (1)$$

$v_p(h)$  is the P-wave velocity at the depth  $h$  (in km/s),  $r_o = 6371$  km is the Earth's radius and  $r_h = r_o - h$ .  $p(\Delta, h) = dT/d\Delta$  is the ray parameter; it corresponds to the gradient of the travel-time curve at the point of observation on the Earth's surface (both in units s/deg) at the epicentral distance  $\Delta$  (in degree) (see Fig. 2.27) and is a function of the hypocentral depth  $h$  (in km). The value of the ray parameter is identical with that of the horizontal component of the of the slowness vector. Tables 1 and 2 give the respective values  $v_p(h)$  and  $p(\Delta, h)$  for P waves.

**Table 1**  $v_p(h)$  according to the IASPEI91 velocity model (Kennett, 1991).

| h (km) | $v_p$ (km/s) | h (km) | $v_p$ (km/s) | h (km) | $v_p$ (km/s) |
|--------|--------------|--------|--------------|--------|--------------|
| 0      | 5.8000       | 120    | 8.0500       | 471    | 9.5650       |
| 20     | 5.8000       | 171    | 8.1917       | 571    | 9.9010       |
| 20     | 6.5000       | 210    | 8.3000       | 660    | 10.2000      |
| 35     | 6.5000       | 271    | 8.5227       | 660    | 10.7900      |
| 35     | 8.0400       | 371    | 8.8877       | 671    | 10.8192      |
| 71     | 8.0442       | 410    | 9.0300       | 760    | 11.0558      |
| 120    | 8.0500       | 410    | 9.3600       |        |              |

**Table 2** Ray parameter  $p = dT/d\Delta$  (= horizontal slowness component) of Pn, P and PKP<sub>df</sub> first arrivals at the Earth's surface as a function of hypocentral depth  $h$  according to IASPEI 1991 Seismological Tables (Kennett, 1991)

|              |                                     | <b>p (in s/deg)</b> |                   |                   |                   |
|--------------|-------------------------------------|---------------------|-------------------|-------------------|-------------------|
| <b>Phase</b> | <b><math>\Delta</math> (in deg)</b> | <b>h = 0 km</b>     | <b>h = 100 km</b> | <b>h = 300 km</b> | <b>h = 600 km</b> |
| Pn (P)       | 2                                   | 13.75               | 12.90             | 7.91              | 4.01              |
|              | 4                                   | 13.75               | 13.49             | 10.96             | 6.91              |
|              | 6                                   | 13.74               | 13.58             | 11.95             | 8.60              |
|              | 8                                   | 13.72               | 13.60             | 12.25             | 9.48              |
|              | 10                                  | 13.70               | 13.59             | 12.26             | 9.90              |
|              | 12                                  | 13.67               | 13.29             | 12.12             | 10.05             |
|              | 14                                  | 13.64               | 12.91             | 11.03             | 10.06             |
|              | 16                                  | 12.92               | 12.43             | 10.91             | 9.17              |
|              | 18                                  | 12.33               | 10.97             | 10.73             | 9.10              |
|              | 20                                  | 10.90               | 10.81             | 10.50             | 9.02              |
| P            | 22                                  | 10.70               | 10.58             | 9.12              | 8.90              |
|              | 24                                  | 9.14                | 9.11              | 9.03              | 8.83              |
|              | 26                                  | 9.06                | 9.02              | 8.91              | 8.76              |
|              | 28                                  | 8.93                | 8.90              | 8.83              | 8.66              |
|              | 30                                  | 8.85                | 8.82              | 8.75              | 8.56              |
|              | 32                                  | 8.77                | 8.74              | 8.65              | 8.45              |
|              | 34                                  | 8.67                | 8.64              | 8.54              | 8.33              |
|              | 36                                  | 8.56                | 8.52              | 8.42              | 8.21              |
|              | 38                                  | 8.44                | 8.40              | 8.29              | 8.08              |
|              | 40                                  | 8.30                | 8.26              | 8.16              | 7.95              |
|              | 42                                  | 8.17                | 8.13              | 8.03              | 7.82              |
|              | 44                                  | 8.03                | 7.99              | 7.89              | 7.69              |
|              | 46                                  | 7.89                | 7.85              | 7.75              | 7.56              |
|              | 48                                  | 7.55                | 7.71              | 7.61              | 7.42              |
|              | 50                                  | 7.60                | 7.56              | 7.47              | 7.29              |
|              | 52                                  | 7.46                | 7.42              | 7.33              | 7.15              |
|              | 54                                  | 7.31                | 7.28              | 7.19              | 7.02              |
|              | 56                                  | 7.17                | 7.13              | 7.05              | 6.88              |
|              | 58                                  | 7.02                | 6.99              | 6.90              | 6.74              |
|              | 60                                  | 6.88                | 6.84              | 6.76              | 6.61              |
|              | 62                                  | 6.73                | 6.70              | 6.62              | 6.47              |

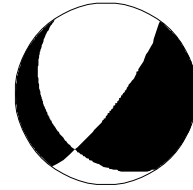
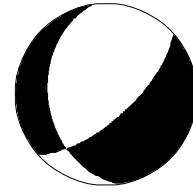
Table 2: cont.

|                   |                   | p (in s/deg) |            |            |            |
|-------------------|-------------------|--------------|------------|------------|------------|
| Phase             | $\Delta$ (in deg) | h = 0 km     | h = 100 km | h = 300 km | h = 600 km |
| P                 | 64                | 6.59         | 6.55       | 6.48       | 6.33       |
|                   | 66                | 6.44         | 6.41       | 6.33       | 6.19       |
|                   | 68                | 6.30         | 6.27       | 6.19       | 6.05       |
|                   | 70                | 6.15         | 6.12       | 6.05       | 5.91       |
|                   | 72                | 6.00         | 5.97       | 5.90       | 5.77       |
|                   | 74                | 5.86         | 5.83       | 5.76       | 5.63       |
|                   | 76                | 5.71         | 5.68       | 5.61       | 5.49       |
|                   | 78                | 5.56         | 5.53       | 5.46       | 5.34       |
|                   | 80                | 5.40         | 5.38       | 5.31       | 5.20       |
|                   | 82                | 5.25         | 5.22       | 5.16       | 5.04       |
|                   | 84                | 5.09         | 5.07       | 5.01       | 4.90       |
|                   | 86                | 4.94         | 4.92       | 4.85       | 4.72       |
|                   | 88                | 4.74         | 4.72       | 4.69       | 4.65       |
|                   | 90                | 4.66         | 4.65       | 4.64       | 4.61       |
|                   | 92                | 4.61         | 4.61       | 4.60       | 4.57       |
|                   | 94                | 4.58         | 4.57       | 4.55       | 4.51       |
|                   | 96                | 4.52         | 4.51       | 4.49       | 4.44       |
|                   | 98                | 4.45         | 4.44       | 4.44       | 4.44       |
| P <sub>diff</sub> | 100-144           | 4.44         | 4.44       | 4.44       | 4.44       |
| PKP <sub>df</sub> | 114               | 1.92         | 1.92       | 1.92       | 1.92       |
|                   | 116-122           | 1.91         | 1.91       | 1.91       | 1.91       |
|                   | 124-126           | 1.90         | 1.90       | 1.90       | 1.90       |
|                   | 130               | 1.88         | 1.88       | 1.88       | 1.88       |
|                   | 136               | 1.84         | 1.84       | 1.84       | 1.83       |
|                   | 140               | 1.80         | 1.79       | 1.79       | 1.78       |
|                   | 142               | 1.76         | 1.76       | 1.76       | 1.75       |
|                   | 144               | 1.73         | (1.72)     | (1.72)     | (1.71)     |
|                   | 146               | 1.68         | 1.68       | 1.67       | 1.66       |
|                   | 148               | 1.63         | 1.62       | 1.62       | 1.60       |
|                   | 150               | 1.57         | 1.56       | 1.55       | 1.54       |
|                   | 152               | 1.49         | 1.49       | 1.48       | 1.47       |
|                   | 154               | 1.42         | 1.41       | 1.40       | 1.39       |
|                   | 156               | 1.33         | 1.33       | 1.32       | 1.30       |
|                   | 158               | 1.24         | 1.23       | 1.23       | 1.21       |
|                   | 160               | 1.14         | 1.14       | 1.13       | 1.11       |
|                   | 162               | 1.04         | 1.03       | 1.03       | 1.01       |
|                   | 164               | 0.93         | 0.93       | 0.92       | 0.91       |
|                   | 166               | 0.82         | 0.82       | 0.81       | 0.80       |
|                   | 168               | 0.71         | 0.70       | 0.70       | 0.69       |
|                   | 170               | 0.59         | 0.59       | 0.58       | 0.58       |
|                   | 172               | 0.47         | 0.47       | 0.47       | 0.47       |
|                   | 174               | 0.36         | 0.36       | 0.35       | 0.35       |
|                   | 176               | 0.24         | 0.24       | 0.24       | 0.23       |
|                   | 178               | 0.12         | 0.12       | 0.12       | 0.12       |
|                   | 180               | 0.00         | 0.00       | 0.00       | 0.00       |

NEIC Moment-tensor solution:  $s_{23}$ , scale  $10^{17}$  Nm;  $M_{rr}$ -3.05;  
 $M_{\theta\theta}$ -0.97;  $M_{\phi\phi}$ 4.03;  $M_{r\theta}$ -2.51;  $M_{r\phi}$ -1.95;  $M_{\theta\phi}$ 2.71. Depth  
 272km; Principal axes: T 6.09, Plg17°, Azm117°; N -136,  
 Plg27°, Azm216°; P -4.73, Plg57°, Azm358°; Best double  
 couple:  $M_o$ 5.4x10<sup>17</sup>Nm; NP1:  $\phi_s$ 172°,  $\delta$ 36°,  $\lambda$ -140°. NP2:  
 $\phi_s$ 48°,  $\delta$ 68°,  $\lambda$ -60°.

HRVD 05<sup>d</sup> 13<sup>h</sup> 24<sup>m</sup> 15<sup>s</sup>.7±0<sup>s</sup>.2, 39°.10N±°.02x15°.39E±°.02,  
 $h$ 295<sup>km</sup>±.8<sup>km</sup>, Centroid moment-tensor solution. Data used:  
 GDSN; LP body waves: s50, c\*\*; Half duration: 1<sup>s</sup>.9.  
 Moment tensor: Scale 10<sup>17</sup>Nm;  $M_{rr}$ -2.17±.06;  
 $M_{\theta\theta}$ -1.97±.10;  $M_{\phi\phi}$ 4.14±.09;  $M_{r\theta}$ -3.51±.09;  $M_{r\phi}$ -3.29±.09;  
 $M_{\theta\phi}$ 0.01±.09. Principal Axes: T 5.83, Plg27°, Azm103°;  
 N 0.32, Plg30°, Azm210°; P -6.15, Plg48°, Azm339°. Best  
 Double couple:  $M_o$ 6.0x10<sup>17</sup>Nm, NP1:  $\phi_s$ 146°,  $\delta$ 33°,  $\lambda$ -157°.  
 NP2:  $\phi_s$ 37°,  $\delta$ 78°,  $\lambda$ -60°.

ISC 05<sup>d</sup>13<sup>h</sup>24<sup>m</sup>11<sup>s</sup>.4±0<sup>s</sup>.13, 39.16±0<sup>s</sup>.16x15°.18E±°.014,  
 $h$ 290<sup>km</sup>±1.3<sup>km</sup>, ( $h$ 286km±2.7<sup>km</sup>:pP-P), n757,  $\sigma$ 1<sup>s</sup>.04/729,  
 Mb5.7/107, 119C-155D, Southern Italy.



|                     |      |     |     |            |      |
|---------------------|------|-----|-----|------------|------|
| OVO Vesuviano       | 1.77 | 340 | ↑iP | 13 24 57.2 | +1.5 |
| MCT Mte Cammarata   | 1.95 | 219 | P   | 13 24 57.7 | +0.6 |
| FG4 Candela         | 1.99 | 8   | P   | 13 24 58.2 | +0.9 |
| MEU Monte Lauro     | 2.07 | 186 | dP  | 13 24 56.8 | -1.3 |
| PZI Palazzolo       | 2.14 | 186 | eP  | 13 24 57   | -1.7 |
| FAI Favara          | 2.21 | 213 | dP  | 13 24 59.5 | +0.1 |
| MSC Monte Massico   | 2.23 | 336 | ↑iP | 13 25 01.1 | +1.6 |
| SGG Gregorio Matese | 2.30 | 345 | ↑iP | 13 25 01.9 | +1.8 |

**Figure 1** Typical section of an ISC bulletin (left) with NEIC (National Earthquake Information Center) and Harvard University (HRVD) moment-tensor fault-plane solutions (right) for the Italy deep earthquake ( $h = 286$  km) of Jan. 05, 1994. Columns 3 to 5 of the bulletin give the following data: 3 - epicentral distance in degrees, 4 - azimuth AZM in degrees, 5 - phase code and polarity.

Table 3 gives respective selected data from the ISC bulletin for five seismic stations at different epicentral distances ( $\Delta$ ) and azimuth (AZM) for the Italy earthquake shown in Figure 1. The polarity readings correspond to the first P ( $\Delta < 100^\circ$ ) or PKP ( $\Delta > 110^\circ$ ) onsets.

**Table 3**

| STA | $\Delta$<br>(deg) | AZM<br>(deg) | POL | $v_P(h)$<br>km/s | $v_P/r_h$<br>(s <sup>-1</sup> ) | $p(\Delta, h)$<br>(s/deg) | AIN<br>(deg) | AZMc<br>(deg) | AINc<br>(deg) |
|-----|-------------------|--------------|-----|------------------|---------------------------------|---------------------------|--------------|---------------|---------------|
| SGG | 2.30              | 345          | +   |                  |                                 |                           |              |               |               |
| KHC | 10.03             | 354          | -   |                  |                                 |                           |              |               |               |
| BTH | 12.25             | 294          | +   |                  |                                 |                           |              |               |               |
| ZAK | 60.02             | 48           | -   |                  |                                 |                           |              |               |               |
| PAE | 154.8             | 324          | -   |                  |                                 |                           |              |               |               |

### 3 Tasks

#### Task 1:

For the data given in Table 3, calculate the missing values in the blank columns for  $v_p(h)$ ,  $v_p/r_h$ ,  $p(\Delta, h)$  and AIN using Table 1 and 2 and assuming an approximate focal depth for the recorded event of  $h = 300$  km. Interpolate linearly as a first approximation.

#### Task 2:

Decide whether your ray has left the upper or lower half of the focal sphere and whether or not you need to calculate AINc and/or AZMc according to Figs. 3.28 and 3.29 in Chapter 3. Complete Table 3 accordingly.

#### Task 3:

Use the values given in Figure 1 for  $\phi$  and  $\delta$  for the nodal planes NP1 and NP2 of the NEIC fault-plane solution. Reconstruct both nodal (fault) planes using the *Lambert-Schmidt net* (Fig. 3.27b) by applying the procedure inverse to the one described in the Exercise : Determination of fault-plane solutions (EX 3.2). Compare your nodal-plane pattern with that of the NEIC "beach-ball" solution (Figure 1 upper right).

#### Task 4:

Find the corresponding equatorial plane to your NP1 and NP2 and mark the locations of the P and T axes on the focal sphere. Draw the P and T vectors towards and from the center of the net, determine their azimuth (Azm) and plunge (Plg) [equivalent to dip, measured from the horizontal]. Compare your respective values with those given by NEIC in Figure 1.

#### Task 5:

Use the values that you calculated for the P-wave take-off angle AINc and ray azimuth AZMc to all 5 stations in Table 3 and mark the point where the ray penetrates the focal sphere and indicate the respective polarity. Check whether they fall into the proper T and P quadrants and whether the short-period polarity readings given in Table 3 are consistent with the fault-plane solution published by the NEIC which is based on long-period waveform data.

### 4 Solutions and discussions

**Table 4** Solutions for Task 1.

| STA | $\Delta$<br>(deg) | AZM<br>(deg) | POL | $v_p(h)$<br>km/s | $v_p/r_h$<br>(s <sup>-1</sup> ) | $p(\Delta, h)$<br>(s/deg) | AIN<br>(deg) | AZMc<br>(deg) | AINc<br>(deg) |
|-----|-------------------|--------------|-----|------------------|---------------------------------|---------------------------|--------------|---------------|---------------|
| SGG | 2.30              | 345          | +   | 8.6286           | $1.4213 \times 10^{-3}$         | 8.368                     | (137.0)      | 165           | 43.0          |
| KHC | 10.03             | 354          | -   |                  |                                 | 12.258                    | 86.5         |               |               |
| BTH | 12.25             | 294          | +   |                  |                                 | 11.984                    | 77.4.1       |               |               |
| ZAK | 60.02             | 48           | -   |                  |                                 | 6.759                     | 33.4         |               |               |
| PAE | 154.8             | 324          | -   |                  |                                 | 1.368                     | 6.4          |               |               |



**Task 2:**

**Note:** In the case of deep earthquakes the values of the ray parameter (and thus slowness) may increase with  $\Delta$ , e.g., in Table 2 for  $h = 300$  km up to  $\Delta = 10^\circ$ . This corresponds to seismic rays that leave the source upwards! Consequently, the value  $AIN = 43.0^\circ$  calculated with Equation (1) for station SGG corresponds, according to the definition given in Fig. 3.28, in fact to an angle of  $180^\circ - AIN = 137.0^\circ$ . Accordingly,  $AZMc$  and  $AINc$  for the equivalent lower hemisphere projection of this ray are  $345^\circ - 180^\circ = 165^\circ$  and  $43.0^\circ$ , respectively.

**Task 3:**

Your manually drawn fault-plane solutions should look very similar to that of the NEIC solution in Figure 1 upper right.

**Task 4:**

Your manually re-constructed values for  $Azm$  and  $Plg$  of the P and T axes should agree with the NEIC solution within a few degrees ( $<10^\circ$ ). If not, check your drawing of the three planes, of the related P and T axis and the measured angles.

**Task 5:**

The short-period polarity data used in this exercise are consistent with the fault-plane solution published by the NEIC which is based on long-period waveform data. All your polarities should fall properly into quadrants of either observed compressional or dilatational P-wave first motions.

**References**

Kennett, B. L. N. (Ed.) (1991). IASPEI 1991 Seismological Tables. Research School of Earth Sciences, Australian National University, 167 pp.

| Topic   | Determination of source parameters from seismic spectra                                                                                                                                                                                                         |
|---------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | Michael Baumbach  , and Peter Bormann (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany);<br>E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version | October 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_EX_3.4">10.2312/GFZ.NMSOP-2_EX_3.4</a>                                                                                                                                                          |

## 1 Aim

This exercise shows how to estimate the source parameters seismic moment, size of the rupture plane, source dislocation and stress drop from data in the frequency domain only and how the results depend on the underlying model assumptions. These parameters could also be estimated in the time domain. However, for estimation in the time domain the records have to be converted into true ground motion (displacement) records. This may be a problem if the bandwidth of the recording system is limited (e.g., short-period records) or if the phase response of the system is not well known. For estimation in the frequency domain only the amplitude response of the instrument is needed.

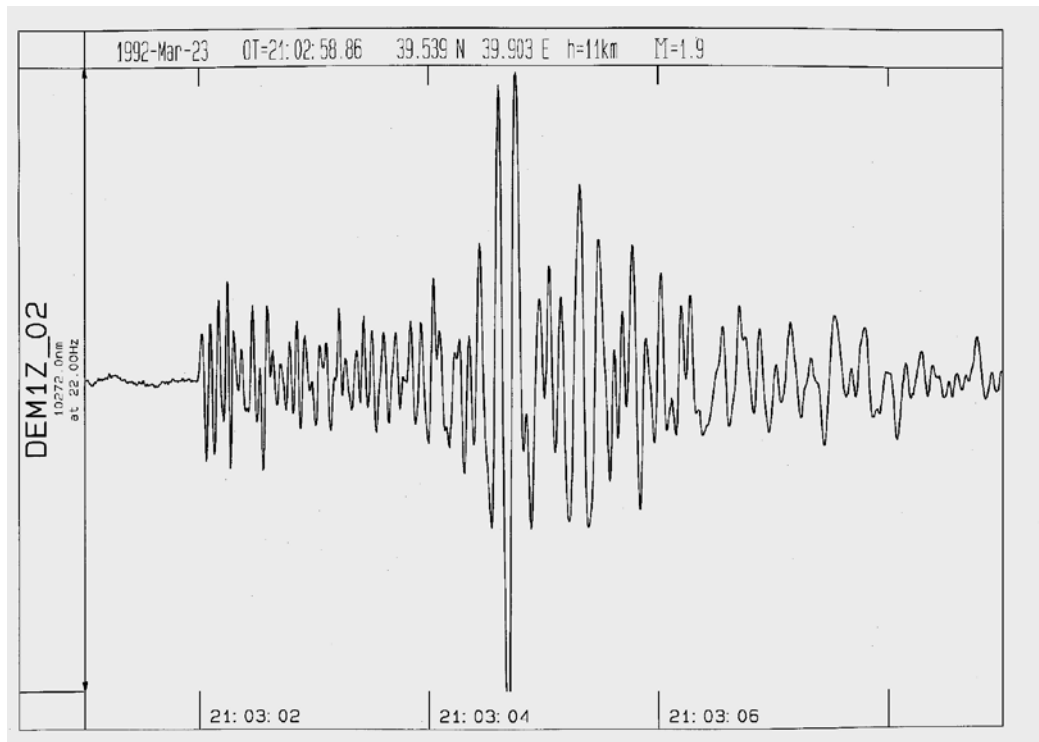
## 2 Data

Figure 1 shows a velocity record (vertical component) of an aftershock of the 1992 Erzincan earthquake (Turkey). Figure 2 shows the corresponding *displacement spectrum* of the P wave. The calculated spectrum was corrected for the amplitude response of the recording system (which includes both response of the velocity seismometer and the anti-aliasing filter of the recorder). Furthermore, the P-wave spectrum was corrected for attenuation,  $\exp(i\omega t/2Q_p)$ .  $Q_p$  had been estimated beforehand from coda  $Q_c$ - observations in the area under study assuming that  $Q_p = 2.25 Q_c$ . This is a good approximation under the assumption that  $v_p/v_s = 1.73$ ,  $Q_c = Q_s$  and the pure compressional  $Q_\kappa$  ( $\kappa$  - bulk modulus) is very large ( $\rightarrow \infty$ ). In Figure 2 also the noise spectrum, treated in the same way as the P-wave spectrum, was computed and plotted in order to select the suitable frequency range for analysis (with signal-to-noise ratio  $SNR > 3$ ).

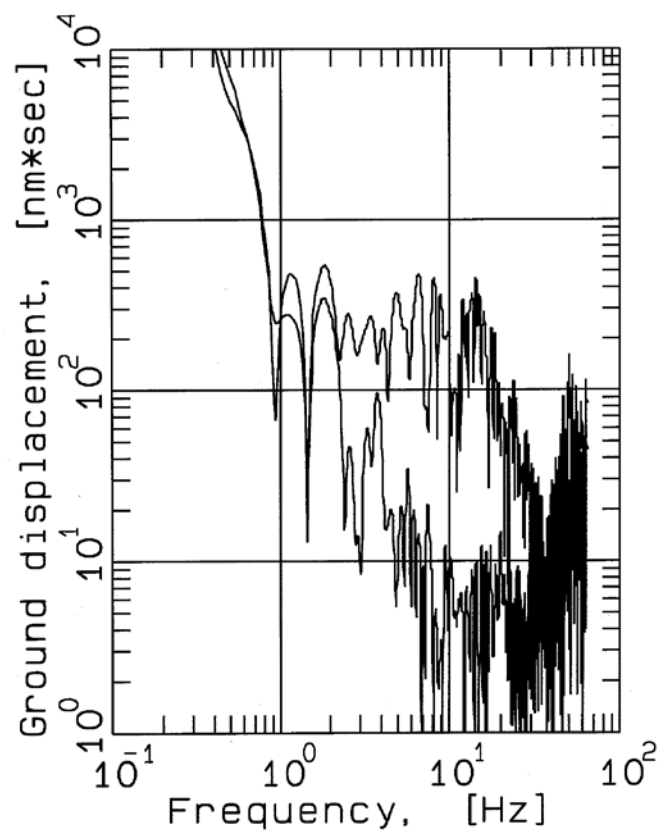
At low frequencies typical P- and S-wave spectra approach a constant amplitude level  $u_0$  and at high frequencies the spectra show a decay that falls off as  $f^{-2}$  to  $f^{-3}$ . Plotted on a log-log scale the spectrum can be approximated by two straight lines. The intersection point is the corner frequency  $f_c$ .  $u_0$  and  $f_c$  are the basic spectral data from which the source parameters will be estimated. Event and material data required for further calculations are the epicentral distance  $\Delta$ , the source depth  $h$ , the rock density  $\rho$ , the P-wave velocity  $v_p$ , and the averaged radiation pattern  $\Theta$  for P waves. Respective values are given under 3.1 below. Other needed parameters can then be calculated.

**Note:** The apparent increase of spectral noise amplitudes in Figure 2 for  $f > 25$  Hz and of signal amplitudes for  $f > 50$  Hz is not real but sampling noise due to anti-aliasing filtering of the record. Thus, this increase should not be considered in the following analysis. The same

holds for spectral amplitudes  $f < 1$  Hz. Since the analyzed P-wave train is only 2 s long, lower frequencies are not properly represented.



**Figure 1** Record of an Erzincan aftershock (vertical component). For the indicated P-wave window the displacement spectrum shown in Figure 2 has been calculated.



**Figure 2** P-wave spectrum (upper curve) and noise spectrum (lower curve) of the record shown in Figure 1, corrected for the instrument response and attenuation.

### 3 Procedures

The parameters to be estimated are:

- Seismic moment  $M_0 = \mu \bar{D} A$  (1)  
(with  $\mu$  - shear modulus;  $\bar{D}$  - average source dislocation;  $A$  - size of the rupture plane)
- Source dislocation  $\bar{D}$
- Source dimension (radius  $R$  and area  $A$ )
- Stress drop  $\Delta\sigma$

#### 3.1 Seismic moment $M_0$

Under the assumption of a homogeneous Earth model and constant P-wave velocity  $v_p$ , the seismic moment  $M_0$  can be determined from the relationship:

$$M_0 = 4 \pi r v_p^3 \rho u_0 / (\Theta S_a) \quad (2)$$

with  $r$  – hypocentral distance,  $\rho$  - density,  $u_0$  – low-frequency level (plateau) of the displacement spectrum,  $\Theta$  - average radiation pattern and  $S_a$  surface amplification for P waves.

In the exercise we use the following values: density

$$\rho = 2.7 \text{ g/cm}^3$$

P-wave velocity

$$v_p = 6 \text{ km/sec}$$

source depth

$$h = 11.3 \text{ km}$$

epicentral distance

$$\Delta = 18.0 \text{ km}$$

hypocentral distance  
(travel path)

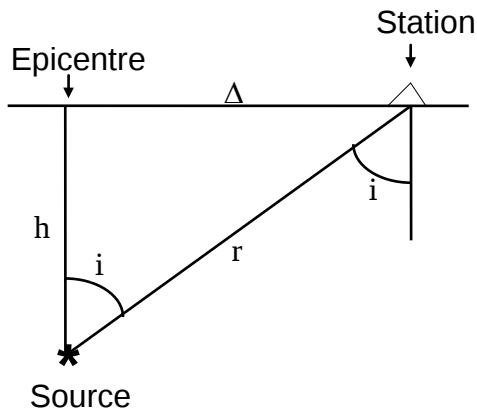
$$r = \sqrt{(h^2 + \Delta^2)}$$

incidence angle

$$i = \arccos(h/r)$$

free surface amplification  $S_a$  for P waves

averaged radiation pattern  $\Theta = 0.64$  for P waves.



**Note** the differences in dimensions used!  $M_0$  has to be expressed in the unit  $\text{Nm} = \text{kg m}^2 \text{s}^{-2}$ .  $S_a$  can be determined by linear interpolation between the values given in Table 1. They were computed for the above given constant values of  $v_p$  and  $\rho$  (homogeneous model) and assuming a ratio  $v_p/v_s = 1.73$ .  $i$  is the angle of incidence, measured from the vertical.

**Table 1** Surface amplification  $S_a$  for P waves;  $i$  is the incidence angle.

| $i$ | $S_a$ | $i$ | $S_a$ | $i$ | $S_a$ |
|-----|-------|-----|-------|-----|-------|
| 0   | 2.00  | 30  | 1.70  | 60  | 1.02  |
| 5   | 1.99  | 35  | 1.60  | 65  | 0.90  |
| 10  | 1.96  | 40  | 1.49  | 70  | 0.79  |
| 15  | 1.92  | 45  | 1.38  | 75  | 0.67  |

|    |      |    |      |    |      |
|----|------|----|------|----|------|
| 20 | 1.86 | 50 | 1.26 | 80 | 0.54 |
| 25 | 1.79 | 55 | 1.14 | 85 | 0.35 |

### 3.2 Size of the rupture plane

For estimating the size of the rupture plane and the source dislocation one has to adopt a kinematic (geometrical) model, describing the rupture propagation and the geometrical shape of the rupture area. In this exercise computations are made for three different circular models (see Table 2), which differ in the source time function and the crack velocity  $v_{cr}$ .  $v_s$  is the S-wave velocity, which is commonly assumed to be  $v_s = v_p / \sqrt{3}$ .

**Table 2** Parameters of some commonly used kinematic rupture models.

|                        |                    |              |              |
|------------------------|--------------------|--------------|--------------|
| 1. Brune (1970)        | $v_{cr} = 0.9 V_s$ | $K_p = 3.36$ | $K_s = 2.34$ |
| 2. Madariaga I (1976)  | $v_{cr} = 0.6 V_s$ | $K_p = 1.88$ | $K_s = 1.32$ |
| 3. Madariaga II (1976) | $v_{cr} = 0.9 V_s$ | $K_p = 2.07$ | $K_s = 1.38$ |

The source radius  $R$  (in km) can then be computed from the relationship

$$R = v_s K_{p/s} / 2\pi f_{c_{p/s}} \quad (3)$$

with  $v_s$  – shear-wave velocity in km/s,  $f_{c_{p/s}}$  – corner frequency of the P or S waves, respectively, in Hz and  $K_p$  and  $K_s$  being the related model constants and  $v_s$ . The differences in  $K_p$  and  $K_s$  between the various models are due to different assumptions with respect to crack velocity and the rise time of the source-time function. Only  $K_p$  has to be used in the exercise (P-wave record!). The size of the circular rupture plane is then

$$A = \pi R^2. \quad (4)$$

### 3.3 Average source dislocation $\bar{D}$

According to (1) the average source dislocation is

$$\bar{D} = M_0 / (\mu A). \quad (5)$$

Assuming  $v_s = v_p / 1.73$  it can be computed knowing  $M_0$ , the source area  $A$  and the shear modulus  $\mu = v_s^2 \rho$ .

### 3.4 Stress drop

The static stress drop  $\Delta\sigma$  describes the difference in shear stress on the fault plane before and after the slip. According to Keilis-Borok (1959) the following relationship holds for a circular crack with a homogeneous stress drop:

$$\Delta\sigma = 7 M_0 / (16 R^3). \quad (6)$$

The stress drop is expressed in the unit of Pascal,  $\text{Pa} = \text{N m}^{-2} = \text{kg m}^{-1} \text{s}^{-2} = 10^{-5} \text{ bar}$ .

## 4 Tasks

### Task 1:

Select in Figure 2 the frequency range  $f_1$  to  $f_2$  that can be used for analysis ( $\text{SNR} > 3$ ):

$$f_1 = \dots\dots\dots \text{ Hz}$$

$$f_2 = \dots\dots\dots \text{ Hz}$$

### Task 2:

Estimate the low-frequency level,  $u_o$ , of the spectrum by approximating it with a horizontal line. Note in Figure 2 the logarithmic scales and that the ordinate dimension is  $\text{nm s} = 10^{-9} \text{ m s}$ .

$$u_o = \dots\dots\dots \text{ m s}$$

### Task 3:

Estimate the exponent,  $n$ , of the high-frequency decay,  $f^{-n}$ ; mark it by an inclined straight line.

$$n = \dots\dots\dots$$

### Task 4:

Estimate the corner frequency,  $f_{c_p}$  (the intersection between the two drawn straight lines).

$$f_{c_p} = \dots\dots\dots \text{ Hz}$$

### Task 5:

Calculate from the given event parameters and relationships given in 3.1 and Table 1 the values for:

$$r = \dots\dots\dots \text{ km}$$

$$i = \dots\dots\dots^\circ$$

$$S_a = \dots\dots\dots$$

$$M_o = \dots\dots\dots \text{ Nm.}$$

### Task 6:

Using the equations (3), (4), (5) and (6) calculate for the three circular source models given in Table 2 the parameters

- source radius  $R$  and source area  $A$ ,
- shear modulus  $\mu$  and average displacement  $\bar{D}$  and
- stress drop  $\Delta\sigma$ .

Write the respective values in Table 3

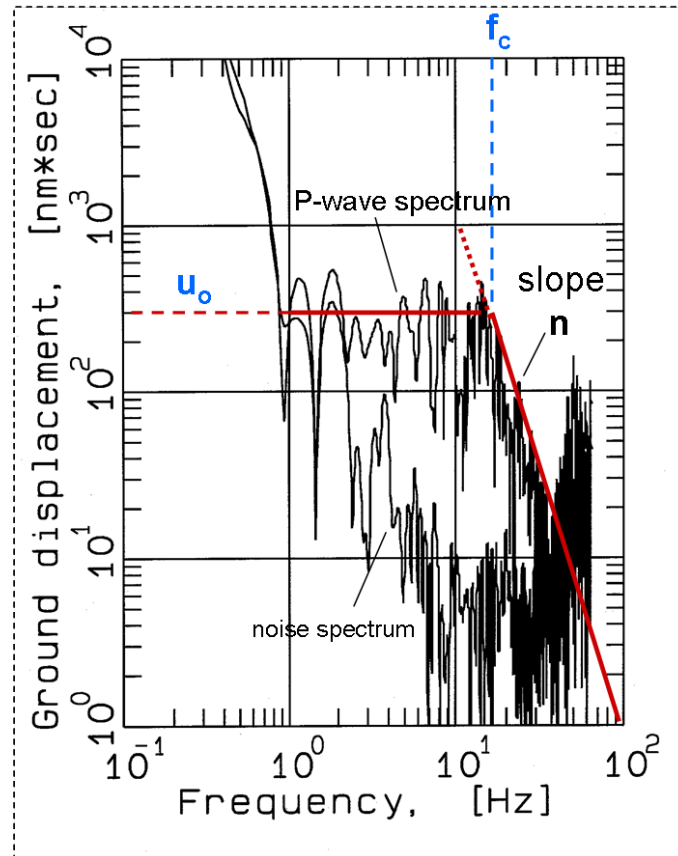
**Table 3**

| Model          | $R$ [m] | $A$ [m <sup>2</sup> ] | $\bar{D}$ [m] | $\Delta\sigma$ [MPa] |
|----------------|---------|-----------------------|---------------|----------------------|
| 1. Brune       |         |                       |               |                      |
| 2. Madariaga I |         |                       |               |                      |

3. Madariaga II

## 5 Solutions

Although individual visual parameter readings from Figure 2 might be subjective, they should not differ by more than about  $\pm 10\%$  from the values given here for tasks 1 to 5 but may be larger for 6. Acceptable average values for the read and calculated parameters are for:



**Task 1:**  $f_1 = 2 \text{ Hz}$ ,  $f_2 = 30 \text{ Hz}$

**Task 2:**  $u_o = 3 \times 10^{-7} \text{ m s}$

**Task 3:**  $n = 3$

**Task 4:**  $f_{cp} = 14.4 \text{ Hz}$

**Task 5:**  $r = 21.3 \text{ km}$   $i = 58^\circ$ ,  $S_a = 1.07$ ,  $M_o = 6.8 \times 10^{13} \text{ N m}$

**Task 6:**

a)  $R_1 = 129 \text{ m}$ ,  $A_1 = 5.23 \times 10^4 \text{ m}^2$   
 $R_2 = 72 \text{ m}$ ,  $A_2 = 1.63 \times 10^4 \text{ m}^2$   
 $R_3 = 79 \text{ m}$ ,  $A_3 = 1.96 \times 10^4 \text{ m}^2$

- b)  $\mu = 3.24 \times 10^{10} \text{ kg m}^{-1} \text{ s}^{-2}$   $D_1 = 4.0 \times 10^{-2} \text{ m}$   
 $D_2 = 1.3 \times 10^{-1} \text{ m}$   
 $D_3 = 1.1 \times 10^{-1} \text{ m}$
- c)  $\Delta\sigma_1 = 13.8 \text{ MPa}$   
 $\Delta\sigma_2 = 79.7 \text{ MPa}$   
 $\Delta\sigma_3 = 60.3 \text{ MPa}$

## 6 Comments

One should always be aware that the interpretation of seismic source spectra is based on simplified model assumptions of the generally unknown *source geometry* (e.g., circular fault or rectangular fault with different, also magnitude dependent aspect ratios), as well as unknown *rupture kinematics* (rupture velocity and direction, e.g. uni- and/or bilateral and – directional; see Figure 6 in IS 3.1), and *dynamics* (stress drop). Different model assumption will yield different results, even for identical spectral parameter readings, as demonstrated with the above solutions. They may differ even for the most common models of small earthquakes by a factor of about two for the source “radius” and more than 5 for the stress drop (because of  $\Delta\sigma \sim R^{-3}$ ).

Moreover, also the correction of measured spectra for wave propagation effects (geometric spreading, scattering and intrinsic attenuation) is largely model-based and thus error-prone. And the reading from rather noisy empirical spectra of the spectral plateau amplitude - and in particular of the corner frequency - is afflicted with relatively large errors too. All this increases further the uncertainty of source radius and especially stress-drop estimates. The latter may be, in the worst case, uncertain up to about two orders of magnitude.

Therefore, in studying possible systematic differences in source parameters derived from spectral data for events in a given area, or within an aftershock sequence, one should always stick to using one type of model only, even if this might not be the best in absolute terms with respect to the given (but usually not yet well known) specific seismotectonic situation. Yet, one should at least be reasonably sure about the validity of assuming that the events have similar modes of faulting and crack propagation. If later models become available which better seem to describe the seismic source processes in the area and/or in the magnitude range under investigation then earlier, maybe inferior but consistent results, may be “rescaled”.

## References

- Brune, J. N. (1970). Tectonic stress and the spectra of shear waves from earthquakes. *J. Geophys. Res.*, **75**, 4997-5009.
- Keilis-Borok, V. I., (1959). On the estimation of displacement in an earthquake source and of source dimension. *Ann. Geofis.*, **12**, 205-214.
- Madariaga, R. (1976). Dynamics of an expanding circular fault. *Bull. Seism. Soc. Am.*, **66**, 639-666.



| Topic   | Moment-tensor determination and decomposition                                                                                                                                                                                                                                                                            |
|---------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | Frank Krüger, Potsdam University, Institute of Geosciences, 14415 Potsdam,<br>E-mail: <a href="mailto:kruegerf@geo.uni-potsdam.de">kruegerf@geo.uni-potsdam.de</a><br><br>Günter Bock  (formerly GeoForschungsZentrum Potsdam, Germany) |
| Version | August 2000                                                                                                                                                                                                                                                                                                              |

## 1 Aim

The tasks in this exercise are aimed at making you more familiar with the use and meaning of the basic equations and matrix formalisms in seismic moment-tensor presentation and decomposition.

## 2 Formulae used

The following formulae are used:

$$d_s(x, t) = M_{kj} [G_{skj}(x, \xi, t) * s(t)] \quad (1)$$

which is identical with Eq. (3.70) in Chapter 3, with  $d_s(x, t)$  - ground displacement at position  $x$  and time  $t$ . (1) simplifies to  $d_s(x, t) = M_{kj} G_{skj}(t)$  if the source-time function  $s(t) = \delta(t)$  is a needle (spike) impulse.

$$M_{kj} = A \{ \Lambda v_i s_i \delta_{kj} + \mu (v_k s_j + v_j s_k) \} \quad (2)$$

which describes the seismic moment tensor for an isotropic medium in the most general way with  $\Lambda$  - elastic Lamé-parameter,  $\mu$  - shear modulus,  $A$  - area of fault rupture,  $s$  - slip vector on the fault and  $v$  - normal to the fault plane. Note that the scalar seismic moment  $M_0 = \mu A |s|$ . The term  $(v_k s_j + v_j s_k)$  forms a tensor  $D$  describing a double-couple source. In case of an explosion it is zero. And the equations:

$$\begin{aligned}
 M_{xx} &= -M_0(\sin\delta \cos\lambda \sin 2\phi + \sin 2\delta \sin\lambda \sin^2\phi) \\
 M_{xy} &= M_0(\sin\delta \cos\lambda \cos 2\phi + 0.5 \sin 2\delta \sin\lambda \sin 2\phi) \\
 M_{xz} &= -M_0(\cos\delta \cos\lambda \cos\phi + \cos 2\delta \sin\lambda \sin\phi) \\
 M_{yy} &= M_0(\sin\delta \cos\lambda \sin 2\phi - \sin 2\delta \sin\lambda \cos^2\phi) \\
 M_{yz} &= -M_0(\cos\delta \cos\lambda \sin\phi - \cos 2\delta \sin\lambda \cos\phi) \\
 M_{zz} &= M_0 \sin 2\delta \sin\lambda
 \end{aligned} \quad (3)$$

with  $\phi$  - strike direction and  $\delta$  - dip angle of the rupture plane, and  $\lambda$  - slip direction (rake angle).

### 3 Tasks

**Task 1:**

By using Equations (2) and (3) above, respectively, determine the Cartesian moment tensors for:

- a) an underground nuclear explosion;
- b) a double-couple focal mechanism with strike  $\phi = 0^\circ$ , dip  $\delta = 90^\circ$ , and rake  $\lambda = 0^\circ$ ;
- c) a double-couple focal mechanism with  $\phi = 0^\circ$ ,  $\delta = 45^\circ$ , and  $\lambda = 90^\circ$ ;
- d) a double-couple focal mechanism with  $\phi = 0^\circ$ ,  $\delta = 90^\circ$ , and  $\lambda = 90^\circ$ .

**Task 2:**

Determine the moment tensor for a tension crack in the direction normal to the fault plane in a homogenous isotropic medium. Use Equation (2) of the exercise.

**Task 3:**

The relation Equation (3) between moment tensor and parameters of a shear dislocation can be expressed as a weighted sum of 4 elementary moment tensors:

$$M = \cos\delta \cos\lambda M_1 + \sin\delta \cos\lambda M_2 - \cos 2\delta \sin\lambda M_3 + \sin 2\delta \sin\lambda M_4.$$

Derive the elements of  $M_1$ ,  $M_2$ ,  $M_3$ , and  $M_4$  and discuss the shear dislocations that are represented by the elementary moment tensors.

### 4 Solutions

**Task 1:**

$$\text{a) } M = M_0 \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix}$$

$$\text{b) } M = M_0 \begin{pmatrix} 0 & 1 & 0 \\ 1 & 0 & 0 \\ 0 & 0 & 0 \end{pmatrix}$$

$$\text{c) } M = M_0 \begin{pmatrix} 0 & 0 & 0 \\ 0 & -1 & 0 \\ 0 & 0 & 1 \end{pmatrix}$$

$$\text{d) } M = M_0 \begin{pmatrix} 0 & 0 & 0 \\ 0 & 0 & -1 \\ 0 & -1 & 0 \end{pmatrix}$$

**Task 2:**

$$M = \begin{pmatrix} \Lambda s_3 & 0 & 0 \\ 0 & \Lambda s_3 & 0 \\ 0 & 0 & (\Lambda + 2\mu)s_3 \end{pmatrix}$$

**Task 3:**

$$M_1 = M_0 \begin{pmatrix} 0 & 0 & -\cos\phi \\ 0 & 0 & -\sin\phi \\ -\cos\phi & -\sin\phi & 0 \end{pmatrix} \rightarrow \delta = 0^\circ; \lambda = 0^\circ, \text{ i.e., horizontal slip on horizontal fault}$$

$$M_2 = M_0 \begin{pmatrix} -\sin 2\phi & \cos 2\phi & 0 \\ \cos 2\phi & \sin 2\phi & 0 \\ 0 & 0 & 0 \end{pmatrix} \rightarrow \delta = 90^\circ; \lambda = 0^\circ, \text{ i.e., strike slip on vertical fault}$$

$$M_3 = M_0 \begin{pmatrix} 0 & 0 & \sin\phi \\ 0 & 0 & -\cos\phi \\ \sin\phi & -\cos\phi & 0 \end{pmatrix} \rightarrow \delta = 90^\circ; \lambda = 90^\circ, \text{ i.e., dip slip on a vertical fault}$$

$$M_4 = M_0 \begin{pmatrix} -\sin^2\phi & \frac{1}{2}\sin 2\phi & 0 \\ \frac{1}{2}\sin 2\phi & -\cos^2\phi & 0 \\ 0 & 0 & 1 \end{pmatrix} \rightarrow \delta = 45^\circ; \lambda = 90^\circ, \text{ i.e. dip slip on a } 45^\circ \text{ dipping fault.}$$



| Title   | A practical on moment tensor inversion using the Kiwi tools                                                                                                                                                                                                                                                                                                                                                             |
|---------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <p style="text-align: center;"><b>Simone Cesca<sup>1,2</sup> and Sebastian Heimann<sup>1</sup></b></p> <p><sup>1</sup>Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Department 2.1, D-14473 Potsdam, Germany; E-mail: <a href="mailto:simone.cesca@gfz-potsdam.de">simone.cesca@gfz-potsdam.de</a></p> <p><sup>2</sup>University of Potsdam, Institute for Earth and Environmental Sciences</p> |
| Version | July 2013; DOI: <b>10.2312/GFZ.NMSOP-2_EX_3.6</b>                                                                                                                                                                                                                                                                                                                                                                       |

## 1 Introduction

This tutorial with exercises introduces to the use of the Kiwi tools (Heimann, 2011; Cesca et al., 2010), an open source seismological software developed at the University of Hamburg and the GFZ Potsdam for the goal of moment tensor inversion and finite source inversion. The following text is accompanied by an electronic supplement, including a virtual machine (KiwiVirtualBox) with a Linux Ubuntu system. The virtual machine includes all required software, manuals, and data examples here discussed. It can be downloaded at [10.5880/GFZ.2.1.2013.001](https://doi.org/10.5880/GFZ.2.1.2013.001).

The aim is to guide the reader through the different modules required to perform a moment tensor inversion. All moment tensor and source inversion routines in seismology require access to the following components: Green's functions, seismological data, inversion framework and results assessment.

The following sections introduce the reader to the Kiwi tools devoted to the Green's function databases handling, data access and visualization, selection of the inversion parameters and inversion process, and results visualization. While describing the most important software tools and their more relevant options, specific examples guide the reader through the different steps of the inversion process. In conclusion, a number of relevant exercises are provided.

## 2 Green's functions

Green's functions (GFs) describe, on the basis of model assumptions or empirically gained data, the wave propagation from the seismic source to a receiver, in response to specific source excitation. In the framework of source inversion problems, they are used to build synthetic seismograms, which are compared to real surface data records in the moment tensor inversion.

In the current setup of the Kiwi tools, only 1D velocity models are considered. This means, Earth models for which all relevant properties (P and S wave velocities, density, attenuation coefficients) only change with depth. A 1D case presents certain advantages in terms of Green's function computation. First, several codes are nowadays freely available to solve the forward problem. Moreover, because of the symmetry of the model, it is only required to compute a limited number of Green's functions. For any combination of source depth and source-receiver distance, 10 Green's functions have to be computed. These 10 Green's

functions can be combined using proper weighting to compute the response to any moment tensor configuration at any azimuth.

Green's functions are typically stored into databases, which we refer here as Green's function database (GFDB). A GFDB may require a significant computational effort, but is then ready for use and significantly reduces computational time of the inversion. The Kiwi tools currently handle GFDB through HDF5 (the dependence on HDF5 may be removed in future Kiwi versions). The system is independent on the software used to generate GFs, however we have successfully worked so far with Qseis (Wang, 1999) and Gemini (Friedrich and Dalkolmo, 1995). GFDBs are built for any required velocity model with user-defined constraints. In particular, the following parameters have to be chosen when preparing a GFDB:

- Description of the velocity model,
- Range of epicentral distances,
- Range of source depths,
- GF sampling.

When preparing an own GFDB, these parameters have to be chosen carefully with respect to the area of study, network geometry, expected range of magnitudes and source depths, and sensor instrumentation (sampling rate, frequency range). The range of distances and depths limits the possible application of the GFDB, e.g. synthetic seismogram for too far away stations or too deep sources cannot be computed when exceeding the spatial boundary of a GFDB. Epicentral distances and depths should be sampled equidistant. The sampling interval limits the resolution potential of a GFDB: first, because it limits the maximal reproducible frequencies, and second because the inversion framework requires a common sampling of data and GFs.

The KiwiVirtualBox includes a single GFDB. This was generated with the software Qseis (Wang, 1999) for a global velocity model, IASP91 (Kennett & Engdahl, 1991). The available GFDB allows to build synthetic seismograms for sources down to 40 km and stations up to 1000 km epicentral distance with a spatial sampling of 1 km. Thus GFs are available every 1 km in distance and depth and the computation of synthetic seismograms in between grid points can either be done through GFs interpolation or by using the GFs of the closest grid point. The sampling is 0.5 s, so that the spectral fit cannot exceed 1 Hz.

The GFDB is found at the directory */GFDB*; its name is *GFDB/db*.

To get some information on the GFDB type:

*gfdb\_info GFDB/db*

In turn you get information on the sampling (dt), spatial GF sampling along epicentral distances and source depths (dx, dz), minimal epicentral distance and source depth (firstx, firstz), number of epicentral distances and source depths (nx, nz), number of GFs per source depth-epicentral distance geometry (ng, typically 10), and some additional technical information on the number of files used to allocate GFs (chunks) and the state of filling of a GFDB (for large GFDBs, with significant GF computation time, the available database may still be incomplete). Temporal and spatial units are always given in seconds and meters.

To extract a Green's function (in the example we extract from the database GFDB/db at a distance of 400000 m and source depth of 20000 m the GF number 1, and store it in the file `gf1.d400km.z20km.mseed`) in miniSEED format type (note that using a different extension than `.mseed` provides the GF in the default ASCII format):

```
gfdb_extract GFDB/db <<EOF
400000 20000 1 gf1.d400km.z20km.mseed
EOF
```

To view the Green's function with the snuffler visualizer type:

```
snuffler gf1.d400km.z20km.mseed
```

A new GFDB can be created using the Kiwi command `gfdb_build`, while a subset of an existing database can be created using the command `gfdb_redeploy`. Testing these commands with no argument, provide more information on their syntax:

```
gfdb_build
gfdb_redeploy
```

Further information on other Kiwi tools to generate and handle Green's function databases are available in the enclosed documentation (in the directory *DOCS*).

### 3 Data

The data, here intended as displacement waveforms, are an obvious prerequisite for any source inversion tool. Their characteristics, e.g., their sampling rate or the limited bandwidth of the flat displacement-proportional frequency response of the recording instrument can severely limit the inversion potential.

Kiwi can currently handle different formats, including the preferred miniSEED (see Chapter 10) and a simple ASCII format (first column time, second column displacement amplitude). Sampling should be common for any single study (e.g., for any treated earthquake case), and common to the adopted GFDB. Amplitudes are given in meters, times in seconds, with respect to the reference system timestamp, January 1, 1970, 00:00:00.

Data preprocessing is not described in detail here (see Scherbaum, 2007). However, displacement traces are typically obtained from velocity seismometers, through data recovery, tapering (see IS 14.1), removal of mean and trend and restitution to displacements, according to instrument response function. Data may also need to be converted to any of the usable data formats.

Displacement files for a single study (earthquake) should be stored in a single event directory, with the following naming:

```
DISPL.$STAT.$CHAN
```

where `$STAT` is the station code, and `$CHAN` the channel code (BHZ: Vertical, BHN: North, BHE: East, BHR: Radial, BHT: Transversal). The adoption of channel “BH” is only a naming convention, and does not limit the application of the Kiwi tools to any instrument type or channel.

The KiwiVirtualBox includes several datasets, which are described in the following paragraphs. All data are stored as miniSEED files, with a sampling of 0.5 s (common to the available GFDB). All data were recorded by velocity broadband sensors, and can be safely analyzed in any frequency band above 0.01 Hz and below 1 Hz.

An example of directory structure for a synthetic dataset can be found at:

```
DATA/SYNTH1
```

which is a synthetic example. Displacement waveforms have been created there for a pure double couple (DC) point source (location, and origin time are given in the last section), with the available GFDB and no noise (therefore they can be perfectly matched in the test inversion).

To visualize the typical content of a data directory type:

```
ls DATA/SYNTH1
```

It indicates files corresponding to displacement traces, plus an additional station file.

To look at the miniSEED waveforms, use the snuffler software:

```
snuffler DATA/SYNTH1/DISPL.*
```

The following command allows to include event information and to perform some data sorting:

```
snuffler DATA/SYNTH1/DISPL.* \  
--stations="META/stations.dat" --event="META/events.txt"
```

An event is then selected by typing “e” when the cursor is on the event label (a rectangle should appear around the label).

A left click on the event label opens a window with possible operations.

For example, it is possible to sort traces according to their epicentral distance from the selected event.

Repeating the operation with the following attribute allows to compare observed waveforms with synthetics for provided focal mechanisms.

```
snuffler DATA/SYNTH1/DISPL.* \  
--stations="META/stations.dat" --event="META/events_w_solutions.txt"
```

It is now possible to modify the frequency filter applied to both the recorded and the synthetic traces, and to visually inspect the effect of frequency filtering on both traces and on their fit.

Apart from the displacement files, the data directory includes one more file: *stations.dat*. This ASCII file, which can be opened with any text editor, is very important; the inversion cannot be performed without it. It include several lines, one per each desired station. Each line is composed by a station number (an increasing integer), the station code, latitude and longitude (in degrees).

Note that, when running the inversion and select a given station file, only stations listed in the station file and for which properly named traces are available in the data directory will be used. This implies that: (a) stations can be temporarily omitted, simply by file renaming (e.g.



removed.DISPL.STAT.BHZ), (b) several station files can be created, to simulate different network configurations, with no need to duplicate data.

## 4 Inversion setup

We refer to the inversion framework, in the most general sense, as to the combination of inversion software, scripts, input files and the selection of inversion parameters. In the general formulation of a source inversion process, the data and GFs have to be made accessible. Then, a source model is first considered, synthetic seismograms are computed on the basis of a selected source model and the adopted GFs, synthetic seismograms are then compared to observations, and a fit among data and synthetics is estimated. On the base of such a fit, the source model is modified, perturbed or “improved”, and the process iteratively repeated, until certain conditions are met. At the end of the process, the information regarding the best fitting model is released, typically together with amisfit estimation and suitable evaluation plots, such as the fault-plane (“beach-ball”) solution (see., e.g., section 3.4 of Chapter 3), waveform fits, etc..

The inversion process is for sure to some extent non unique, since the results are affected both by the choice of the (usually oversimplified or sometimes even not appropriate) velocity model and of the inversion parameters. E.g., , with respect to the latter only, a specific station selection covering different azimuthal and distance ranges, different frequency bands analyzed, the fitting of body waves only or of full waveforms or the performance of the inversion through either waveform or amplitude spectra fitting, all could contribute to significant differences in the inversion results. If a good quality dataset should ensure a minor deviation between results from different inversion setups, then this requires the reasonable selection of the underlying model and inversion parameters and of the chosen inversion approach. These should be tuned as good as possible to the event size, its depth, the reliability of the available velocity model, the seismic instrumentation, data acquisition and processing, and the source-stations geometry.

Here, we describe how to use the *rapidinv* python code to perform the inversion. *rapidinv* allows the user to customize the inversion procedure, within certain limits, and make use of the Kiwi tools inversion modules without directly dealing with these tools. *rapidinv* acts as a user friendly interface to these tools. Its potential may be further exploited by advanced users. In particular, *rapidinv* internally calls several times the Kiwi tool *minimizer*, which is used to assess source model parameters, generate synthetic seismograms, filter and taper synthetics and observed seismograms, compare them and compute misfits, perform grid search or Levenberg-Marquardt inversion for selected source parameters, and export seismograms in different formats.

*rapidinv* always performs the inversion in 3 separate steps, as described in Cesca et al. (2010) (DC and finite source inversion), and Cesca et al. (2013) (DC and full MT inversion). The first and second steps have to be carried out to obtain a point source solution, the third step is only required to perform an additional finite, kinematic source inversion.

The first step is typically performed as an amplitude spectra inversion. The pure DC results include strike, dip, rake (4 configurations because of the polarity ambiguity of the focal mechanism), centroid depth, scalar moment and magnitude. The full moment tensor (MT)

results additionally include moment tensor components (2 configurations because of the polarity ambiguity) and moment tensor decomposition (see also IS 3.9 and EX 3.5).

The second inversion step is always performed in the time domain, to resolve the polarity ambiguity and to search for the preferred centroid locations. Previous results are improved through the solution of the polarity ambiguity. Additionally, the relative epicentral location and time of the centroid are found. Location and times are defined with respect to the hypocentral location (in terms of spatial shift along North/South and East/West) and nucleation time (in terms of relative time). Both these reference parameters are defined by the user prior to the inversion, and are typically based upon a specific source location result or seismic catalogue information.

After completion of the first and the second inversion step we leave the framework of point source inversion and try to invert for kinematic parameters of the extended source. The background for doing so and the source models used to keep the inversion problem feasible is given in Heimann (2011) and Cesca et al.(2010). The third inversion step for kinematic parameters of the extended source is preferably performed in the frequency domain, including higher frequencies, which should help to estimate finite source parameters. Among the possible resolved parameters are: the preferred rupture plane orientation (discrimination between rupture and auxiliary plane), rupture directivity, rupture velocity, size and duration.

The inversion is preferably performed in the WORK directory, where all local python (<http://www.python.org>) codes (rapidinv, Cesca et al. 2010; mopad, Krieger & Heimann 2013) are locally stored.

The following command gives a list of the initial directory content:

```
ls WORK/*
```

\*.py and \*.pyc are locally stored python scripts. rapidinv\* (\* denotes different version which may be available at different stages) is the script to perform the source inversion. mopad is a tool to perform moment tensor decomposition and plotting (see Krieger and Heimann 2013), which is internally used for these purposes.

*rapidinv.inp.\** are example of input files, for different events or inversion setups;  
*rapidinv.defaults* includes the default setup of inversion variables;  
*rapidinv.acceptables* is a control file to verify very roughly the inversion variables choice.

## 5 Selection of inversion parameters

The inversion is customized after building an own input file. To have an idea, how such an input file looks like, we can open with an editor the file *rapidinv.inp.synth1*. It can be seen that the file includes a list of variables (first column) and a list of values (second column). The most important variables are described in the following. A complete list of variable definition and accepted values is given in the rapidinv manual in the VirtualBox in the directory DOCS.

The input file should be kept as short as possible, and only contains those variables which are specific for the specific event inversion. All other variables, those kept to the defaults values and those common for all studied events, should be stored in the file *rapidinv.defaults*.

### 5.1 Important global variables:

DATA\_DIR: Path to the directory containing all data and station information.

DATA\_FORMAT: Expected format of input displacement data (see kinherd.org for more details).

STAT\_INP\_FILE: (Default: stations.dat): Name of the input file with list of stations to be used. This file, with the proper format, should exist and be saved in the data directory (see DATA\_DIR). Specific file format (manual). All stations herein indicated will be considered. If data are missing, the station will be excluded. If data are partially missing, only existing components will be used. If data files exist, but the station name is not included in the list, data will not be used.

INVERSION\_DIR: Path to the directory where all output files will be saved.

NUM\_INV\_STEPS: Number of inversion steps to be realized (1 = only focal mechanism, 2 = focal mechanism and centroid location, 3 = full point and kinematic inversion).

COMP\_2\_USE: The string should include one letter for each spatial component to consider. Letters follow the convention from kinherd.org (d: down; u: up; n: North; s: South; e: East; w: West; r: transversal, rightward as seen from source to receiver; l: transversal, leftward as seen from source to receiver; a: radial, away from source; c: radial, backward to source).

EPIC\_DIST\_MIN, EPIC\_DIST\_MAX: Minimum and maximum epicentral distance to be considered for all steps.

LATITUDE\_NORTH, LONGITUDE\_EAST: Original latitude and longitude (deg) of the epicenter. Further relocation will be relative to this value.

SW\_FILTERNOISY: Switch to detect and remove noisy traces.

SW\_WEIGHT\_DIST: Apply a distance-dependent weight, w:  $w = \text{station\_epicentral\_distance} / \text{maximal\_epic\_distance}$ .

SW\_FULLMT: Switch to additionally run full moment tensor inversion

### 5.2 Important variables for inversion step 1:

BP\_F1\_STEP1, BP\_F2\_STEP1, BP\_F3\_STEP1, BP\_F4\_STEP1: Bandpass frequency (e.g. 0.009 0.01 0.1 0.11) at inversion step 1.

DEPTH\_BOTTOMLIM, DEPTH\_UPPERLIM: Minimum and maximum accepted depth (km) after inv. step 1 (all other solutions will be removed)

DEPTH\_1, DEPTH\_2, DEPTH\_STEP; SCAL\_MOM\_1, SCAL\_MOM\_2, SCAL\_MOM\_STEP; STRIKE\_1, STRIKE\_2, STRIKE\_STEP; DIP\_1, DIP\_2, DIP\_STEP; RAKE\_1, RAKE\_2, RAKE\_STEP: Min, Max, Increment of depth (km), moment (Nm) and fault angles (deg) to define starting configurations, during inversion step 1.

SW\_RAPIDSTEP1: If set TRUE will ignore STRIKE\_1, STRIKE\_2, ... and test 10 default configurations of strike-dip-rake.

INV\_MODE\_STEP1: Defines the strategy to carry out inversion step 1. The following possibilities are allowed (e.g. invert\_dmsdst: gradient inversion of invert strike, dip, rake, moment, depth)

MISFIT\_MET\_STEP1: Defines the misfit definition for inversion step 1 (ampspec\_l1norm: amplitude spectra will be compared, using L1 norm; ampspec\_l2norm: amplitude spectra will be compared, using L2 norm; l1norm: time traces will be compared, using L1 norm; l2norm: time traces will be compared, using L2 norm).

PHASES\_TO\_USE\_ST1: Seismic phases to be used for inv. Step 1 (p: P phases on all used components, as defined in COMP\_2\_USE; b: Bodywaves, P on vertical component, S on remaining components; a: Full waveform on all used components, as defined in COMP\_2\_USE).

WEIGHT\_A\_ST1, WEIGHT\_P\_ST1, ...: Defines a weight for the time window for phase a (p, ...) during inversion step 1. It is used only if the phase is activated in PHASES\_TO\_USE\_ST1.

WIN\_LENGTH\_A\_ST1, WIN\_LENGTH\_P\_ST1, ...: Defines the length (s) of the time window for phase a (p, ...) during inversion step 1.

WIN\_START\_A\_ST1, WIN\_START\_P\_ST1, ...: Defines the starting time of the time window for phase a (p, ...) during inversion step 1. The value indicates the position of theoretical first P (or S) phase within the time window, defined as percentage of the window length (e.g. 0.02 for a window length of 100s, indicates that the window will start to have the theoretical first P phase arrival time at 2% of the window length).

WIN\_TAPER\_A\_ST1, WIN\_TAPER\_P\_ST1, ...: Defines the tapering of the time window for phase a (p, ...) during inversion step 1. A taper is applied in the time domain, smoothing the time window at both sides. The length of the smoothed part of each side is defined by the given real value, which is defined as a percentage of the time window length.

### 5.3 Plotting variables for inversion step 1:

DATA\_PLOT\_STEP1: Define if data format for plotting result after inversion step 1 is amplitude spectra or displacements). Acceptable values are amsp or seis.

AMPL\_PLOT\_STEP1: Amplitudes of plotted spectra/waveforms (amax: amplitudes of different traces have the same scale; norm: amplitudes are normalized, they are enlarged to the maximal possible scale).

FILT\_PLOT\_STEP1 Defines the format (plain = unfiltered, filtered, tapered) of time series for plotting results after inversion step 1.

START\_PLOT\_STEP1, LEN\_PLOT\_STEP1, TICK\_PLOT\_STEP1: Defines the start, length and tick interval (s) of time series for plotting results after inversion step 1. These are used only if DATA\_PLOT\_STEP1 is seis.

#### 5.4 Important (and plotting) variables for inversion step 2:

EPIC\_DIST\_MAXLOC: Additional constraint on maximum epicentral distance to consider for inversion step 2 (where maximum distance will be the minimum between EPIC\_DIST\_MAX and EPIC\_DIST\_MAXLOC).

GFDB\_STEP2: GFDB used to calculate synthetic seismogram during inversion step 2.

MISFIT\_MET\_STEP2: Defines the misfit definition for inversion step 1 (l1norm: time traces will be compared, using L1 norm; l2norm: time traces will be compared, using L2 norm)

REL\_EAST\_1, REL\_EAST\_2, REL\_EAST\_STEP; REL\_NORTH\_1, REL\_NORTH\_2, REL\_NORTH\_STEP: Min, max and increment of relative location along East and North direction (meters) to define the group of starting configuration, during inversion step 2.

REL\_TIME1, REL\_TIME\_2, REL\_TIME\_STEP: Min, max and increment of relative time offset (seconds) to define the group of starting configuration, during inversion step 2.

PHASES\_TO\_USE\_ST2; WEIGHT\_\*\_ST2, WIN\_LENGTH\_\*\_ST2, WIN\_START\_\*\_ST2, WIN\_TAPER\_\*\_ST2: Analogous to the case as for inversion step 1.

BP\_F1\_STEP2, BP\_F2\_STEP2, BP\_F3\_STEP2, BP\_F4\_STEP2: Bandpass inversion step 2.

DATA\_PLOT\_STEP2: Define if data format (amsp, seis) for plotting result after inversion step 2 (amplitude spectra/displacements)

#### 5.5 Important (and plotting) variables for inversion step 3:

EPIC\_DIST\_MAXKIN: Additional constraint on maximum epicentral distance to consider for inversion step 3 (where maximum distance will be the minimum between EPIC\_DIST\_MAX and EPIC\_DIST\_MAXKIN).

GFDB\_STEP3: Defines the GFDB used to calculate synthetic seismogram during the inversion step 3.

MISFIT\_MET\_STEP3: Defines the misfit definition for inversion step 3 (ampspec\_l1norm: amplitude spectra will be compared, using L1 norm; ampspec\_l2norm: amplitude spectra will be compared, using L2 norm; l1norm: time traces will be compared, using L1 norm; l2norm: time traces will be compared, using L2 norm).

REL RUPT\_VEL\_1, REL RUPT\_VEL\_2, REL RUPT\_VEL\_S: Min, max, increment of relative rupture velocity for starting configurations of inv. Step 3.

PHASES\_TO\_USE\_ST3; WEIGHT\_\*\_ST3, WIN\_LENGTH\_\*\_ST3, WIN\_START\_\*\_ST3, WIN\_TAPER\_\*\_ST3: Analogous as for inversion step 1 (use of bodywaves may work better for location)

BP\_F1\_STEP3, BP\_F2\_STEP3, BP\_F3\_STEP3, BP\_F4\_STEP3: Bandpass inversion step 3.

DATA\_PLOT\_STEP3: Define if data format (amsp, seis) for plotting result after inversion step 2 (amplitude spectra/displacements)

SW\_APPDURATION: Activate fast directivity inversion (uses BP and taper from inversion step 3)

Once all parameters are modified we are ready to perform the inversion

## 6 Moment tensor inversion step by step

The inversion procedure is here described for the case of the SYNTH1 event, according to the original rapidinv.inp.synth1 inversion setup.

Go to the working directory and open the chosen input file.

Check DATA\_DIR points to the correct data directory

Check GFDB\_STEP1, GFDB\_STEP2, GFDB\_STEP3 point to the correct database

Adapt inversion parameters to user wishes.

Check that specific relevant parameters are fixed, in particular those defining the source parameters which are considered to be known, when starting the moment tensor inversion, This include the event origin time, its hypocentral location, and a first moment estimate. Both depth and moment will be . This information is typically available from global catalogue, such as NEIC or GCMT for earthquakes above Mw 5.5, or from local agencies in the case of smaller magnitude events of regional interest. These parameters are provided for all test datasets. For synthetic dataset, the true source parameters are given, while for real dataset, source parameters were chosen according to the information provided either by NEIC, GCMT or EMSC-CSEM.

Known source parameters for the event SYNTH1 are as follows:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2012-03-15 15:00:00 |
| Lat N                  | 47.6667             |
| Lon E                  | 12.0167             |
| Depth (rough)          | 15km                |
| Mw (rough)             | 6.0                 |
| M <sub>0</sub> (rough) | 1e18Nm              |

We are now ready to run the inversion. This is done by typing in the working directory:

```
python rapidinv12.py rapidinv.inp.synth1
```

The code first read rapidinv.defaults and assign default values to all variables.

Then it reads the input file and overwrites all selected variables (repeated variable assignment imply the last assignment will overwrite previous ones)

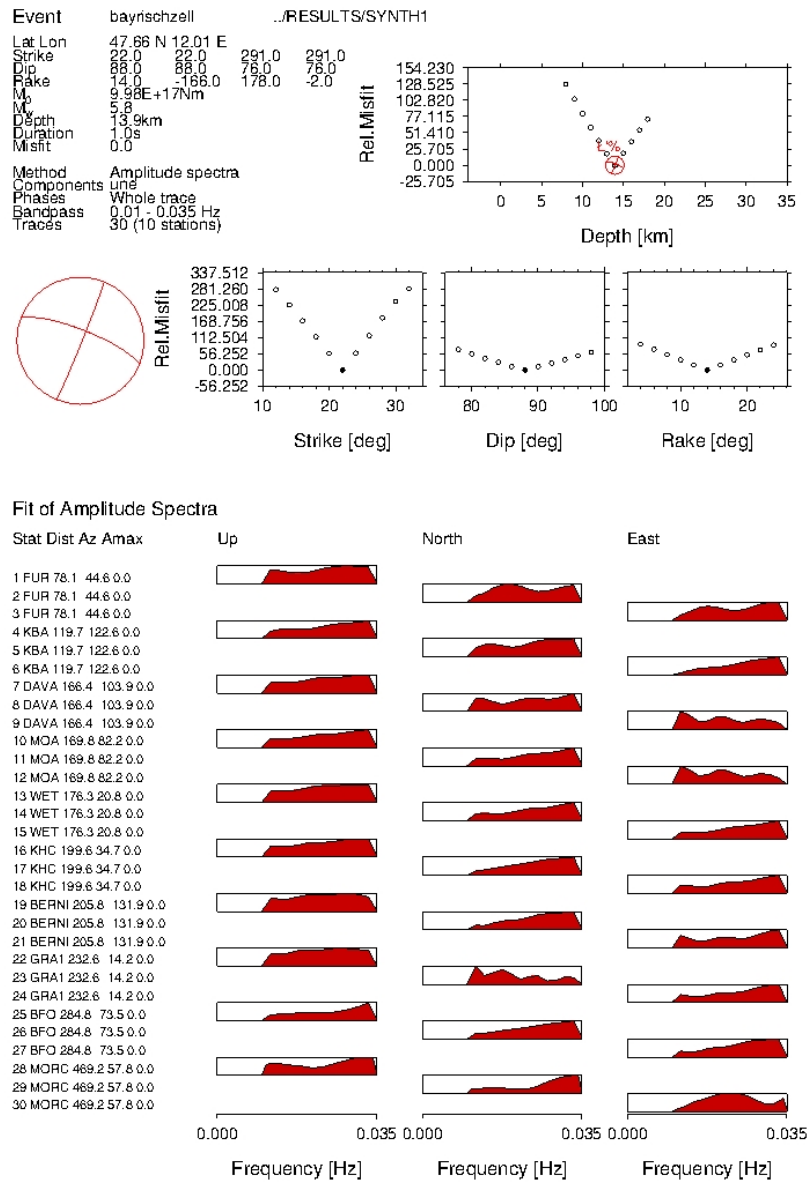
Finally, a rough control on variable values is performed using the file rapidinv.acceptables

Then, the following tasks are performed:

- Create/overwrite results directory;
- Filter stations according to selected epicentral distance ranges;
- A number of starting source configuration for the DC inversion is built, according to input parameters, and the DC inversion is then performed.
- The best DC solution, together with a number of non-DC solutions, constitute the starting source configurations for the full MT inversion, and the MT inversion is then performed.
- The first inversion step is successfully completed when two GMT scripts are created, executed and two postscript files generated: step1.ptsolution.ps and step1.mtsolution.ps

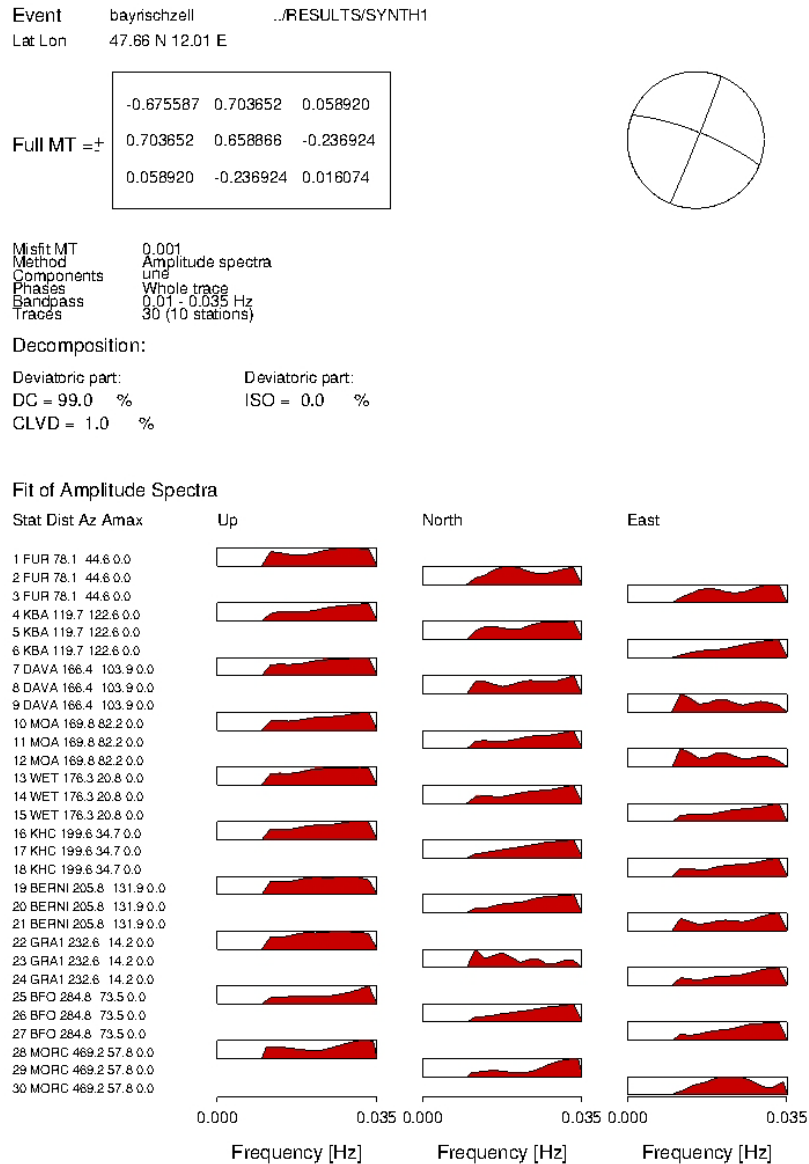
*step1.ptsolution.ps* (see Figure 1, obtained for event SYNTH1) shows the results of the pure DC inversion after inversion step 1. Source parameters, including location, depth, moment, magnitude, strike-dip-rake configurations and misfit are given in the top left section of the figure. The top right section is dedicated to the relative misfit variations with respect to the best solution, when perturbing single source parameters (depth, strike, dip, rake). These curves are expected to have a minimum in correspondence of the preferred solution, the variability of the minima broadness provides information on the relative resolution of different source parameters (e.g. strike vs. dip vs. rake).

The bottom panel of *step1.ptsolution.ps* is dedicated to waveform or spectra comparison (red colors are used for data, black for synthetics). Waveform plots additionally show a gray area denoting the applied time domain taper. Stations are sorted on the base of epicentral distances, with station name, distance, azimuth and maximal amplitudes provided on the left side.



**Figure 1** *step1.ptsolution.ps*

*step1.mtsolution.ps* (figure 2, obtained for event SYNTH1) provides a similar representation for the full MT inversion results (for the definition of the full MT parameter set (centroid location and time, CLVD, isotropic components etc. see IS 3.9):



**Figure 2** *step1.mtsolution.ps*

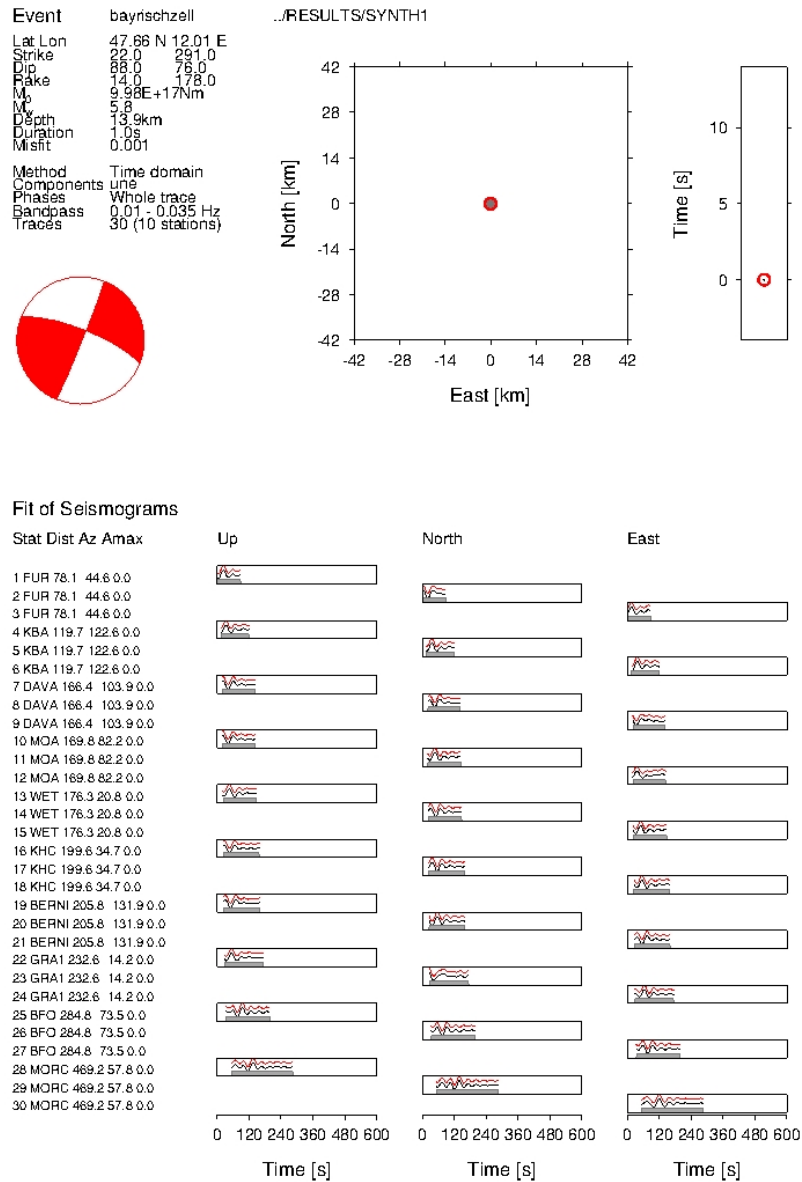
The second inversion step is carried out in the time domain, by comparing observed and synthetic displacement waveforms, for a reduced set of focal mechanisms, decided upon results of inversion step 1. The inversion step 2 for DC solution is performed as follows:

- Two possible DC focal mechanisms, with opposite polarities are used to build synthetics. After waveform cross-correlation, waveforms fits are evaluated and the polarity chosen.
- A grid search for the centroid location and centroid time (as relative location and time



with respect to the original epicentral location and origin time respectively) is performed, and best centroid location/time found.

- Polarities are also tested for the best location also for the best full MT solution, and the polarity chosen. A full MT decomposition, through DC, compensated linear vector dipole (CLVD) and isotropic components is then performed.
- The second inversion step is successfully completed when two GMT scripts are created, executed and two postscript files generated: *step2.ptsolution.ps* (figure 3) and *step2.mtsolution.ps*

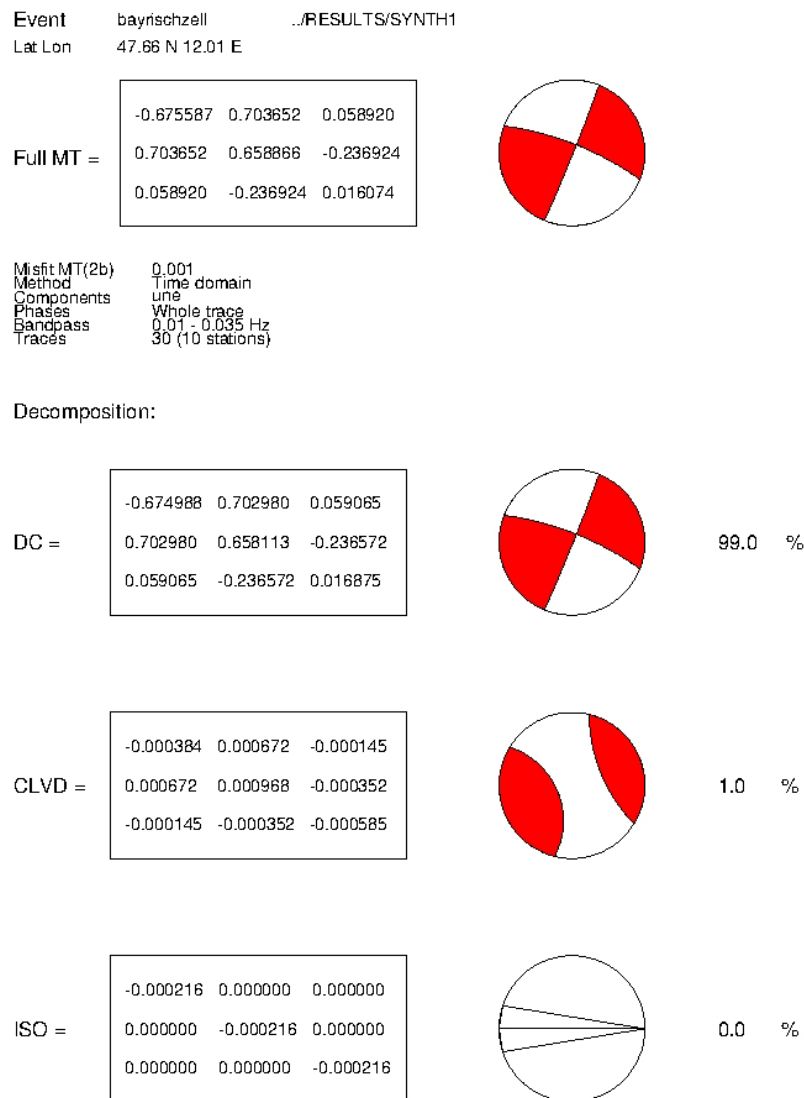


**Figure 3** *step2.ptsolution.ps*

*step2.ptsolution.ps* shows the results of the pure DC inversion after inversion step 2. Improved source parameters are given in the top left section of the figure. The top right section is dedicated to the relative misfit variations for the centroid location and centroid time

grid search (reference location is at 0, 0; reference time at 0). Larger circle sizes correspond to lower misfits. Red circles denote best solutions. The bottom panel is again dedicated to waveform (or spectra) comparison (red colors are used for data, black for synthetics). Waveform plots additionally show a gray area denoting the applied time domain taper. Stations are sorted on the base of epicentral distances, with station name, distance, azimuth and maximal amplitudes provided on the left side.

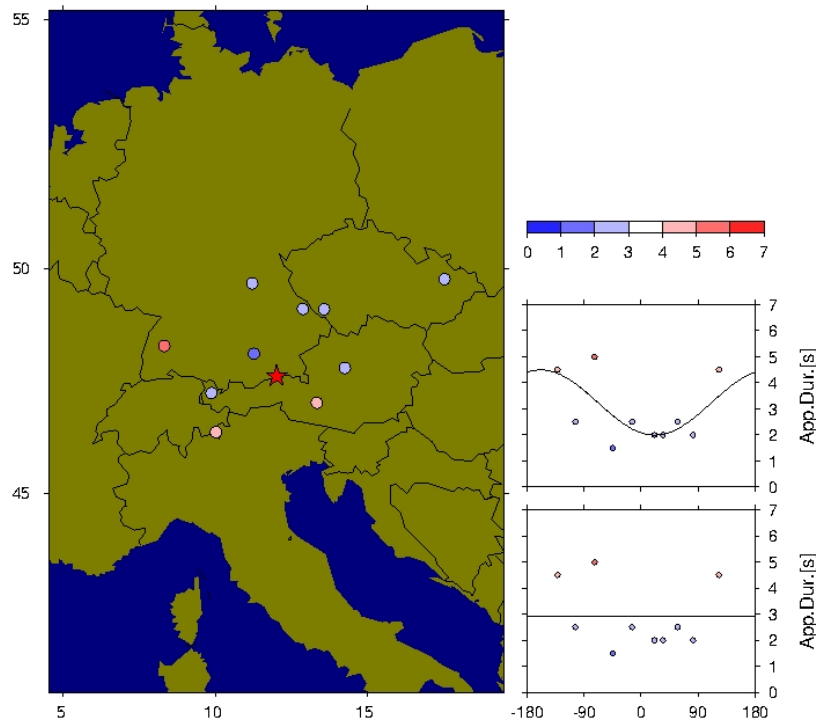
The moment tensor polarity is similarly obtained, by comparing waveform fits for both full MT solution with opposite polarity. The resulting plot (step2.mtsolution.ps, shown in figure 4 for event SYNTH1) illustrates the result of moment tensor decomposition, according to DC, compensated linear vector dipole (CLVD) and isotropic (ISO) components.



**Figure 4** *step2.mtsolution.ps*

Prior to the complete finite source inversion, a quick directivity inversion is carried out (see Cesca et al. 2011 for details), by inverting the apparent source duration at single stations, and plot its dependance on the station azimuth. Results can be seen in the file *apparentdurations.ps*. This file shows the results of the quick directivity detection analysis. The map on the right side of the plot illustrate the study region, the epicenter (star), and the station locations (circles). Circle colors denote the apparent duration seen at that station, according to the color bar. The azimuthal distribution of the apparent durations are shown on the left graphs, and fitted through a preferred directivity model (more details on expected curve shapes are discussed in Cesca *et al.*, 2011), or through a common, average source time function length (bottom).

Figure 5 shows an example of the file *apparentduration.ps* for event SYNTH8 where rupture directivity towards NNE is correctly resolved.



**Figure 5** *apparentdurations.ps*

- A range of tested found source solutions is built, with two possible fault plane orientations, variable size and directivity (currently hardcoded, and link to the event magnitude), and possibly variable rupture velocity.
- Finally, performs the third inversion step, which is carried out by grid search, evaluating the misfits for all considered finite source solutions. This last step is successfully completed when a GMT script is created, executed and a postscript file generate: *step3.eiksolution.ps*

*step3.eiksolution.ps* (see figure 6, obtained for event SYNTH8) shows the results of the finite source inversion after inversion step 3. Point and finite source parameters (e.g. Fault plane orientation, radius, area, average slip, etc.) are given in the top left section of the figure. The top right section is dedicated to the rupture propagation (color denote times of excitation,

lines correspond to rupture isochrones) along the rupture plane (x axis: along strike; y axis: along dip) and relative misfit variations for solutions of different radius and relative rupture velocity).

The bottom panel is again dedicated to waveform (or spectra) comparison (red colors are used for data, black for synthetics). Waveform plots additionally show a gray area denoting the applied time domain taper. Stations are sorted on the base of epicentral distances, with station name, distance, azimuth and maximal amplitudes provided on the left side.

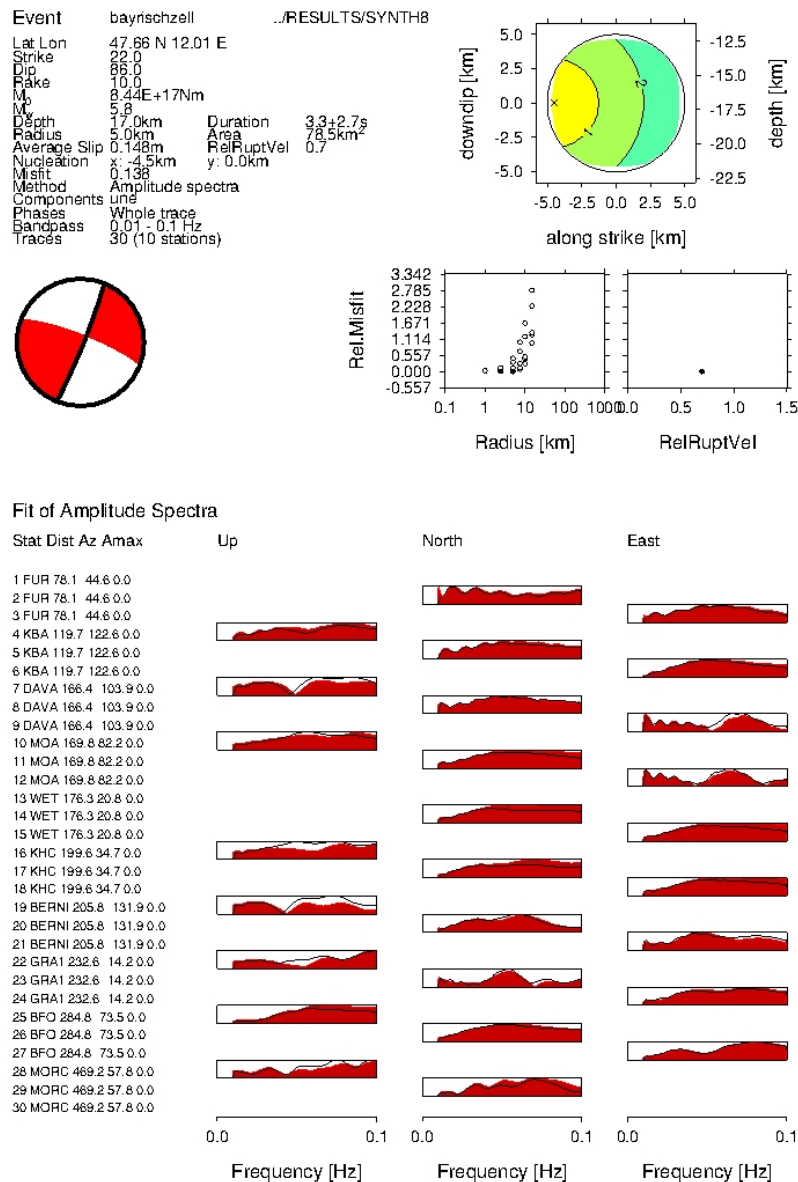


Figure 6 *step3.eiksolution.ps*

## 7 Testing cases/exercises

Data, including displacement waveforms and metadata, are provided for 20 earthquakes: 8 of them are synthetic events, the remaining 12 are real earthquakes. Because of the spatial

limitations of the here presented available GFDB all events are shallow (crustal) earthquakes, with an hypocentral depth between 1 and ca. 30km. For the same reason only data for stations at local/regional distances up to a maximum of 1000 km are available. Magnitudes of the events range between Mw 4 and 7.

Real datasets have been selected in order to illustrate possible inversion targets, such as the inversion of pure double couple solutions (moderate magnitude natural earthquakes), full moment tensor solutions (mine collapse induced event), and finite source inversion (larger magnitude earthquakes).

The following events (and directories) are available for the reader to test:

| <i><b>Directory</b></i> | <i><b>Event name</b></i> | <i><b>Dataset short description</b></i>             |
|-------------------------|--------------------------|-----------------------------------------------------|
| SYNTH1                  | Bayrischzell_D           | Synthetic data (DC point source)                    |
| SYNTH2                  | Bayrischzell_D           | Synthetic data (MT point source)                    |
| SYNTH3                  | Bayrischzell_D           | Synthetic data (kinematic bilateral model)          |
| SYNTH4                  | Bayrischzell_D           | Synthetic data (kinematic unilateral model)         |
| SYNTH5                  | Bayrischzell_D           | Synthetic data, wrong model (DC point source)       |
| SYNTH6                  | Bayrischzell_D           | Synthetic data, wrong model (MT point source)       |
| SYNTH7                  | Bayrischzell_D           | Synthetic data, wrong model (kinematic bil. model)  |
| SYNTH8                  | Bayrischzell_D           | Synthetic data, wrong model (kinematic unil. model) |
| GER110996A              | Teutschenthal_D          | Real data, Local distances, Potash mine collapse    |
| GER220203A              | St_Die_Vosges_F          | Real data, Local distances                          |
| GER201004A              | Rotenburg_D              | Real data, Local distances, Reservoir induced       |
| GER051204A              | Waldkirch_D              | Real data, Local distances                          |
| GRE110406A              | Zakynthos1_GR            | Real data, Regional distances                       |
| GRE110406B              | Zakynthos2_GR            | Real data, Regional distances                       |
| GRE120406A              | Zakynthos3_GR            | Real data, Regional distances                       |
| GRE080608A              | Andravida_GR             | Real data, Regional distances, Directivity          |
| ITA060409A              | L'Aquila_I               | Real data, Regional distances                       |
| ITA250112A              | Emilia_I                 | Real data, Local distances, Dense network           |
| TUR231011A<br>2011      | Van1_TR                  | Real data, Regional distances, main event Van       |
| TUR091111A<br>2011      | Van2_TR                  | Real data, Regional distances, aftershock Van       |

Details for each event are provided on the following pages. Together with the event reference name and a short description, the reader should make use of the “Known source parameters”, which are the only needed parameters to build a proper input file for the inversion.

### **SYNTH1**

Event Name: Bayrischzell, Germany Description:

Description:

Synthetic dataset for local inversion. The IASP91 GFDB was used to create synthetic seismograms and no noise is included. Seismograms are obtained for a pure DC point source model.

Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2012-03-15 15:00:00 |
| Lat N                  | 47.6667             |
| Lon E                  | 12.0167             |
| Depth (rough)          | 15km                |
| M <sub>w</sub> (rough) | 6.0                 |
| M <sub>0</sub> (rough) | 1e18Nm              |

Network configuration:

|                   |             |
|-------------------|-------------|
| Stations number   | 10          |
| Station distances | up to 500km |

## **SYNTH2**

Event Name: Bayrischzell, Germany

Description:

The IASP91 GFDB was used to create synthetic seismograms and no noise is included. Seismograms are obtained for a full MT point source model (DC term similar to SYNTH1).

Known source parameters and network configuration: see SYNTH1

## **SYNTH3**

Event Name: Bayrischzell, Germany

Description:

The IASP91 GFDB was used to create synthetic seismograms and no noise is included. Seismograms are obtained for one extended source model (DC similar to SYNTH1).

Known source parameters and network configuration: see SYNTH1

## **SYNTH4**

Event Name: Bayrischzell, Germany

Description:

The IASP91 GFDB was used to create synthetic seismograms and no noise is included. Seismograms are obtained for a different extended source model (DC similar to SYNTH1).

Known source parameters and network configuration: see SYNTH1

## **SYNTH5**

Event Name: Bayrischzell, Germany

Description:

A different crustal model was used to create synthetic seismograms and no noise is included. Seismograms are obtained for a pure DC point source model (same as SYNTH1).

Known source parameters and network configuration: see SYNTH1

## **SYNTH6**

Event Name: Bayrischzell, Germany

Description:

A different crustal model was used to create synthetic seismograms and no noise is included. Seismograms are obtained for a full MT point source model (same as SYNTH2).

Known source parameters and network configuration: see SYNTH1

### **SYNTH7**

Event Name: Bayrischzell, Germany

Description:

A different crustal model was used to create synthetic seismograms and no noise is included. Seismograms are obtained for one extended source model (same as SYNTH3).

Known source parameters and network configuration: see SYNTH1

### **SYNTH8**

Event Name: Bayrischzell, Germany

Description:

The IASP91 dataset was used to create synthetic seismograms and no noise is included. Seismograms are obtained for a different extended source model (same as SYNTH4).

Known source parameters and network configuration: see SYNTH1

### **GER220203A**

Event Name: St. Die, Vosgian\_Mts, France

Description:

Real data (GRSN, MEDNET, GEOSCOPE networks).

Local earthquake with good quality data.

Point source (DC and MT) inversion should be well constrained and easily retrieved.

Kinematic models are not easily resolved, because the rupture model is likely very small for such moderate event.

Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2003-02-22 20:41:01 |
| Lat N                  | 48.33               |
| Lon E                  | 6.67                |
| Depth (rough)          | 12km                |
| Mw (rough)             | 4.5                 |
| M <sub>0</sub> (rough) | 1e16Nm              |

Network configuration:

|                   |             |
|-------------------|-------------|
| Stations number   | 17          |
| Station distances | up to 500km |

### **GER051204A**

Event Name: Waldkirch, Black Forest, Germany

Description:

Real data (GRSN networks).

Local earthquake with good quality data, but poor azimuthal coverage.

Point source (DC and MT) inversion should be well constrained and easily retrieved.

Kinematic models are not easily resolved, because the rupture model is likely very small for such moderate event.

Known source parameters:

|               |                     |
|---------------|---------------------|
| Date/Time     | 2004-12-05 01:52:37 |
| Lat N         | 48.115              |
| Lon E         | 8.077               |
| Depth (rough) | 10km                |
| Mw (rough)    | 4.5                 |

|                        |             |
|------------------------|-------------|
| $M_0$ (rough)          | 1e16Nm      |
| Network configuration: |             |
| Stations number        | 8           |
| Station distances      | up to 500km |

### **GER110996A**

Event Name: Teutschenthal Collapse, Germany

Description:

Real data (GRSN network) from 1996.

Local earthquake with challenging dataset. A complex rupture model is expected, with energy released in different subevents. The collapse of the mine is expected to show relevant implosive components and CLVD.

Point source (DC and MT) inversion should be well constrained and easily retrieved.

Kinematic models are not easily resolved, because the rupture model is very small ( $\leq 1$ km radius).

Known source parameters:

|               |                     |
|---------------|---------------------|
| Date/Time     | 1996-09-11 03:36:35 |
| Lat N         | 51.48               |
| Lon E         | 11.79               |
| Depth         | 1km                 |
| $M_w$ (rough) | 4.5                 |
| $M_0$ (rough) | 1e16Nm              |

Network configuration:

|                   |             |
|-------------------|-------------|
| Stations number   | 10          |
| Station distances | up to 500km |

### **GER201004A**

Event Name: Rotenburg, induced event, Germany

Description:

Real data (GRSN network). Local earthquake.

Point source (DC and MT) inversion should be well constrained and easily retrieved.

Kinematic models are not easily resolved, but perhaps possible, because the rupture velocity was reported as very slow.

Known source parameters:

|               |                     |
|---------------|---------------------|
| Date/Time     | 2004-10-20 06:59:18 |
| Lat N         | 53.01               |
| Lon E         | 9.63                |
| Depth         | 5km                 |
| $M_w$ (rough) | 4.5                 |
| $M_0$ (rough) | 1e16Nm              |

Network configuration:

|                   |             |
|-------------------|-------------|
| Stations number   | 20          |
| Station distances | up to 500km |

### **GRE110406A**

Event Name: Zakynthos\_1, Greece

Description:

Real data (GEOFON, MEDNET, IRIS networks). The earthquake is part of a seismic sequence close to Zakynthos, Greece (other events from the sequence are available).

Regional earthquake.



Point source DC (and MT) and kinematic inversion can be performed.

Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2006-04-11 00:02:40 |
| Lat N                  | 37.64               |
| Lon E                  | 20.75               |
| Depth                  | 20km                |
| M <sub>w</sub> (rough) | 5.5                 |
| M <sub>0</sub> (rough) | 1e17Nm              |

Network configuration:

|                   |              |
|-------------------|--------------|
| Stations number   | 17           |
| Station distances | up to 1000km |

### **GRE110406B**

Event Name: Zakynthos\_2, Greece

Description:

Real data (GEOFON, MEDNET, IRIS networks). The earthquake is part of a seismic sequence close to Zakynthos, Greece (other events from the sequence are available).  
Regional earthquake.

Point source DC (and MT) and kinematic inversion can be performed.

Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2006-04-11 17:29:27 |
| Lat N                  | 37.64               |
| Lon E                  | 20.75               |
| Depth                  | 20km                |
| M <sub>w</sub> (rough) | 5.5                 |
| M <sub>0</sub> (rough) | 1e17Nm              |

Network configuration:

|                   |              |
|-------------------|--------------|
| Stations number   | 15           |
| Station distances | up to 1000km |

### **GRE120406A**

Event Name: Zakynthos\_3, Greece

Description:

Real data (GEOFON, MEDNET, IRIS networks). The earthquake is part of a seismic sequence close to Zakynthos, Greece (other events from the sequence are available).  
Regional earthquake.

Point source DC (and MT) and kinematic inversion can be performed.

Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2006-04-12 16:51:59 |
| Lat N                  | 37.63               |
| Lon E                  | 20.74               |
| Depth                  | 20km                |
| M <sub>w</sub> (rough) | 5.5                 |
| M <sub>0</sub> (rough) | 1e17Nm              |

Network configuration:

|                   |              |
|-------------------|--------------|
| Stations number   | 17           |
| Station distances | up to 1000km |

### **GRE080608A**

Event Name: Andravida, Greece

Description:

Real data (GEOFON, MEDNET, IRIS networks) with very good quality.  
Magnitude 6.3 earthquake in NW Peloponnese, Greece, with stations at regional distances.  
Point source DC (and MT) and kinematic inversion can be performed.  
Directivity effects can be clearly seen.

Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2008-06-08 12:25:27 |
| Lat N                  | 37.93               |
| Lon E                  | 21.63               |
| Depth                  | 25km                |
| M <sub>w</sub> (rough) | 6.3                 |
| M <sub>0</sub> (rough) | 4e18Nm              |

Network configuration:

|                   |              |
|-------------------|--------------|
| Stations number   | 20           |
| Station distances | up to 1000km |

**ITA250112A**

Event Name: Emilia, Italy

Description:

Real data from different networks (MEDNET, IRIS, GEOFON, INGV, Univ.Genova).  
The event is relatively weak but the stations distribution unusually dense (inversion may be time consuming if all stations are used).  
Stations are at local distances, most below 200km.  
Point source (DC and MT) inversion should be well constrained and easily retrieved.  
Kinematic models are not easily resolved, because of the small earthquake size/magnitude.

Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2012-01-25 08:06:36 |
| Lat N                  | 44.85               |
| Lon E                  | 10.54               |
| Depth                  | 20km                |
| M <sub>w</sub> (rough) | 4.5                 |
| M <sub>0</sub> (rough) | 1e16Nm              |

Network configuration:

|                   |             |
|-------------------|-------------|
| Stations number   | 38          |
| Station distances | up to 300km |

**ITA060409A**

Event Name: L'Aquila, Italy

Description:

A well-known earthquake for (Italian) seismologists.  
Real data from different networks (MEDNET, IRIS, GEOFON, GEOSCOPE, GRSN) are available at regional distances.  
Point source (DC and MT) and kinematic inversions can be tested.

Known source parameters:

|           |                     |
|-----------|---------------------|
| Date/Time | 2009-04-06 01:32:41 |
| Lat N     | 42.33               |
| Lon E     | 13.34               |
| Depth     | 5km                 |

|                        |              |
|------------------------|--------------|
| M <sub>w</sub> (rough) | 6.2          |
| M <sub>0</sub> (rough) | 3e18Nm       |
| Network configuration: |              |
| Stations number        | 15           |
| Station distances      | up to 1000km |

### **TUR231011A**

Event Name: Van, Turkey

#### Description:

The largest event in the dataset. This event has recently affected Eastern Turkey, the epicenter close to the city of Van.

Real data from different networks are available at regional distances.

Station distribution is poor and uneven.

Point source (DC and MT) and kinematic inversions can be tested.

#### Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2011-10-23 10:41:22 |
| Lat N                  | 38.67               |
| Lon E                  | 43.58               |
| Depth                  | 5km                 |
| M <sub>w</sub> (rough) | 7.0                 |
| M <sub>0</sub> (rough) | 5e19Nm              |

#### Network configuration:

|                   |              |
|-------------------|--------------|
| Stations number   | 5            |
| Station distances | up to 1000km |

### **TUR091111A**

Event Name: Van, Turkey

#### Description:

The strongest aftershock of the 23.10.2011 earthquake (see TUR231011A).

Real data from different networks are available at regional distances.

Station distribution is poor and uneven. The azimuthal gap is very large.

Point source (DC and MT) and kinematic inversions can be tested.

#### Known source parameters:

|                        |                     |
|------------------------|---------------------|
| Date/Time              | 2011-11-09 19:23:35 |
| Lat N                  | 38.35               |
| Lon E                  | 43.40               |
| Depth                  | 20km                |
| M <sub>w</sub> (rough) | 5.7                 |
| M <sub>0</sub> (rough) | 5e17Nm              |

#### Network configuration:

|                   |              |
|-------------------|--------------|
| Stations number   | 6            |
| Station distances | up to 1000km |

## **Acknowledgment**

The facilities of GEOFON and IRIS Data Management System, and specifically the IRIS Data Management Center, were used to access part of the waveform and metadata required in this study. Part of data and metadata relative to earthquakes in Germany and surrounding areas were provided by BGR in Germany. We acknowledge all institutions providing seismic

data used in this research. Figures have been prepared with GMT (Wessel and Smith 1998) and MoPaD (Krieger and Heimann 2013).

## References

- Cesca, S., Heimann, S., Stammer, K., and Dahm, T. (2010). Automated procedure for point and kinematic source inversion at regional distances. *J. Geophys. Res.*, . **115**, B6304; doi: 10.1029/2009JB006450
- Cesca, S., Heimann, S., and Dahm, T. (2011). Rapid directivity detection by azimuthal spectra inversion. *J. Seismol.*, **15** (1), 147-164; doi: 10.1007/s10950-010-9217-4
- Cesca, S., Rohr, A., and Dahm, T., (2013). Discrimination of induced seismicity by full moment tensor inversion and decomposition. *J. Seismol.*, **17** (1), 147-163; doi: 10.1007/s10950-012-9305-8
- Friedrich, W., and Dalkolmo, J. (1995). Complete synthetic seismograms for a spherically symmetric earth by a numerical computation of the green's function in the frequency domain, *Geophys. J. Int.*, **122**, 537-550.
- Heimann, S., 2011. A Robust Method to estimate kinematic earthquake source parameters. PhD thesis, University of Hamburg, Hamburg, Germany. <http://ediss.sub.uni-hamburg.de/volltexte/2011/5357/>
- Kennett, B. L. N., and Engdahl, E. R. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, **105**, 429-465; [doi:10.1111/j.1365-246X.1991.tb06724.x](https://doi.org/10.1111/j.1365-246X.1991.tb06724.x).
- Krieger, L., and Heimann, S., 2013. MoPaD – Moment Tensor Plotting and Decomposition: a Tool for Graphical and Numerical Analysis of Seismic Moment Tensors, *Seismol. Res. Lett.*, doi: 10.1785/gssrl.83.3.589
- Scherbaum, F. (2007). *Of poles and zeros; Fundamentals of Digital Seismometry*. 3<sup>rd</sup> revised edition, Springer Verlag, Berlin und Heidelberg.
- Wang, R. (1999). A simple orthonormalization method for the stable and efficient computation of Green's functions. *Bull. Seismol. Soc. Am.*, **89**, 733-741.
- Wessel, P., and Smith, W. H. F. (1998). New, improved version of the generic mapping tools released, *Eos Trans. AGU*, **79**, 579; doi:10.1029/98EO00426.

| Topic   | Theoretical source representation                                                                                                                                                                                                                                                                                                                                                                                      |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Helmut Grosser</b> , and <b>Peter Bormann</b> (both formerly GFZ German Research Centre for Geosciences, Dept. 2, D-14473 Potsdam, Germany);<br>E-mail: <a href="mailto:helmut_grosser@arcor.de">helmut_grosser@arcor.de</a> and <a href="mailto:pb65@gmx.net">pb65@gmx.net</a><br><b>Agustin Udias</b> (formerly Universidad Complutense de Madrid, Departamento de Geofisica y Meteorologia, 28040 Madrid, Spain) |
| Version | April 2003; references added 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_IS_3.1">10.2312/GFZ.NMSOP-2_IS_3.1</a>                                                                                                                                                                                                                                                                                            |

|                                                      | page      |
|------------------------------------------------------|-----------|
| <b>1 Introduction</b>                                | <b>1</b>  |
| <b>2 Continuum mechanics</b>                         | <b>2</b>  |
| <b>3 Kinematic source models</b>                     | <b>6</b>  |
| <b>4 Dynamic source models</b>                       | <b>14</b> |
| <b>5 Energy, moment, dislocation and stress drop</b> | <b>17</b> |
| <b>Acknowledgments</b>                               | <b>19</b> |
| <b>References</b>                                    | <b>20</b> |

## 1 Introduction

In seismology the problem of understanding and describing the seismic source consists in relating observed seismic waves (i.e., seismograms) generated by this source to suitably conceived geometric, kinematic and dynamic parameters of a mechanical source model that represents the physical phenomenon of a brittle fracture in the Earth's lithosphere. Representations of the source are defined by parameters whose number depends on the complexity of the source models (e.g., Aki and Richards, 1980 and 2002; Ben-Menahem and Singh, 1981; Das and Kostrov, 1988; Lay and Wallace, 1995; Udías, 1999 and 2002). In the direct problem, theoretical seismic wave displacements are determined from source models and in the inverse problem parameters of source models are derived from observed wave displacements. In the following we will consider only source models related to earthquakes and explosions (see Chapter 3), volcanic tremors (see Chapter 13) and rock bursts. Here we will not discuss sources of seismic noise (see Chapter 4).

Strong non-linear and non-elastic processes take place in a seismic source volume. Parts of it may crack, phase transitions may take place, the temperature may increase, and so on. These kinds of processes are not described by most seismic source theories; however, there are special theories to model such processes, e.g., the time-dependent pressure within an explosion cavity, the rupture propagation on an earthquake fault, and the material behavior on a crack tip (crack criteria). We limit ourselves to the phenomenological description of a seismic source. The aforementioned complicated processes need not to be considered when looking only for their integral effect on a surface surrounding the seismic source, i.e., by replacing a volume integral by a surface integral (see, e.g., Aki and Richards, 1980).

## 2 Continuum mechanics

The description of the source mechanism is based on the solution of the equation of motion. In a deformable solid medium this equation is derived from classical Newtonian mechanics. The linearized equation of motion (i.e., by neglecting density changes and other second order effects) is

$$\rho \ddot{u}_i(x_s, t) - \sigma_{ij,j}(x_s, t) = f_i^b(x_s, t) . \quad (1)$$

In this equation  $\rho$  is the density of the solid body,  $u_i$  are the components ( $i = 1, 2, 3$ ) of the displacement field that describe the deformation of the body,  $\sigma_{ik}$  is the stress tensor,  $f_i^b$  is the body force density acting per unit volume,  $\ddot{u}_i$  is the second time derivative  $\partial^2/\partial t^2$  of the displacement and the comma between two subscripts, e.g., in  $\sigma_{ik,k}$  indicates the spatial derivative of the considered quantity. We generally use the *summation convention* which requires that one has to sum when a subscript appears twice, e.g.,

$$\sigma_{ik,k} = \frac{\partial}{\partial x_1} \sigma_{i1} + \frac{\partial}{\partial x_2} \sigma_{i2} + \frac{\partial}{\partial x_3} \sigma_{i3} .$$

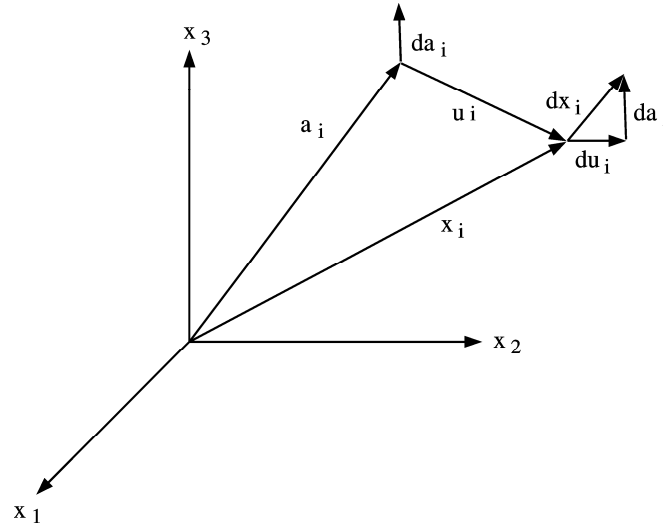
The displacement is a function of the spatial co-ordinates  $x_i$  and the infinitesimal deformation is defined as

$$du_i = u_{i,k} dx_k \quad (2)$$

with

$$u_{i,k} = \beta_{ik} \quad (3)$$

as the distortion tensor. We now consider the location of a particle before and after it is deformed, described by the vectors  $a_i$  and  $x_i$ , respectively. Accordingly, an infinitesimal vector  $da_i$  at the point  $a_i$  is moved (i.e., deformed) to the vector  $dx_i$ , as shown in Figure 1.



**Figure 1** Coordinates and vectors describing the displacement field (see text).

Introducing  $ds^2$ , which is the difference between the square of the length of the vectors  $dx_i$  and  $da_i$ , and, thus, a measure of the deformation of the body, i.e.,  $ds^2 = dx_i dx_i - (dx_i - du_i)(dx_i - du_i)$ , we get with (2) and (3)

$$ds^2 = (\beta_{ij} + \beta_{ji} - \beta_{ki} \beta_{kj}) dx_i dx_j = 2\varepsilon_{ij} dx_i dx_j. \quad (4)$$

Equation (4) is the definition of the *strain tensor*  $\varepsilon_{ij}$ . It is a symmetric tensor. For small deformations it can be approximated by its linear terms

$$\varepsilon_{ij} = \frac{1}{2} (\beta_{ij} + \beta_{ji}) = \frac{1}{2} (u_{i,j} + u_{j,i}). \quad (5)$$

Thus, the strain tensor  $\varepsilon_{ij}$  is the symmetric part of  $\beta_{ij}$ . Any symmetric tensor can be transformed into a co-ordinate system such that

$$\varepsilon_{ij} = \varepsilon^{(i)} \delta_{ij} \quad (6)$$

where  $\delta_{ij}$  is the dimensionless *Kronecker symbol*, defined by

$$\delta_{ij} = \begin{cases} 1 & \text{if } i = j \\ 0 & \text{if } i \neq j \end{cases} \quad (7)$$

and  $\varepsilon^{(i)}$  are the eigenvalues of the strain tensor. The co-ordinate system where  $\varepsilon_{ij}$  is of the form (6) is called system of principle axes. The three eigenvalues describe the relative deformation in direction of the principle axes.

In continuum mechanics one distinguishes between body forces and surface forces. The body forces are sometimes also termed volume forces because they act on volume elements  $dV$  of the body. In Equation (1) we consider infinitesimal masses  $\rho dV$  where  $dV$  is the infinitesimal volume of the mass element. Accordingly, an infinitesimal body force (Aki and Richards, 1980) can be written as  $dF_i = f_i^b(x_s, t) dV$ . Typical examples of body forces are the gravity field and the centrifugal force.

In contrast, surface forces such as cohesion, the sliding friction, or the internal stress during the deformation of the body, act on surface elements  $dS$  of the volume  $dV$ . The stress is a tensor of second order, i.e., it has two subscripts, because it is characterized both by the orientation of the force and by the orientation of the surface on which the force acts. A second-order tensor has generally 9 independent components which can be written explicitly as

$$\sigma_{ij} = \begin{pmatrix} \sigma_{11} & \sigma_{12} & \sigma_{13} \\ \sigma_{21} & \sigma_{22} & \sigma_{23} \\ \sigma_{31} & \sigma_{32} & \sigma_{33} \end{pmatrix}.$$

In general,  $\sigma_{ij}$  depends on position and time. It acts only between adjacent particles. Because of the conservation law of angular momentum this tensor has to be symmetric, i.e.,

$$\sigma_{ij} = \sigma_{ji}. \quad (8)$$

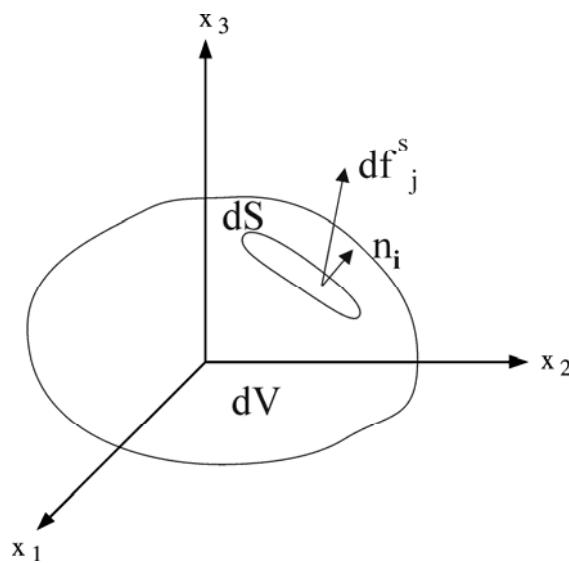
The relation between the incremental body force density  $df_i^s$  which acts on an internal surface element  $dS$  and the stress is

$$df_i^s = \sigma_{ij} n_j dS \quad (9)$$

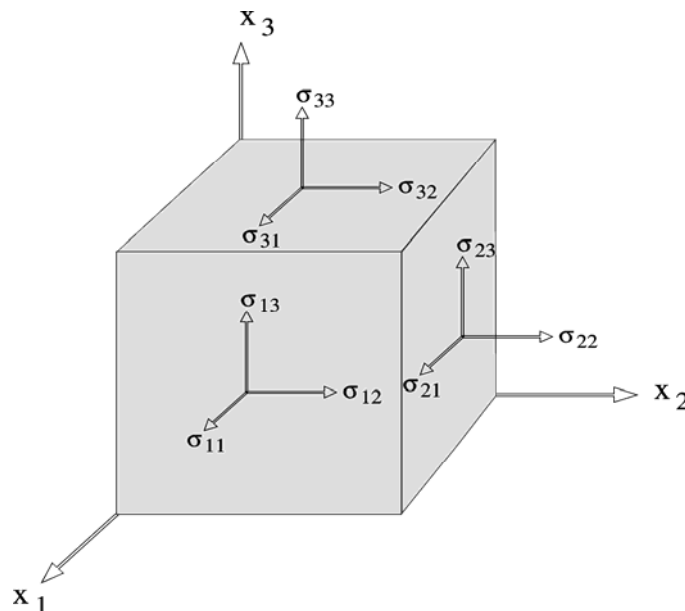
where  $n_j$  is the normal vector of the surface elements (see Figure 2).  $\sigma_{ij} n_j$  is called the *traction* of the stress tensor. The pressure and the surface tension in fluids are special examples of internal surface forces. Figure 3 shows the different components of  $\sigma_{ij}$  which act on the surfaces of an infinitesimal cube.

In the linear theory of elasticity, the strain and the stress tensor are linearly coupled. A relatively simple stress-strain relation is the generalized *Hook's law*

$$\sigma_{ij} = c_{ijkl} \varepsilon_{kl}. \quad (10)$$



**Figure 2** Schematic depiction of the considered source volume  $dV$ , a surface element  $dS$  (with its normal vector  $n_i$ ) on which the force  $df_j^s$  acts.





**Figure 3** The nine components of the stress tensor.  $\sigma_{ij}$  are the components of the stress tensor parallel to  $x_j$  on planes having  $n_i$  as their normals.

The body that obeys the relation (10) is said to be *linearly elastic*. The  $c_{ijkl}$  are called *elastic constants* because they are independent of strain, however, in the case of an inhomogeneous medium, they depend on the position in the body. Due to the symmetry of strain (see Equation (5)) and stress tensor (see Equation (8)) and because of the energy balance in the body, the fourth-order tensor  $c_{ijkl}$  has the following three symmetries:

$$c_{ijkl} = c_{jikl}, \quad c_{ijkl} = c_{ijlk}, \quad \text{and} \quad c_{ijkl} = c_{klij}. \quad (11)$$

These symmetries reduce the independent components in  $c_{ijkl}$  from 81 to 21. In the case of an *isotropic medium*, i.e., when the elastic properties are independent of the orientation in the body, the elastic constants reduce to just two. Then  $c_{ijkl}$  has the form

$$c_{ijkl} = \lambda \delta_{ij} \delta_{kl} + \mu (\delta_{ik} \delta_{jl} + \delta_{il} \delta_{jk}). \quad (12)$$

The two parameter  $\lambda$  and  $\mu$  are known as the *Lamé constants*.

If attenuation has to be included the relatively general *Boltzmann law*

$$\sigma_{ij}(t) = \int_{-\infty}^t b_{ijkl}(t - \tau) \varepsilon_{kl}(\tau) d\tau \quad (13)$$

can be used.

It is advantageous to introduce now the *Fourier transform*  $f(\omega)$  of a time dependent function  $f(t)$ . Here,  $\omega$  is the angular frequency  $2\pi f$ , where  $f$  is frequency in units of Hz.. We use the definitions

$$f(\omega) = \int_{-\infty}^{\infty} f(t) e^{-i\omega t} dt \quad \text{and} \quad f(t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} f(\omega) e^{i\omega t} d\omega \quad (14)$$

where  $i = \sqrt{-1}$  is the imaginary unit, and  $f(\omega)$  is a complex function, called the complex spectrum of  $f(t)$ . It can be represented by

$$f(\omega) = a(\omega) + i b(\omega) = A(\omega) e^{i\Phi(\omega)}$$

where  $A(\omega)$  is the amplitude spectrum and  $\Phi(\omega)$  the phase spectrum.  $a(\omega)$  and  $b(\omega)$  are the real and the imaginary parts of  $f(\omega)$ , respectively. When applying the Fourier transformation to Equation (13) the integral is replaced by the product of  $b_{ijkl}(\omega)$  and  $\varepsilon_{kl}(\omega)$ . The imaginary part of  $b_{ijkl}$  describes a linear attenuation for a propagating displacement field.

With Eqs. (5), (10), and (14) the equation of motion (1) becomes (Udías, 1999)

$$\rho \omega^2 u_i(x_s, \omega) + \sigma_{ij,j}(x_s, \omega) = -f_i^b(x_s, \omega) \quad (15)$$

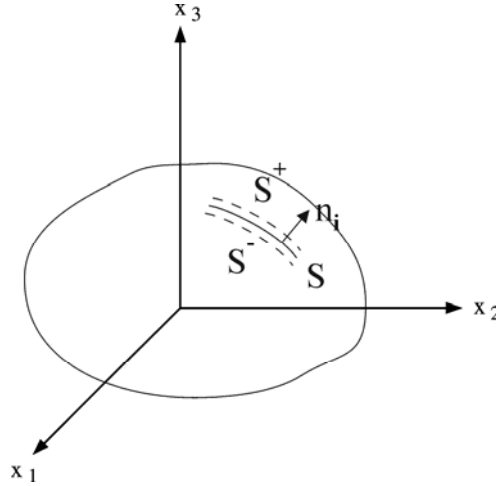
and in a linear elastic but inhomogeneous medium

$$\rho \omega^2 u_i(x_s, \omega) + (c_{ijkl} u_{k,l}(x_s, \omega))_{,j} = -f_i^b(x_s, \omega). \quad (16)$$

The second term on the left side is the stress due to the displacement  $u_k$ . In order to specify  $u_i$  in a unique way, the initial conditions have to be fixed for the displacement  $u_i$  and the related velocity  $\dot{u}_i$  as well as the boundary conditions for the displacement or the traction. The homogeneous initial condition, that both  $u_i$  and  $\dot{u}_i$  are zero before the beginning of the seismic event, is the precondition for the existence of the related Fourier transform  $u_i(x_s, \omega)$ . Boundary conditions can be specified for the displacement  $u_i$  or the traction  $\sigma_{ij} n_j$  on internal surfaces  $S$  (or *external surfaces* such as the Earth's free surface) (see Figure 4), namely

$$u_i(\xi_s, \omega) \quad \text{or} \quad \sigma_{ij}(\xi_s, \omega) n_j \quad \text{on the internal surface } S(\xi_s) \quad (17)$$

where  $S(\xi_s)$  may consist of several unconnected surfaces. The Greek letter  $\xi_s$  used as coordinates should indicate that the quantities  $u_i$  and  $\sigma_{ij}$  are lying on the surface  $S(\xi_s)$  which is generally curved. These boundary conditions are indispensable for modeling seismic sources and computing the wave propagation through a layered medium.



**Figure 4** Illustrating the definition of boundary conditions for seismic faults representation.

### 3 Kinematic source models

The first mathematical formulation of the mechanism of earthquakes used the representation of the processes at the source by a distribution of the body force density  $f_i^b(\xi_s, t)$  acting inside the source volume  $V_0$ . Since these forces must represent the phenomenon of fracture, they are called equivalent forces. If it is assumed that no other body forces are present (gravity, etc.), and that on its surface  $S$  displacements and tractions are zero, we can use the representation theorem in terms of the *Green's function* to write the elastic displacements in an infinite medium in the time domain as

$$u_i(x_s, t) = \int_{-\infty}^{\infty} d\tau \int_{V_0} f_k^b(\xi_s, t) G_{ik}(x_s, t, \xi_s, \tau) dV \quad (18)$$

or in the frequency domain by

$$u_i(x_s, \omega) = \int_{V_0} f_k^b(\xi_s, \omega) G_{ik}(x_s, \xi_s, \omega) dV. \quad (19)$$

The Green's function  $G_{ki}$  is the solution of the equation of motion (16) for special impulsive single point forces, termed *Dirac* or *needle impulses*, which act inside the body. The spectrum of the Dirac impulse is 1 for all frequencies and, thus, does not appear in Equation (20) below. According to Ben-Menahem and Singh (1981) and Udías (1999), the following equation holds for the Green's function

$$\rho \omega^2 G_{in}(x_r, \xi_r, \omega) + (c_{ijkl} G_{kn,l}(x_r, \xi_r, \omega))_{,j} = -\delta_{in} \delta(x_r - \xi_r) \quad (20)$$

where  $\delta(x_r - \xi_r)$  is the three-dimensional Dirac delta function which is the product of three one-dimensional Dirac delta functions, i.e.,  $\delta(x_r - \xi_r) = \delta(x_1 - \xi_1) \delta(x_2 - \xi_2) \delta(x_3 - \xi_3)$ . Note that  $\delta(x_r - \xi_r)$  has the dimension of 1/(unit volume). The three one-dimensional Dirac functions define the point in space where the three perpendicular point forces, as described by the Kronecker symbol in Equation (7), act.

The Green's function acts as a "propagator" of the effects of forces  $f_i^b$ , from the points where they are acting ( $\xi_i$  inside  $V_0$ ) to points  $x_i$  outside  $V_0$ , where the elastic displacement  $u_i$  produces the seismogram. A simplification, often used in the practice, is made by applying the point source approximation. It is valid if the source dimension is much smaller than the considered wavelength and the distance of the observation point from the source. For a point source at  $x_s^0$  we develop the Green's function in Equation(19) in a Taylor series at this point:

$$\begin{aligned} u_i(x_s, \omega) &= \int_{V_0} \left[ f_k^b(x_s^0 + s_s, \omega) G_{ik}(x_s, x_s^0, \omega) + s_j f_k^b(x_s^0 + s_s, \omega) \frac{\partial}{\partial x_j^0} G_{ik}(x_s, x_s^0, \omega) + \dots \right] dV(s_s) \\ &= F_k(x_s^0, \omega) G_{ik}(x_s, x_s^0, \omega) + M_{jk}^f(x_s^0, \omega) G_{ik,j}(x_s, x_s^0, \omega) + \dots \end{aligned} \quad (21)$$

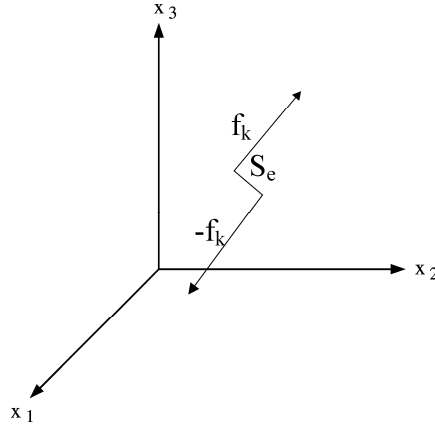
If the source volume is small the Taylor series can be finished after the second term with the first derivative to the source co-ordinates  $x_i^0$ . Then (21) defines the force  $F_k$  and a seismic moment tensor  $M_{kl}^f$  for which the following relations hold:

$$F_k(x_s^0, \omega) = \int_{V_0} f_k^b(x_s^0 + s_s, \omega) dV(s_s) \quad (22)$$

and

$$M_{jk}^f(x_s^0, \omega) = \int_{V_0} s_j f_k^b(x_s^0 + s_s, \omega) dV(s_s). \quad (23)$$

If  $f_k^b$  is a single point force then  $M_{kl}^f$  as a whole describes a force couple (see Figure 5).



**Figure 5** Schematic presentation of a general force couple  $f_i s_j$

Equation (21) contains the partial spatial derivatives of the Green's function. In a homogeneous infinite body they can be written as

$$G_{ki,j} = \frac{1}{r^4 \omega^2} \sum_{n=0}^3 A_{ijk}^{(n)P} \left( \frac{\omega r}{v_p} \right)^n + \frac{1}{r^4 \omega^2} \sum_{n=0}^3 A_{ijk}^{(n)S} \left( \frac{\omega r}{v_s} \right)^n \quad (24)$$

where the  $A_{ijk}^{(n)}$  are complex coefficients proportional to the amplitudes and phases of the P and S waves (see 2.2). The term of  $G_{ij,k}$  with  $n = 3$  is called the *far-field term* because it can still be observed at rather large distances  $r$  between the point source and the point of observation (seismic recording). In contrast, the terms with  $n = 0, 1$  and  $2$  are called the *near field terms* because they decay with distance more rapidly than the far-field term, namely proportional to  $r^{-2}$ ,  $r^{-3}$ , and  $r^{-4}$ , respectively.

Elastic displacements are given now by the time convolution of the forces acting at the focus with the Green's function for the medium. The simplest Green's function is that corresponding to an homogeneous infinite medium (full space). Internal sources must be in equilibrium, thus satisfying the condition that their resulting total force and moment are zero. Therefore, we consider as a seismic source only the symmetric part of  $M_{kl}^f$  as a seismic moment tensor, i.e.,

$$M_{jk} = M_{jk}^f + M_{kj}^f. \quad (25)$$

Fig. 3.34 shows all possible 6 couples and three dipoles of the seismic moment tensor  $M_{jk}$ .

If we want to represent the shear motion on a fault, the equivalent system of forces is that of two couples with no resulting moment, called a double-couple model (DC) (see Figure 8). If the couples are oriented in the direction of the two perpendicular unit vectors  $e_i$  and  $l_i$ , respectively, with  $e_i l_i = 0$ , and if their scalar seismic moment is  $M_0(\omega) = \lim_{|s_i| \rightarrow 0} |s_i| |F_k|$ , where  $|s_i|$  is the length of the arm of the couple and  $|F_k|$  the amount of the force, the displacement caused by the double-couple source is given by

$$u_i^{DC}(x_s, \omega) = M_0(\omega)(e_k l_j + e_j l_k) G_{ik,j}(x_s, x_s^o, \omega). \quad (26)$$

Note that in the given case the comma in the subscripts of  $G$  represents the partial derivative with respect to the source co-ordinates.

If an earthquake is produced by a fault in the Earth's crust, a mechanical representation of its source can be given in terms of fractures or dislocations in an elastic medium. A displacement dislocation consists of an internal surface  $S$  with two sides ( $S^+$  and  $S^-$ ) inside of the elastic medium (see Figure 5) across which there exists a discontinuity of displacement; however, stress is continuous. Thus,  $S$  is a model of a seismic fault. Coordinates on this surface are  $\xi_k$  and the normal at each point is  $n_i$ . From one side to the other of this surface there is a discontinuity in displacement  $D_i$ , which is termed the slip or dislocation on the fault:

$$D_i(\xi_k, \omega) = u_i^+(\xi_k, \omega) - u_i^-(\xi_k, \omega). \quad (27)$$

The plus and minus signs refer to the displacement at each side of the surface  $S$ . If there are no body forces ( $F_i = 0$ ), and the stresses are continuous through  $S$ , then, for an infinite medium, the equation relating the displacement to the dislocation  $D_i$ , results in

$$u_n(x_s, \omega) = \int_S D_i(\xi_s, \omega) c_{ijkl} n_j(\xi_s) G_{nk,l}(x_s, \xi_s, \omega) dS(\xi_s). \quad (28)$$

Equation (27) corresponds to a kinematic model of the source, that is a model in which elastic displacements  $u_i$  are derived from slip vector  $D_i$ . The latter represents a non-elastic displacement of the two sides of a fault (i.e., of the model surface  $S$ ). In a kinematic model slip is assumed to be known. It is not derived from stress conditions in the focal region as it is in dynamic models. Equation (28) contains the Green's function discussed in conjunction with Equation (24). When seismic waves, generated by the source, are observed in the far-field, i.e., at distances  $r$  much larger than the wavelength and the linear source dimension, then the Green's function is proportional to  $\omega$ . Accordingly, the dominant term of the integrant in Equation (28) is  $\omega D_i$  which is, in the time domain, proportional to the slip velocity. Thus, the elastic displacement observed in the far-field does not depend on the slip in the source but on the slip velocity and, similarly, on the seismic moment rate  $\partial M_{ik}(t)/\partial t = \dot{M}_{ik}$  (see Fig. 2.4). Or, in the frequency domain, the displacement is proportional to  $i\omega M_{kl}(\omega)$ . This means that the source radiates elastic energy only while it is moving; when motion at the source stops it ceases to radiate energy.

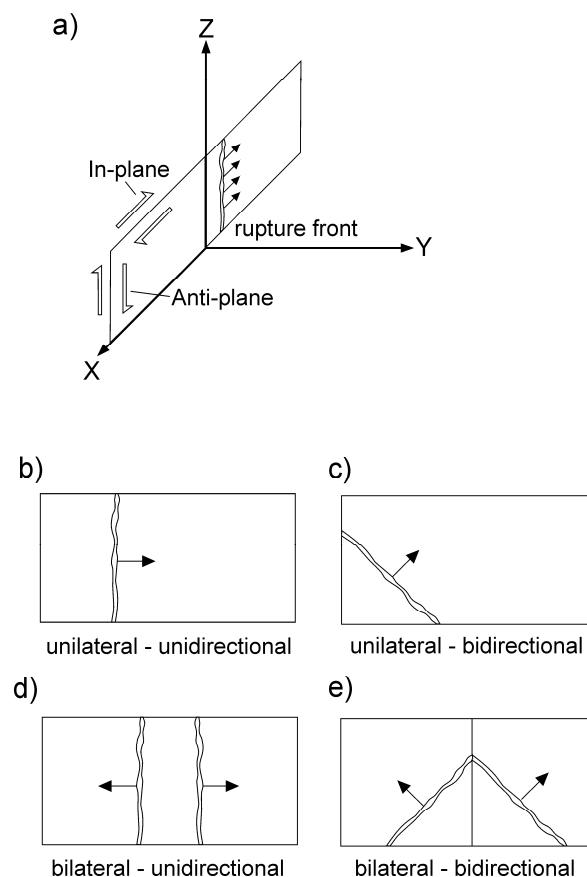
The most common model for the source of an earthquake is a shear fracture, that is, a fracture in which the slip  $D_i$  is perpendicular to the normal of the fault. For a fault plane  $S$  of area  $A$  and normal  $n_i$ , the slip  $D_i(\xi_s, t)$  is in the direction of the unit vector  $l_i$  contained in the plane. Accordingly,  $l_i$  and  $n_i$  are perpendicular and the scalar product  $n_i \cdot l_i = 0$ . For an infinite, homogeneous isotropic medium, displacement according to Equation (28) is given by

$$u_i(x_s, \omega) = \int_S \mu |D_l(\xi_s, \omega)| (l_k n_j + l_j n_k) G_{ik,j}(x_s, \xi_s, \omega) dS(\xi_s) \quad (29)$$

For modeling a shear dislocation source, the parameters on the right-hand side of Equation (29) have to be known. Implicitly these parameters include information about the rupture propagation, i.e., on the shape of the crack front, its propagation direction and propagation velocity (crack velocity), and shape of the final ruptured surface  $S$ .

The circular fault and the rectangular fault are the most important approximations. In the first case the rupture begins at the center and the crack front is described by an outward propagating circle. However, the direction of the dislocation is not necessarily radially symmetric. This circular model, described by Brune (1970) and Madariaga (1976), should be valid for small earthquakes with magnitudes smaller than about 4 to 5. Another approximation, for large earthquakes in the Earth's crust in particular, is a rectangular fault model, also called Haskell-model (Haskell, 1964). The length of the fault, generally assumed to be horizontal, is larger than its width (depth) by a factor of 2 to 10 or even more for very large earthquakes. This is due to the limited thickness of the seismogenic zone of the upper lithosphere, usually ranging between about 10 and 25 km, where brittle fracturing is possible. On the other hand, large crustal earthquakes may have a rupture length of 200 km or even more, e.g., about 450 km for the Alaska earthquake of 1964 and about 1000 km for the Chile earthquake of 1960. This rectangular model is also useful for describing deeper earthquakes in subduction zones.

When the Haskell-model is used the behavior of the rupture front must be known. The first approximation is that the rupture starts along a line and propagates unilaterally or bilaterally over the rectangular fault plane (see Figure 6). This approximation is useful for long ruptures with small width (the line-source approximation). It is also suitable for distinguishing between an in-plane and an anti-plane fault geometry. In the case of an in-plane fault the rupture moves into the direction of the slip whereas in the anti-plane case the direction of slip is parallel to the rupture front (see Figure 6).



**Figure 6** Several models of rupture propagation

For describing the rupture propagation in the case of a rectangular fault the following four terms and definitions, shown in Figure 6, are important:

- unilateral rupture propagation – one rupture front propagates over the entire fault plane;
- bilateral rupture propagation – two rupture fronts with different directions propagate over the rupture plane;
- unidirectional rupture propagation – the direction of rupture propagation is parallel to the length of the fault plane; and
- bidirectional rupture propagation – the rupture starts at a point and propagates across the fault plane.

Other models for describing the shape of the fault plane, the shape of the rupture front, and the mode of the rupture propagation are possible.

With respect to the velocity of rupture propagation the most common models assume values between about 0.6 to 0.9 of the shear-wave velocity  $v_s$  (see 2.2) in the source region; however, detailed field and laboratory investigations have shown that both slower (so-called “silent earthquakes”) and supersonic ( $> v_s$ ) rupture propagation velocities are possible (e.g., Tibi et al., 2001). Rupture velocity depends on the material properties, the internal friction of the unbroken material, the frictional conditions along the fractured surface and the stress conditions (ambient and on the crack tip) in the given case.

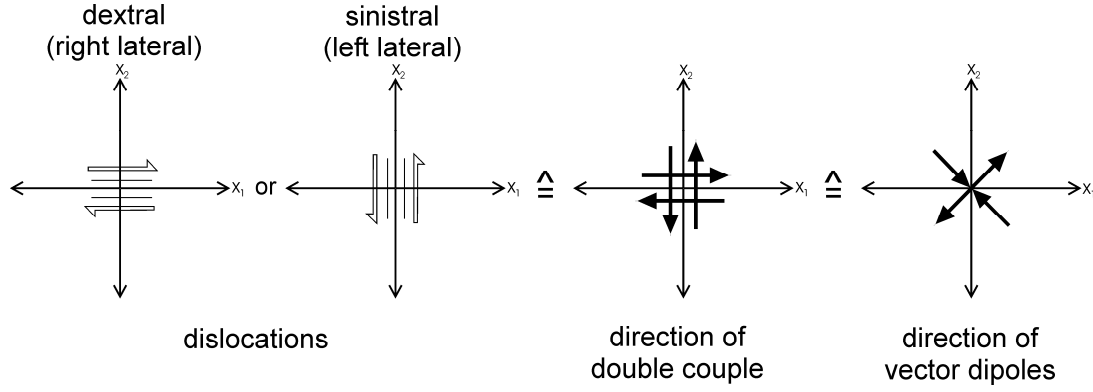
For the point source approximation Equation (29) takes the simpler form

$$u_i(\omega) = \mu A |D_l(\omega)| (l_k n_j + l_j n_k) G_{ik,j}(\omega) \quad (30)$$

or, in the time domain,

$$u_i(t) = \mu A (l_k n_j + l_j n_k) \int_{-\infty}^{\infty} |D_l(\tau)| G_{ik,j}(t - \tau) d\tau. \quad (31)$$

Displacements are given by temporal convolution of slip with the derivatives of the Green’s function. The geometry of the source is now defined by the orientation of the two unit vectors  $n_i$  and  $l_i$ . These two vectors, which refer to the geophysical co-ordinate system of axes (North, East, Nadir), define the orientation of the source, namely  $n_i$  the orientation of the fault plane and  $l_i$  the direction of slip. These two vectors can be written in terms of the three angles that define the motion on a fault, namely, azimuth  $\phi$ , dip  $\delta$  and rake  $\lambda$ . The shear fracture itself is equivalent to a DC source in terms of forces (see Figure 7).

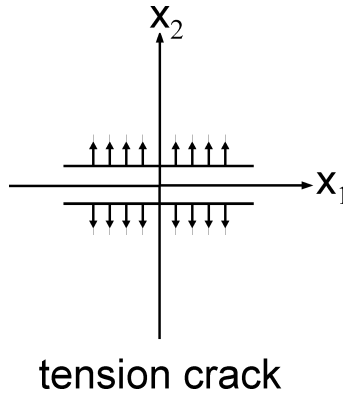


**Figure 7** Depiction of the equivalence of a shear dislocation with the force double couple and the vector dipole models.

In the case that  $l_i$  and  $n_i$  are not perpendicular, Equation (29) has to be replaced by

$$u_k(x_s, \omega) = \int_S [\lambda \delta_{jk} n_l l_l + \mu (l_k n_j + l_j n_k) G_{ik,j}(x_s, \xi_s, \omega)] D_n(\xi_s, \omega) dS(\xi_s). \quad (32)$$

The special case when  $l_i$  and  $n_i$  are parallel is often used to model tensional volcanic earthquakes (Figure 8).



**Figure 8** Illustration of a tension crack which is often used in modeling volcanic earthquakes.

Another more general representation of seismic source is given by the seismic moment tensor density  $m_{ij}$ . The moment tensor density represents that part of the internal strain drop which is dissipated in non-elastic deformations at the source. So far we have modeled the seismic source by means of the forces in the equation of motion (see Equation (1)) or by boundary conditions for the displacement (see Eqs. (17) and (28)). Now we take another approach and divide the true strain tensor  $\varepsilon_{ij}^{true}$  into an elastic and inelastic part, i.e.,

$$\varepsilon_{ij}^{true} = \varepsilon_{ij}^{ek} - \varepsilon_{ij}^{inel}. \quad (33)$$

With this we define the true stress



$$\sigma_{ij}^{true} = \sigma_{ij} - m_{ij}^V \quad (34)$$

where  $\sigma_{ij}$  is the elastic stress related to the strain by Equation (10) or (13) and  $m_{ij}^V$  is given by

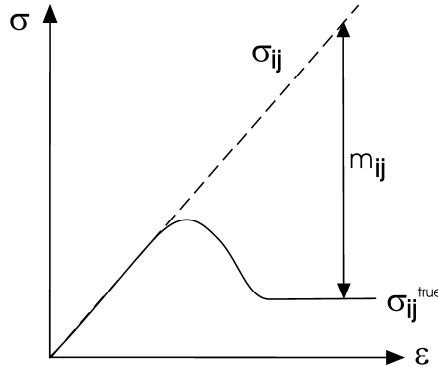
$$m_{ij}^V = c_{ijkl} \varepsilon_{kl}^{inel} . \quad (35)$$

Equation (35) defines the seismic moment tensor density  $m_{ij}^V$ . The superscript  $V$  indicates that it is a volumetric density. Rice (1980) and Madariaga (1983) denote  $\varepsilon_{ij}^{inel}$  as the stress-free strain or transformation strain, and  $m_{ij}^V$  as the stress glut. The seismic moment tensor  $M_{ij}$  is, thus, defined by

$$M_{ij}(\omega) = \int_{V_o} m_{ij}^V(x_k, \omega) dV(x_k) . \quad (36)$$

The quantities  $m_{ij}^V$  and  $M_{ij}$  play a fundamental role in the theory of seismic sources. The relations between the different kinds of stress are shown in Figure 9. When  $\sigma_{ij}$  in (15) is substituted by  $\sigma_{ij}^{true}$  an additional force term appears on the right side. It can be interpreted as an equivalent force density  $f_i^{eq}$  or as an equivalent force  $F_i^{eq}$

$$f_i^{eq}(x_k, \omega) = -m_{ij,j}^V(x_k, \omega) \quad \text{and} \quad F_i^{eq} = -\int_{V_o} m_{ij,j}^V(x_k, \omega) dV(x_k) . \quad (37)$$



**Figure 9** Relationship between the elastic stress  $\sigma_{ij}$ , related to the strain  $\varepsilon$ , the true stress  $\sigma_{ij}^{true}$  and the seismic moment density tensor  $m_{ij}^V$ .

In replacing the body force in Equation (19) by the equivalent force density in Equation (37) an additional volume integral  $-\int_{V_o} G_{ij} m_{jk,k}^V dV$  appear. After an integration by parts and assuming that  $m_{jk}^V$  vanishes on  $S$ , i.e., the inelastic volume is bordered by  $S$ , the displacement produced by  $m_{ij}^V$  is

$$u_i(x_i, \omega) = \int_{V_o} G_{ik,j}(x_s, \xi_s, \omega) m_{jk}^V(\xi_s, \omega) dV(\xi_s) . \quad (38)$$

When comparing Eqs. (38) and (28) one realizes that the integrands have the same form but the integration in (38) is over a volume while it is over a surface in Equation (28).

Accordingly, the stress glut  $m_{ij}^V$  is equivalent to a dislocation when the inelastic volume can be approximated by an inelastic internal surface. Naming this stress glut by  $m_{ij}^S$  from Equation (28) we see that

$$m_{kl}^S = c_{ijkl} D_i n_j \quad (39)$$

for the general linear elastic case and for the shear crack in an isotropic medium holds

$$m_{ij}^S = \mu (D_i n_j + D_j n_i). \quad (40)$$

For the spatially averaged dislocation  $\overline{D_i}(\omega)$ , the seismic moment tensors  $M_{ij}$  in these two cases become

$$M_{ij}(\omega) = c_{ijkl} \overline{D_k}(\omega) n_l A \quad \text{and} \quad M_{ij}(\omega) = \mu [\overline{D_i}(\omega) n_j + \overline{D_j}(\omega) n_i] A, \quad (41)$$

respectively. In the latter case, when  $\overline{D_i}$  and  $n_j$  are perpendicular, the scalar seismic moment is

$$M_0(\omega) = \mu |\overline{D_i}(\omega)| A. \quad (42)$$

In the general case of an arbitrary moment tensor the scalar seismic moment is defined by

$$M_0 = \sqrt{\frac{1}{2} M_{ik} M_{ik}}. \quad (43)$$

## 4 Dynamic source models

Dynamic source models, or crack models, use a given stress on an internal surface (fault) to describe a seismic source. In this case Equation (19) is not valid. In general, two terms must be added to the right side of Equation (19). These terms include boundary conditions for the displacement and the stress. Note that only one of these conditions can be freely chosen while the other one has to be calculated. The computation of the Green's function requires boundary conditions as well, either for the Green's function itself or for the stress produced by it. These boundary conditions do not influence the result of the computation of the displacement  $u_i(x_s, \omega)$ . Therefore, we can freely select any suitable boundary conditions. When selecting a Green's function which produces a vanishing stress on the internal surface  $S$  this Green's function is called  $G_{ij}^{free}$  because the related internal surface behaves like a free surface. The advantage is, that this kind of source representation does not require a knowledge of the displacement produced by the given stress on the internal surface. When no body force acts it holds that

$$u_i(x_i, \omega) = \int_S G_{ik}^{free}(x_s, \xi_s, \omega) n_j \sigma_{kj}(\xi_s, \omega) dS(\xi_s). \quad (44)$$

Equation (44) simplifies the computation of the displacement or the dislocation on the fault when the stress on the fault is given. When using other kinds of representations an inhomogeneous integral equation for  $u_i(\xi_s, \omega)$  on the fault has to be solved.

In the dynamic models the static stress drop  $\Delta\sigma_{ij}$  plays an important role. It is defined as the difference between the stress distribution  $\sigma_{ij}^o$  on the fault plane before the occurrence of the earthquake and the stress  $\sigma_{ij}^1$  after the earthquake. This static stress drop is

$$\Delta\sigma_{ij}(\xi_s) = \sigma_{ij}^o(\xi_s) - \sigma_{ij}^1(\xi_s) \quad (45)$$

with  $\sigma_{ij}^1(\xi_s) = \lim_{t \rightarrow \infty} \sigma_{ij}(\xi_s, t) = \lim_{\omega \rightarrow 0} i\omega \sigma_{ij}(\xi_s, \omega)$ . A more general time dependent stress on the fault is shown in Figure 10 (Yamashita, 1976).

A case of practical importance is that of a circular shear fault. It is probably a good approximation for small earthquakes in the Earth's crust with magnitudes smaller than 4 as long as only frequencies  $< 5$ -10 Hz are considered. If a homogeneous shear stress drop  $\Delta\sigma_{12}$  in the  $x_1$ - $x_2$  plane is assumed, the static dislocation on the fault is

$$D_1 = \frac{8}{\mu\pi} \frac{\lambda + 2\mu}{3\lambda + 4\mu} \Delta\sigma_{12} (R_0^2 - r^2)^{1/2} \quad (46)$$

where  $R_0$  is the final radius of the broken fault and  $r$  the radial co-ordinate. If  $r > R_0$  the dislocation in Equation (46) is zero. By inserting Equation (46) in (41) we get for the static seismic moment

$$M_0 = \frac{16}{3} \frac{\lambda + 2\mu}{3\lambda + 4\mu} \Delta\sigma_{12} R_0^3 \quad (47)$$

and for  $\lambda = \mu$  the well known result derived by Keilis-Borok (1959) is given by

$$M_0 = \frac{16}{7} \Delta\sigma_{12} R_0^3. \quad (48)$$

Similar relations hold for rectangular shear cracks of the length  $L$  and a width  $W$ :

$$M_0 = C L^2 W \Delta\sigma_{12} \quad (49)$$

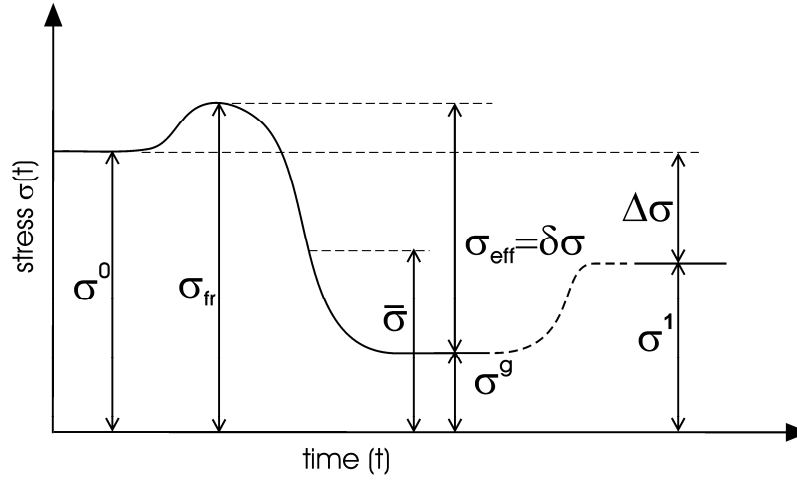
where  $C$  is a model-dependent constant in the order of 1 and  $\Delta\sigma_{12}$  is uniform over the fault. In the case of a buried in-plane shear crack holds

$$C = \frac{\pi}{8} \frac{\lambda + 2\mu}{\lambda + \mu} \quad (50)$$

and for a buried anti-plane case

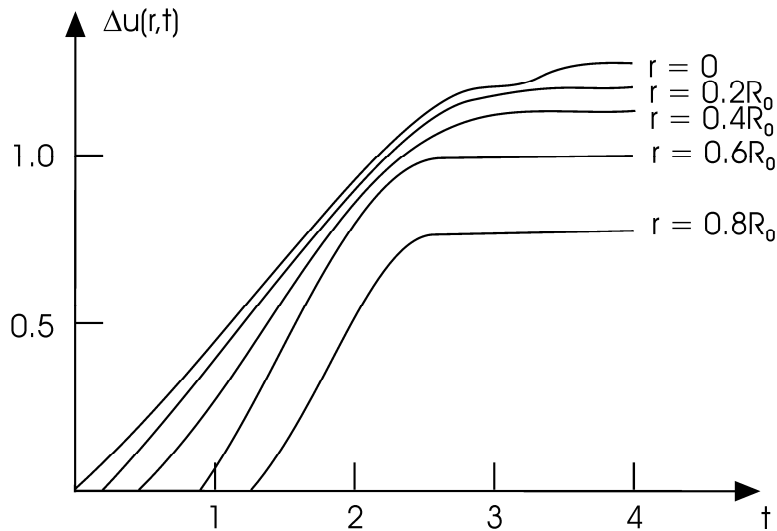
$$C = \frac{\pi}{4}. \quad (51)$$

When the fault is perpendicular to the Earth's surface and outcropping then  $C$  in the Eqs. (50) and (51) is twice as large.



**Figure 10** Time dependence of stress at a point on the fault surface during an earthquake.  $\sigma^0$  and  $\sigma^1$  – stress before and after the earthquake,  $\sigma_{fr}$  – fracture strength,  $\bar{\sigma}$  – mean stress,  $\sigma^g$  – friction stress,  $\sigma_{eff}$  – effective stress = dynamical stress drop  $\delta\sigma$  and  $\Delta\sigma$  – static stress drop.

The dynamic relation between the shear stress drop  $\Delta\sigma$  and the dislocation can be calculated numerically. An example is shown in Figure 11. The rupture starts at  $t=0$  and  $r=0$  and expands with constant velocity. The time  $t$  and the dislocation  $|D_I(r,t)|$  are normalized to  $R_0/V_p$  and  $\Delta\sigma R_0/\mu$  where  $V_p$  is the velocity of the P wave ( $V_p^2 = (\lambda + 2\mu)/\rho$  with  $\rho$  as the density of the medium).



**Figure 11** Dislocation function  $D(r, t)$  at several distances from the center on the circular crack plotted against the normalized time  $t$ . For explanation of symbols see text (according to Madariaga, 1976; modified from Aki and Richards, 1980).

## 5 Energy, moment, dislocation and stress drop

The radiated energy of an earthquake can be computed assuming a specific source model and its source parameters. We describe the earthquake as a shear rupture on a surface. In a relatively general form Kostrov (1975) writes for the radiated seismic energy  $E_s$

$$E_s = \int_0^{t_{\max}} dt \int_{S(t)} dS (\xi_k) (\sigma_{ij}^0 - \sigma_{ij}) \dot{D}_i n_j - \frac{1}{2} \int_A dS \Delta \sigma_{ij} D_i^0 n_j - \int_A g dS \quad (52)$$

where  $t_{\max}$  is the maximum duration of the motion on the fault plane,  $S(t)$  the rupture plane developing during the rupture,  $A$  the final rupture plane with  $A = \lim_{t \rightarrow \infty} S(t)$ ,  $\sigma_{ij}^0(\xi_s)$  the stress before the earthquake occurred,  $\sigma_{ij}(\xi, t)$  the stress on the broken fault surface,  $\dot{D}_i(\xi_s, t) = \partial/\partial t D_i(\xi_s, t)$  the dislocation velocity,  $\Delta \sigma_{ij}(\xi)$  the static stress drop (see Figure 10),  $n_j$  the normal vector of the fault surface,  $D_i^0(\xi_s)$  the static dislocation, and  $g$  the specific energy required to generate a new surface. Equivalent to Equation (52) is the often used form

$$E_s = \int_0^{t_{\max}} dt \int_{S(t)} dS \dot{D}_j n_j (\bar{\sigma}_{ij} - \sigma_{ij}) - \int_A g dS \quad (53)$$

where  $\bar{\sigma}_{ij} = (\sigma_{ij}^0 + \sigma_{ij}^1)/2$  denotes the mean stress,  $\sigma_{ij}^0$  is the stress before the earthquake, and  $\sigma_{ij}^1$  is the final stress, which may be equal to the frictional stress. When taking into account the grow of the rupture area during the earthquake in the formulation of the dislocation (source time) function  $D_i(t)$ , Equation (53) becomes

$$E_s = \int_A dS \int_0^{D_i^f} dD_i n_j [\bar{\sigma}_{ij} - \sigma_{ij}(D_k)] - \int_A g dS \quad (54)$$

where  $\sigma_{ij}(D_k)$  is the stress-dislocation relation on the fault plane and  $D_i^f$  the final dislocation.. In the Eqs. (52) to (54) the seismic energy  $E_s$  is composed of released deformation energy  $E_{tot}$ , frictional energy  $E_f$ , and rupture (crack) energy  $E_r$

$$E_s = E_{tot} - E_f - E_r \quad (55)$$

with

$$\begin{aligned} E_{tot} &= \int dS \bar{\sigma}_{ij} D_i n_j \\ E_f &= \int dS \int dD_i n_j \\ E_r &= \int g dS \end{aligned} \quad (56)$$

With this we define the seismic efficiency  $\eta$

$$\eta = \frac{E_s}{E_{tot}}$$

(57)

and the apparent stress  $\sigma_{app}$

$$\sigma_{app} = \frac{1}{A|\overline{D_i}|} \int e dS \quad (58)$$

where  $|\overline{D_i}|$  is the spatial averaged absolute value of the dislocation. The energy density  $e$  is identical with the integrant of the surface integral. Therefore the following relation between the seismic energy and the scalar seismic moment holds:

$$E_s = \sigma_{app} M_0 / \mu \quad (59)$$

Further special cases are:

a)

$\sigma_{ij}^0$ ,  $\sigma_{ij}$ ,  $\sigma_{ij}^1$  are homogeneous and  $\sigma_{ij}$  equal to the time-independent friction stress  $\sigma_{ij}^g$  Eqs. (3), (6) and (7) yield

$$E_s = (\overline{\sigma}_{ij} - \sigma_{ij}^g) D_i^f n_j S_0 - E_r \quad (60)$$

$$\eta = \frac{(\sigma_{ij}^0 + \sigma_{ij}^1 - 2\sigma_{ij}^g) e_i n_j - 2\bar{g}}{(\sigma_i^0 + \sigma_{ij}^1) e_i n_j} \quad (61)$$

$$\sigma_{app} = \frac{1}{2} (\sigma_{ij}^0 + \sigma_{ij}^1 - 2\sigma_{ij}^g) e_i n_j - \frac{\bar{g}}{|\overline{D_i}|} \quad (62)$$

where  $\bar{g}$  is the averaged specific rupture energy and  $e_i$  a unit vector in the direction of the dislocation. With this we get

$$\sigma_{app} = \eta \overline{\sigma}_{ij} e_i n_j. \quad (63)$$

b)

For a shear fracture, and  $\sigma_{ij}^g = \sigma_{ij}^1$  with  $g \approx 0$  as an approximation or  $g = 0$  in the case of an *anti-plane* brittle rupture propagating with shear-wave velocity or of an *in-plane* brittle rupture propagating with Rayleigh-wave velocity, respectively, we get

$$E_s = \frac{1}{2} D_i^f n_j \Delta\sigma_{ij} S_0 = \frac{1}{2\mu} \Delta\sigma_{ij} M_{ij} \quad (64)$$

with

$$\Delta\sigma_{ij} = \sigma_{ij}^0 - \sigma_{ij}^1. \quad (65)$$

Ohnaka (1978) gives the following relationship for the seismic energy of a circular shear fracture propagating with the crack velocity  $v_c = 0.8 v_s$ :

$$E_s = \frac{M_o \overline{D_0}}{2R} \quad (66)$$

with  $M_o$  – scalar seismic moment,  $\overline{D_0}$  – static averaged dislocation and  $R$  – source radius. For rectangular shear fractures of length  $L$  and with unilateral fracture propagation a similar approximate relationship holds:

$$E_s \approx \frac{M_o \overline{D_0}}{3L} \quad (67)$$

and in case of partial incoherence

$$E_s \approx \frac{M_o \overline{D}}{L}. \quad (68)$$

Further,  $E_s$  can be determined directly by integrating over the displacement field. It holds

$$E_s = \sum_k \int_{-\infty}^{\infty} dt \int_S dS \rho v^{(k)} \dot{u}_i^{(k)} \dot{u}_i^{(k)} \quad (69)$$

with  $S$  – a surface surrounding the source,  $\rho$  – density distribution on this surface,  $\dot{u}_i^{(k)}$  – velocity of ground motion. The sum is over all kinds of waves which leave the volume enclosed by the surface  $S$  with the velocity  $v^{(k)}$ . However, one has to take into account that on the way from the source to  $S$  part of the energy has already been transformed into heat by inelastic effects of wave propagation.

Equation (69) forms the theoretical background for the simple relationship between seismic energy and magnitude  $M$

$$\log E_s = a M + b \quad (70)$$

which is based on rather simple assumptions. Nevertheless, the corresponding relationship given by Gutenberg and Richter (1956) is

$$\log E_s[J] = 1.5 M_s + 4.8 \quad (71)$$

with  $M_s$  – surface wave magnitude (see 3.2.5.1). Equation (71) has proven to yield rather good estimates of  $E_s$ . More details on direct energy determination based on digital broadband recordings is outlined in 3.3.

## Acknowledgments

The authors acknowledge with thanks the fruitful discussions with Rongjiang Wang and Dietrich Stromeyer of the GFZ Potsdam and careful reviews by George Choy and Günter Bock <sup>‡</sup>. Their suggestions have helped to improve the first draft of this information.

## References

- Aki, K., and Richards, P. G. (1980). Quantitative seismology. *Freeman*, San Francisco, Vol. I and II, 932 pp.
- Aki, K., and Richards, P. G. (2002). Quantitative seismology. Second Edition, ISBN 0-935702-96-2, *University Science Books*, Sausalito, CA., xvii + 700 pp.
- Ben-Menahem, A., and Singh, S. J. (1981). Seismic waves and sources. *Springer-Verlag*, New York, Heidelberg, Berlin, 1108 pp.
- Brune, J. N. (1970). Tectonic stress and the spectra of shear waves from earthquakes. *J. Geophys. Res.*, **75**, 4997-5009.
- Das, S., and Kostrov, B. V. (1988). Principles of earthquake source mechanics. *Cambridge University Press*.
- Gutenberg, B., and Richter, C. F. (1956a). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Haskell, N. A. (1964). Total energy and energy spectral density of elastic wave radiation from propagating faults. *Bull. Seism. Soc. Am.*, **54**, 1811-1841.
- Keilis-Borok, V. I., (1959). On the estimation of displacement in an earthquake source and of source dimension. *Ann. Geofis.*, **12**, 205-214.
- Kostrov, B. V. (1975). *Mechanika otschaga tektonitscheskogo semletrjasenija. Isdatelstwo Nauka*, Moskva.
- Lay, T., and Wallace, T. C. (1995). Modern global seismology. ISBN 0-12-732870-X, *Academic Press*, 521 pp.
- Madariaga, R. (1976). Dynamics of an expanding circular fault. *Bull. Seism. Soc. Am.*, **66**, 639-666.
- Madariaga, R. (1983). Earthquake source theory: a review. In: Earthquakes: Observation, Theory and Interpretation (Ed. Kanamori, H., Boschi, E.). Proceedings of the International School of Physics "Enrico Fermi". *North-Holland Publishing Comp.*, Amsterdam, New York, Oxford, 1-44.
- Ohnaka, M. (1978). Application of some dynamic properties of stick-slip to earthquakes. *Geophys. J. R. Astr. Soc.*, **53**, 311-318.
- Rice, J. R. (1980). The mechanics of earthquake rupture. In: Physics of the Earth's interior (Ed. Dziewonski A. A., Boschi, E.). *North Holl. Publ. Co.*, Amsterdam, 555-649.
- Yamashita, T. (1976). On the dynamic process of the fault motion in the presence of friction and inhomogeneous initial stress. Part 1., Rupture propagation. *J. Phys. Earth*, **24**, 417-444.
- Udías, A. (1999). Principles of Seismology. *Cambridge University Press*, United Kingdom, 475 pp.
- Udias, A. (2002). Theoretical seismology: An introduction. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). International Handbook of Earthquake and Engineering Seismology, Part A. *Academic Press, Amsterdam*, 81-102.
- Yamashita, T. (1976). On the dynamic process of the fault motion in the presence of friction and inhomogeneous initial stress. Part 1., Rupture propagation. *J. Phys. Earth*, **24**, 417-444.



| Title   | Proposal for unique magnitude and amplitude nomenclature                                                                                               |
|---------|--------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Peter Bormann (formerly GeoForschungsZentrum Potsdam,Telegrafenberg, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version | June 2013; DOI: 10.2312/GFZ.NMSOP-2_IS_3.2                                                                                                             |

|                                                                                                                                              | page         |
|----------------------------------------------------------------------------------------------------------------------------------------------|--------------|
| <b>1 Introduction</b>                                                                                                                        | <b>1</b>     |
| <b>2 Seismic parameter data formats</b>                                                                                                      | <b>6</b>     |
| <b>3 IASPEI standard magnitudes</b>                                                                                                          | <b>7</b>     |
| <b>4 Magnitude and amplitude nomenclature applied by the USGS-NEIC</b>                                                                       | <b>9</b>     |
| <b>5 Magnitude nomenclature reported to the ISC</b>                                                                                          | <b>10</b>    |
| <b>6 Proposal for a more generalized unique nomenclature for amplitude and magnitude data</b>                                                | <b>13</b>    |
| <b>7 Author or agency specific magnitude nomenclature</b>                                                                                    | <b>18</b>    |
| <b>8 The role of the ISC in implementing IASPEI standard magnitude measurements and assuring unique amplitude and magnitude nomenclature</b> | <b>20</b>    |
| <b>Acknowledgment</b>                                                                                                                        | <b>22</b>    |
| <b>References</b>                                                                                                                            | <b>22-26</b> |

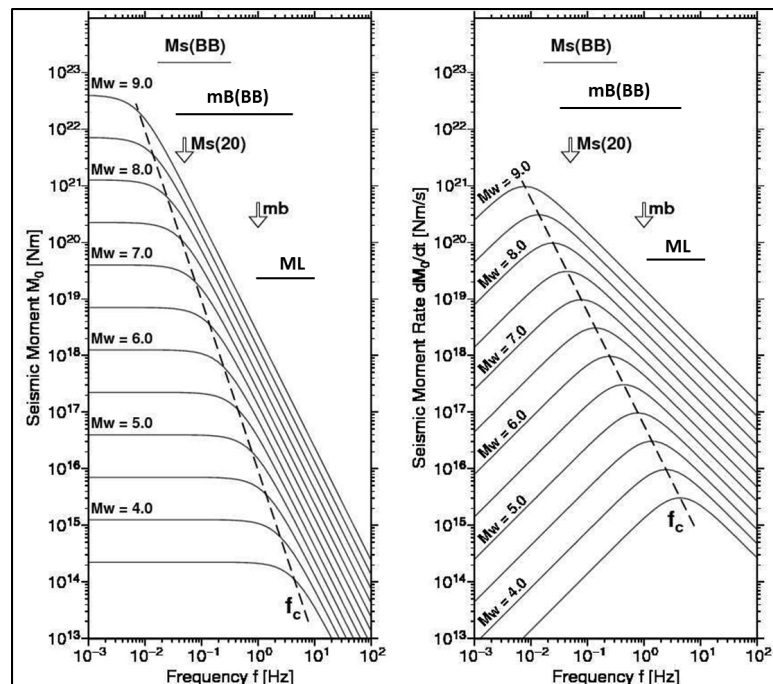
## 1 Introduction

Besides location the magnitude is the most frequently determined and important parameter to characterize a seismic source and for earthquake statistics, aimed at assessing seismic activity and hazard, even the most important one. Classical magnitudes are based on the measurement of amplitudes and periods of seismic waves which also allow inferences on the structural and physical properties of the propagation medium. The usefulness of amplitude, period and magnitude data for application and research essentially depends on their long-term availability, homogeneity, reproducibility, compatibility and reliability. But these properties require globally agreed and applied measurement, calculation and data representation (nomenclature) standards. Inconsistencies in these crucial aspects increase data scatter, may even results in systematic biases, wrong inferences and thus strongly reduce the value of such data. This the more so since it has become obvious that no single magnitude is able to characterize the “sizes” of an earthquake in all of their geometric, kinematic and dynamic aspects. Better characterization of different aspects of earthquake size, desirable in order to better anticipate associated earthquake effects, requires a complementary set of magnitude types. Although this is outlined in detail in sections 3.1.2, 3.2 and 3.3 of Chapter 3 we summarize here some of the essential related aspects in order to better appreciate the need for a sufficiently detailed and unambiguous magnitude nomenclature.

The effort by Gutenberg and Richter (1954) to classify in “Seismicity of the Earth” earthquakes of different size by just ONE unspecified magnitude value  $M$  was one of the outstanding seismological milestones of the 20<sup>th</sup> century. It is now recognized, however, that a relative-size classification by a single type of magnitude implicitly presupposes an over-

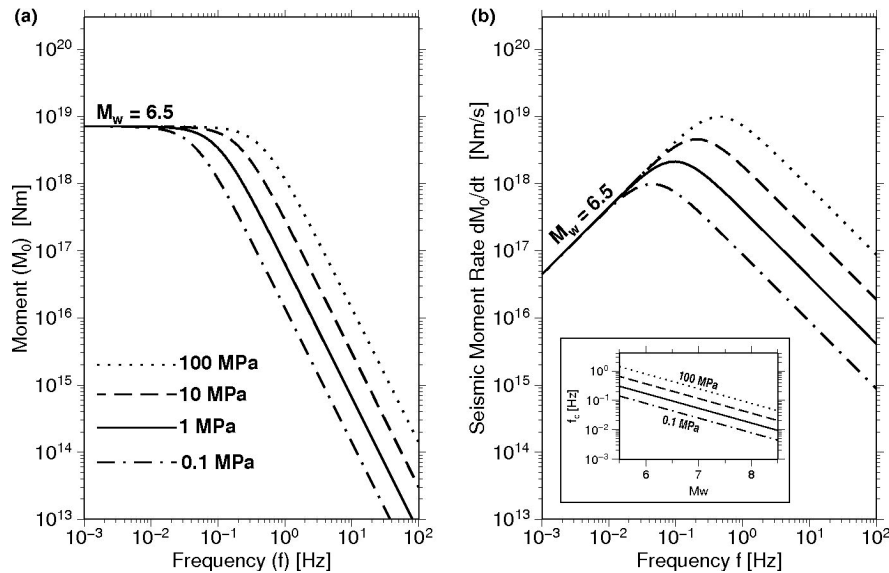
simplified assumption of the similarity of the source process of all earthquakes and thus of their spectra, as, e.g., in Figure 1. The condition of source similarity (see section 3.3.1 of Chapter 3) holds only under certain conditions of geometric and dynamic similarity, such as constant stress drop  $\Delta\sigma$ . But for individual earthquakes  $\Delta\sigma$  may vary by several orders of magnitude and thus the corner frequency  $f_c \sim (\Delta\sigma)^{1/3}$  of the source spectrum up to a factor of 10 or even more. This means, however, that for a given static seismic moment or rupture “size” the ratio between released high and low frequency energy and thus the earthquake’s potential to cause shaking damage or to generate a tsunami is not constant. Therefore, magnitudes based on amplitudes measured in either the long-period or more medium to short-period range of the source spectrum may be rather different and thus the conclusions with respect to the size or “strength” of an earthquake and its related hazard potential (e.g., Choy and Kirby, 2004; Di Giacomo et al., 2008 and 2010; Bormann and Di Giacomo, 2011).

Abe and Kanamori were aware of this and therefore “decomposed” the Gutenberg-Richter (1954) and later published unspecified  $M$  for large earthquakes on the basis of Gutenberg’s original note pads and other sources into the original Gutenberg (1945a) 20 s surface-wave magnitude  $M_s$  and the more broadband 2-20s body-wave magnitude  $mB$  (Gutenberg and Richter 1956) [Abe (1981), (1982), and (1984); Abe and Kanamori (1979) and (1980)]. And the short-period “generic” magnitudes such as teleseismic  $mb$  and local  $ML$ , which are based on amplitudes at frequencies in the range  $0.3 \text{ Hz} < f < 10 \text{ Hz}$ , reveal in their relationship to the long-period event magnitudes even more clearly such hazard relevant shifts in source spectra corner frequencies for equal seismic moments away from or towards frequencies that are of particular interest to earthquake engineers (see Figures 1 and 2 below).



**Figure 1 Left:** "Source spectra" of a simplified “omega-squared” model, modified according to Bormann et al. (2009). Plotted are the ground displacement amplitudes  $A$  for “average” seismic shear sources with constant stress drop  $\Delta\sigma = 3 \text{ MPa}$  as a function of frequency  $f$ , scaled to seismic moment  $M_0$  and the equivalent moment magnitude  $M_w$ . **Right:** The same as left, but for ground motion velocity amplitudes  $V$ , scaled to seismic moment rate and  $M_w$ .

Note that the maximum of seismic energy  $E_s \sim V^2$  is radiated around the corner frequency  $f_c$ . The open arrows point to the center frequencies on the abscissa at which the 1 Hz body-wave magnitude  $m_b$  and the 20 sec surface-wave magnitude  $M_s(20)$ , respectively, are determined. The horizontal interval bars mark the range of frequencies within which broadband body- and surface-wave magnitudes  $mB(BB)$  and  $M_s(BB)$  and the local magnitude  $M_L$  should be measured.



**Figure 2** Far-field displacement and (b) velocity source spectra scaled to seismic moment and moment rate, respectively, for a model earthquake with  $M_w = 6.5$  but different stress drop  $\Delta\sigma$  in units of MPa. The spectra were calculated based on the same model assumptions as in Figure 1. The inset in Figure 2 (b) shows the variation of the corner frequency  $f_c$  obtained according to the Brune (1970, 1971) equation  $f_c = c v_s (\Delta\sigma/M_0)^{1/3}$  in a wider range of  $M_w$  for varying  $\Delta\sigma$  in increments of one order between 0.1 and 100 MPa. (Copy of Fig. 2, p. 415 in Bormann and Di Giacomo, 2011, J. Seismology, **15** (2), 411-427; © Springer Publishers).

This questions the whole currently still dominating concept of the seismic hazard community of "unifying" magnitudes in earthquake catalogs to moment magnitudes with  $M_0$  being a purely static-geometric parameter of rupture size corresponding to the zero-frequency plateau amplitude in Figure 1. This neglects the "dynamic variability" of the individual rupture process which is reflected, e.g., in the position of the corner frequency towards higher frequencies for equal seismic moment (Figure 2) and in the empirically proven variability of the decay of spectral displacement amplitudes towards higher frequencies with slopes varying between about 1 and 5 for  $f \gg f_c$ . Moreover, for the majority of instrumentally recorded events no accurate direct seismic moment measurements are available. This necessitates to estimate inferior  $M_w$  proxies via often rather noisy empirical correlation relationships between proper  $M_w$  with classical magnitudes (see ISC-GEM project: Global Instrumental Earthquake Catalogue (1900-2009), Di Giacomo et al., 2013). No single value magnitude type, including the supposedly non-saturating  $M_w$ , will ever allow to quantify both the (geometric) "size" and the seismic radiation efficiency and thus the potential "shaking strength" of an earthquake. Some of the currently common  $M_w$  procedures do not even fulfill the primary justification for having  $M_w$  as a non-saturating complement/extension to  $M_s$  (compare  $M_{wp}$ ,  $M_{wb}$  and  $M_{wc}$  in Table 1).

Therefore, future improvements in seismic hazard assessment necessitate a multi-magnitude approach. But this requires a deeper understanding of the physical relevance of the different magnitude scales, of their mutual relationship, as well as more “clean” complementary high quality data that have been measured over long time spans according to internationally agreed standards or otherwise clearly defined procedures, linked to easily accessible documentation published with unique nomenclatures that allows to unambiguously identify and correctly use such data.

The classical generic magnitudes are ML (Richter, 1935), Ms (Gutenberg, 1945a; since 1967 using the IASPEI standard Ms formula according to Vaněk et al., 1962), mB (Gutenberg 1945 b and c) and mb (Engdahl and Gunst, 1966), both using the Gutenberg and Richter (1956) calibration function for mB although mb as a more band-limited and high-frequency magnitude means a physically different thing. According to Figure 1 these “generic” magnitudes cover the range between 3 and 9 reasonably well. They are also the most frequently measured and at data centers collected magnitudes, with the exception of mB. Despite its fundamental importance (see Chapter 3, section 3.2.5.2) the determination of mB had been terminated in most parts of the world with the deployment of the narrowband instrumentation of the World-wide Seismic Standard Network (WWSSN) with peak magnifications between 1 and 2 Hz and 10 to 20 s, respectively (see Chapter 3, Fig. 3.20).

However, the procedures for measuring “generic” mb and Ms have not been so unique as global standards would have required for so widely used magnitudes. Different recording responses, measurement time windows after the P-onset and calibration functions were applied at different agencies or changed with time, resulting in increased data scatter, some systematic distance and/or magnitude-dependent biases and also in mixing of sometimes incompatible data. E.g., in the extreme case of the great Sumatra-Andaman Islands Mw9.3 earthquake of 26 December 2004 the following values for mb were reported by some of the leading seismological agencies: mb(IDC/CTBTO) = 5.7, mb(CENC, China) = 6.4, and mb(NEIC) = 7.2. Such differences for magnitudes of the same type and code name are not acceptable. Less dramatic but still significant are both distance- and magnitude-dependent biases up to 0.3-0.5 m.u. when using the IASPEI standard Ms formula, which had been derived for scaling A/T ratios in a wide range of periods between 2 and about 25 s, for scaling  $20 \pm 2$  s surface waves only. Facts, causes and the effects of lacking standards have been outlined and discussed in detail in Chapter 3.

Besides these classical magnitudes, which have been determined already for decades, mostly from analog records, others, such as the moment magnitude Mw and the energy magnitude Me, require digital broadband recordings and their spectral analysis or integration in the time-domain. Up to now they have been regularly determined only by a few specialized data centers. However, the broader use of these modern magnitude concepts is rapidly growing.

Short-comings of the current procedures to determine and annotate classical and also newly proposed magnitudes are, e.g.:

- Although already earlier IASPEI recommendations, published in the old Manual of Seismological Practice (Willmore, 1979), aimed at
  - expending and homogenizing magnitude measurements and nomenclature,
  - determining magnitudes from all seismic waves and component readings for which calibration functions are available,

- indicating the type of instruments and component on which the parameter readings (amplitudes, periods and/or duration) for a given magnitude value were made, these recommendations have become common practice only in a few countries;
- Body-wave magnitudes are nowadays determined from *vertical component* P-waves only although Gutenberg and Richter (1956) published body-wave calibration functions  $Q(\Delta, h)$  for both vertical and horizontal component readings of P and PP as well as for horizontal component readings of S;
- mb has been determined since the 1960s at “western” data centers and most stations and networks only from *short-period* recordings although the body-wave Q-functions have been derived mainly from medium- to long-period, more or less broadband recordings. Only some “eastern” countries, such as the former Soviet Union and its allies, or now in Russia, and in China, have seismograph networks continued to determine the classical Gutenberg and Richter (1956) medium-period broadband magnitude mB ( $m_B$ ) as well;
- Some other specialized international data centers such as the IDC of the CTBTO use for mb and Ms determination other than the common filter responses, measurement time-windows and/or calibration functions, resulting in the more or less strong incompatibility of their mb and/or Ms values (for comparison of mb(IDC) and mb(NEIC) see Murphy and Barker, 2003; Granville et al., 2002 and 2005; Bormann et al., 2007 and 2009);
- Currently still dominating “generic” magnitude nomenclature provides no hints to data users about such possible incompatibilities of magnitudes of supposedly the “same kind”;
- Calculating event mb averages, as done, e.g., at the ISC for many years from such incompatible NEIC and IDC mb data with an average (although magnitude-dependent) bias of about 0.4 m.u. is incorrect;
- Long-term continuity of classical standard magnitudes is a matter of high priority and necessitates proper scaling of modern magnitudes based on digital data with their forerunners that were based on analog data analysis. This has not always been done. Jumps in detection thresholds and catalog completeness due to unknown or not properly documented changes in measurement procedures may result in wrongly inferred changes of the relative frequency of occurrence of weaker and stronger earthquakes and be misinterpreted as changes in the seismic regime and the time-dependent seismic hazard (see case study for Southern California Ml by Hutton and Jones, 1993; also Habermann, 1995). This is not acceptable.

In order to overcome or at least mitigate these problems IASPEI established in 2001 within its Commission on Seismic Observation and Interpretation (CoSOI) a Working Group on Magnitude Measurements (in the following for short termed the WG). It has been entrusted with critically screening the most common procedures of amplitude measurement and magnitude determination practiced at seismic stations and various data centers and proposing measurement and nomenclature standards. Its members were: J. Dewey and P. Bormann (co-chairs), P. Firbas, S. Gregersen, A. Gusev, K. Klinge, B. Presgrave, L. Ruifeng, K. Veith, W.-Y. Kim, H. Patton, R. A. Uhrhammer, I. Gabsatarowa, J. Saul, and S. Wendt. The essential results are summarized below.

## 2 Seismic parameter data formats

The Willmore (1979) proposal for reporting magnitude parameter data was still based on the classical and costly telegraphic format. Modern seismic networks can provide far more information and low-cost e-mail transmissions and data exchanges via Internet have eliminated the restrictions and high costs of old telex messages. Consequently, as outlined in Chapter 10, most seismic parameter data have been stored and exchanged since at least 1990 in modern formats that are more complete, simpler and usually more transparent than the telegraphic format. But only rather recently agreement has been reached on some modern generally accepted standard formats for reporting parameter data.

A major step forward in this direction was made by the Group of Scientific Experts (GSE) organized by the United Nations Conference on Disarmament. It developed GSE/IMS formats (see Chapter 10, section 10.2.4) for exchanging parametric seismological data in tests of monitoring the Comprehensive Test Ban Treaty (CTBT). Seismological research, however, has a broader scope than the International Monitoring System (IMS) for the CTBT. Therefore, a new **IASPEI Seismic Format** (ISF; <http://www.isc.ac.uk/standards/#dataformats>), which is compatible with the IMS format but with essential extensions, has been developed and adopted by the Commission on Seismological Observation and Interpretation (CoSOI) of the International Association of Seismology and Physics of the Earth's Interior (IASPEI) at its meeting in Hanoi, August 2001. Examples for ISF compatible parameter data plots are given in the Information Sheet IS 10.2 and in Table 1 below. Both the International Seismological Centre (ISC) and the USGS National Earthquake Information Center (NEIC) present, exchange and archive their parameter data in the ISF format.

With respect to magnitude data, however, the ISF format limits the length of the magnitude names (nomenclature) to 5 letters/symbols. Further specification is however possible in combination with the related “amplitude phase name” which allows for more letters/symbols.

Since 2001 new developments have been introduced due to technology changes. The introduction of XML, i.e., an Extensible Markup Language, allows to present hierarchic structured data in form of [text](#) files and thus enables a platform and implementation independent exchange of data between computer systems, especially via [Internet](#). This again resulted in the requirement of a standard scheme for the representation of seismological parameter data. QuakeML, initiated and maintained by the Swiss Seismological Service, is such a collaborative effort. Detailed documentation and contact addresses are accessible via <https://quake.ethz.ch/quakeml/QuakeML>. The latest version 1.2 has been released only on 14 Febr. 2013 with the basic event descriptions mostly stable as of this writing but with the related metadata presentation still being under development.

QuakeML does not specify a magnitude or a magnitude name and its length. This is considered to be the duty of other standards. In QuakeML a magnitude name may be arbitrarily long. Rather, QuakeML permits to describe in detail how the amplitudes, periods, duration or other magnitude-relevant parameters have been measured by whom, when, where, and how, i.e., according to which procedure. QuakeML also allows to describe how the station and event magnitudes have been calculated from the measurement parameters, taking into account station counts and azimuthal gaps, the evaluation mode and statistics, how moment tensors have been determined etc. For details see sections 3.5.5 to 3.5.9 in Euchner and Schorlemmer (2013).

This latest development has to be closely followed up further because it might offer a comfortable and flexible complement to the limitations of the fixed ISF magnitude format for allowing the badly needed sufficiently detailed and unique description of the procedure of magnitude determination in the metadata.

### 3 IASPEI standard magnitudes

By May 2002, the WG reached the following first general understanding:

- the “generic” magnitudes ML, mb and Ms are most common, and related data for their determination are regularly reported by seismic stations and networks/arrays to international data centers;
- these “generic” magnitude names are widely accepted and applied by a diversity of user groups. Therefore, their **names should be kept when reporting magnitude data to a broader public.**
- this nomenclature is also considered to be adequate for many scientific communications, but **on the understanding that these magnitudes have been determined according to well established rules and procedures;**
- the WG realized, however, that different data producers make their related measurements for determining these magnitudes on records with different response characteristics and bandwidths, on different components and types of seismic waves and sometimes also use different period and measurement time windows. This increases data scatter, may produce baseline shifts and prevents long-term stable, unique and reproducible magnitude estimates that are in tune with original definitions, earlier practices and inter-magnitude conversion relationships;
- this situation is no longer acceptable, therefore, the WG felt a **need to introduce an obligatory more “specific” nomenclature** for reporting magnitude, amplitude (and related period) measurement data for databases and for use **in scientific correspondence and publications in which the ambiguity inherent in the “generic” nomenclature might cause misunderstanding;**
- the WG notes that the recently accepted IASPEI Seismic Format (ISF, see section 10.2.5 of Chapter 10) and the flexibility of internet data communication allow such specifications in nomenclature and even complementary remarks to be reported to international data centers, and to store and retrieve such data from modern relational databases.

Therefore, the WG proposed within its mandate:

- a) standard measurement procedures for amplitude-based magnitudes that agree as far as possible and reasonable with magnitudes of the same type that have been measured for decades from analog seismograms according to original definitions,
- b) that promote the best possible use of the advantages of digital data and processing,
- c) a unique nomenclature for these standard magnitudes and related amplitude data which is compatible with the IASPEI Seismic Format (ISF) and allows an unambiguous discrimination between standard and non-standard data;

- d) a questionnaire for dissemination (and deposition of the responses) by the ISC to seismic parameter data reporting stations and agencies on their specific procedures for recording and analyzing data relevant for magnitude calculation (see IASPEI, 2013: [http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf) and Annex 1 in IS 3.4)

with the main aim to

- e) minimize bulletin magnitude biases that result from procedure-dependent single-station or network magnitude biases;
- f) increase the number of seismological stations and networks with well-defined procedures;
- g) avail at the ISC as the final authoritative depository of the most complete and accurate seismological parameter data set also a directory of all standard and agency-specific procedures and nomenclatures;
- h) increase essentially the accuracy, representativeness, homogeneity and long-term global compatibility of magnitude data and their usefulness for seismic hazard assessment and research.

First recommendations for standard magnitude measurement procedures were published by IASPEI (2005). The latest update is posted under IASPEI (2013). The following standard types of magnitudes have been proposed and the procedures of their determination been specified. For more details, also on filter parameters and calibration functions, distance ranges of applicability, admissible deviations etc., see NMSOP-2, Chapter 3, sections 3.2.3.2, 3.2.4, 3.2.5 and IS 3.3:

- Local magnitude **ML** according to the Hutton and Boore (1987) formula measured on simulated Wood-Anderson records according to the parameter specifications of Uhrhammer and Collins (1990);
- Regional magnitude  $mb(Lg) = \mathbf{mb\_Lg}$  based on amplitude measurement of the Lg phase in a defined group-velocity window on WWSSN-SP simulated records;
- Short-period body-wave magnitude **mb** by measuring on simulated WWSSN-SP records the largest amplitude with periods  $< 3$  s in the whole P-wave train prior to PP and calibrating it with the Gutenberg and Richter (1956) calibration function for vertical component P waves;
- Broadband body-wave magnitude  $mB(BB) = \mathbf{mB\_BB}$  by measuring the largest velocity amplitude in the whole P-wave train on velocity-broadband records in a wide period range between 0.2 s and 30 s and calibrating it as for mb;
- Surface-wave magnitude  $Ms(20) = \mathbf{Ms\_20}$  by measuring the largest amplitude in the Rayleigh-wave train on simulated long-period WWSS-LP records at periods between 18 s and 22 s and calibrating A/T with the IASPEI standard Ms formula according to Vaněk et al. (1962) in the distance range between  $20^\circ$  and  $160^\circ$ ;
- Surface-wave magnitude  $Ms(BB) = \mathbf{Ms\_BB}$  by measuring the largest velocity amplitude in the Rayleigh-wave train on broadband velocity records in a wide range of periods between 3 s and 60 s in the wider distance range between  $2^\circ$  and  $160^\circ$  and calibrating them with the Vaněk et al. (1962) formula for Ms. This formula had in fact been developed on the basis of data in this wider range of period and distances.



With the exception of standard ML, which is measured on single horizontal components, not vector summed but each counted as an individual magnitude datum, all other standard magnitudes are measured on vertical component records only.

For international data exchange and archiving the magnitude names (nomenclature) should be written in the maximum 5-letter code (as above in **bold**) reserved in the ISF format for magnitude data.

In order to associate unambiguously the measured amplitudes and periods with the respective standard magnitudes the following standard nomenclature in the ISF format has been agreed upon for the so-called “amplitude phase names” (which may exceed 5 letters):

IAML, IAmb, IAmb\_Lg, IAMs\_20, IVmb\_BB and IVMs\_BB.

I stands for IASPEI International standard, A for displacement amplitude in units of nm and V for velocity amplitude in nm/s.

#### 4 Magnitude and amplitude nomenclature applied by the USGS-NEIC

The USGS/NEIC runs since 2006 its new fully automatic HYDRA location and analysis system. It measures routinely in experimental mode the amplitudes and periods related to all recently recommended IASPEI standard magnitudes according to the filter and measurement procedures outlined in IASPEI (2013) and IS 3.3. These data are presented in the event and station parameter plots according to the recommended standard nomenclature (see previous section) for amplitude phase names and related magnitudes (Table 1).

Additionally, HYDRA measures some agency-specific magnitudes, amongst them several versions of moment magnitude based on broadband body or surface waves (Mwb according to ???, Mwc according to ???, and Mwp according to ???), a time-domain, variable-period surface-wave magnitude Ms(VMAX) for application at regional and teleseismic distances according to a procedure proposed by Bonner et al. (2006) and a not specified average event magnitude M out of all these data (Table 1). All these magnitudes are given in the *maximum 5-letters* ISF format, which required to shorten, e.g., Ms(VMAX) to Ms\_VX.

**Table 1** Cut-out from an event and station parameter plot in IMS1.0/ISF parameter format produced by the USGS/NEIC HYDRA automatic location and analysis system.

```
BEGIN IMS1.0
MSG_TYPE DATA
MSG_ID 30000Z9N_43 HYDRA_ORANGE

DATA_TYPE BULLETIN IMS1.0:SHORT
The following is an UNCHECKED, FULLY AUTOMATIC LOCATION from the USGS/NEIC Hydra System
Event 15694
    Date      Time      Err    RMS Latitude Longitude  Smaj  Smin  Az  Depth
2009/09/29 17:48:13.26  2.94  1.45 -15.5267 -172.0703   6.5   5.8 133  28.4

Magnitude  Err Nsta Author      OrigID
Mb_Lg      6.0 0.5    1 NEIC      0
Ms_VX      8.2 0.1    23 NEIC      0
mb         7.2 0.0   243 NEIC      0
Mwp        7.8 0.0   179 NEIC      0
mB_BB      7.7 0.0   246 NEIC      0
Ms_BB      8.3 0.1   134 NEIC      0
Mwb        7.7 0.0    97 NEIC      0
```

|       |     |     |      |      |   |
|-------|-----|-----|------|------|---|
| Ms_20 | 8.0 | 0.0 | 161  | NEIC | 0 |
| Mwc   | 8.1 | 0.0 | 71   | NEIC | 0 |
| M     | 7.9 | 0.1 | 1064 | NEIC | 0 |
| Mwp   | 7.9 | 0.0 | 4    | PMR  | 0 |

| Sta  | Dist  | EvAz  | Phase   | Time         | Amp       | Per   | Magnitude | ArrID       |
|------|-------|-------|---------|--------------|-----------|-------|-----------|-------------|
| KNTN | 12.68 | 1.6   | IAmb_Lg | 17:55:00.090 | 7785.9    | 0.98  | Mb_Lg     | 6.0 BHZIU00 |
| KNTN | 12.68 | 1.6   | IVMs_BB | 17:56:03.398 | 2169276.1 | 10.00 | Ms_BB     | 7.7 LHZIU00 |
| TARA | 22.39 | 317.2 | P       | 17:53:11.829 |           |       |           | BHZIU00     |
| TARA | 22.39 | 317.2 | IAmb    | 17:53:45.055 | 12080.1   | 1.25  | mb        | 7.2 BHZIU00 |
| TARA | 22.39 | 317.2 | MMwp    | 17:53:53.279 | 1728974.9 | 41.45 | Mwp       | 7.6 BHZIU00 |
| TARA | 22.39 | 317.2 | IVmB_BB | 17:54:07.329 | 206688.6  | 5.40  | mB_BB     | 7.8 BHZIU00 |
| OUZ  | 23.45 | 210.6 | P       | 17:53:22.710 |           |       |           | HHZNZ10     |
| OUZ  | 23.45 | 210.6 | IAmb    | 17:53:47.450 | 11261.4   | 1.22  | mb        | 7.3 HHZNZ10 |
| OUZ  | 23.45 | 210.6 | IVmB_BB | 17:53:49.880 | 314772.2  | 9.74  | mB_BB     | 8.0 HHZNZ10 |
| OUZ  | 23.45 | 210.6 | IAMS_20 | 18:01:02.679 | 6314.2    | 20.00 | Ms_20     | 8.1 LHZNZ10 |
| OUZ  | 23.45 | 210.6 | IVMs_BB | 18:02:47.679 | 3821858.4 | 16.00 | Ms_BB     | 8.4 LHZNZ10 |
| OUZ  | 23.45 | 210.6 | AMS_VX  | 18:02:54.449 | 6326077.5 | 15.00 | Ms_VX     | 8.3 LHZNZ10 |

This way of an unambiguous presentation and annotation of magnitude and related amplitude data may serve as a good example for other seismological agencies as well, provided that the data users are somehow guided, e.g., via some explanatory pages that accompany data reports or bulletins and link to relevant documents/publications in which the respective procedures and - hopefully - also the relationship of this type of magnitude to other well established and widely used magnitudes are explained (see next section 5). Regrettably, the parameter data plot in Table 1 does not allow to say in which component the measurement has been made. One just has to know that all above amplitudes and magnitudes have been measured exclusively on the vertical component. However, standard ML should be measured on either the N or E or both components and any teleseismic S-wave amplitude or magnitude measurement, currently still greatly disregarded despite its undoubted scientific importance (see section 6), should preferably be made on horizontal components as well. Even the standard magnitude “amplitude phase names” do not yet account for a differentiation between vertical and horizontal component measurements.

## 5 Magnitude nomenclature reported to the ISC

The ISC (2013) Bulletin Summary for January-June 2010 includes the description of the upgraded new ISC procedures and the nomenclature for agency-specific magnitudes reported to the ISC. The latter are outlined in detail in Tables 9.6 and 9.7 of the ISC (2013) Bulletin Summary. For the period January-June 2010 **148 agencies contributed 29 different types of magnitudes**. This is due to the different seismological practices both at regional/national and global level in computing magnitudes from a multitude of combinations of wave types, period and amplitude measurement standards, distance ranges, instruments, components, algorithms etc. They are listed along with the agency codes and the number of earthquakes for which these magnitudes have been determined. Their main features are presented in tabulated form together with references to documents in which the procedure and/or data peculiarities are outlined. Table 2 is an abridged sample of the types of magnitudes contained in the ISC Bulletin for the time January-June 2010.

In Table 2 the proper IASPEI standard magnitude names according to the finally in 2011/2013 agreed nomenclature have not yet been used and also not yet been reported to the ISC. However, in a special investigation Bormann et al. (2009) compared for large event data sets in the wide magnitude range  $4 < M < 9$  all types of traditional Chinese magnitudes, measured at the China Earthquake Network Centre (CENC), with independently measured related IASPEI standard magnitudes. E.g., Ms7 in Table 2 is measured on vertical component

records with a WWSSN-LP response, as the standard  $M_s(20)$ . But, in contrast to the latter, which is measured only at periods between 18 s and 22 s and in the distance range  $20^\circ$ - $160^\circ$ ,  $M_s7$  is based on surface-wave displacement amplitude maxima with periods  $T > 6$  s measured in the distance range  $3^\circ$ - $177^\circ$ . Nevertheless, independent  $M_s7$  and  $M_s(20)$  for the same events agreed almost perfectly within the tolerance limit of 0.1 m.u. set by the WG. The orthogonal regression relationship is  $M_s7 = 0.99M_{s\_20} + 0.14$  with a standard deviation of  $\pm 0.20$  m.u.. The latter is typical for high-quality  $M_s$  determinations. Thus,  $M_s7$  and  $M_{s\_20}$  are fully compatible and new  $M_{s\_20}$  could be merged with older  $M_s7$  data without any loss in quality or increase in data scatter, which means assured data continuity.

**Table 2** Examples of magnitudes reported to the ISC by data contributing seismological stations or agencies for the Summary of the ISC Bulletin January-June 2010 (ISC, 2013). All references cited in the table, and agency abbreviations, are given in ISC (2013) but those typed in *Italics* also in the reference list to this IS.

| Magnitude type | Description                                                      | References                                                         | Comments                                                                     |
|----------------|------------------------------------------------------------------|--------------------------------------------------------------------|------------------------------------------------------------------------------|
| M              | Unspecified magnitude type                                       |                                                                    | Often related to real or near-real time magnitude estimations                |
| mB             | Broad-band body-wave magnitude                                   | <i>Gutenberg (1945b,c); IASPEI (2005); Bormann and Saul (2008)</i> |                                                                              |
| mb             | Short-period body-wave magnitude                                 | <i>IASPEI (2005)</i>                                               | Classical mb computed from station between $20^\circ$ - $100^\circ$ distance |
| mb1            | Short-period body-wave magnitude                                 | IDC (1999) and references therein                                  | Reported only from IDC; it includes also station below $20^\circ$ distance   |
| mb1mx          | Maximum likelihood short-period body-wave magnitude              | Ringdal (1976); IDC (1999) and references therein                  | Reported only from IDC                                                       |
| mbtmp          | short-period body-wave magnitude with depth fixed at the surface | IDC (1999) and references therein                                  | Reported only from IDC                                                       |
| mbLg           | Lg-wave magnitude                                                | <i>Nuttli (1973); IASPEI (2005)</i>                                | Reported only by MDD and by other North American agencies as MN              |
| Mc             | Coda magnitude                                                   |                                                                    |                                                                              |
| MD (Md)        | Duration magnitude                                               | <i>Bisztricsany (1958); Lee et al. (1972)</i>                      |                                                                              |
| ME (Me)        | Energy magnitude                                                 | <i>Choy and Boatwright (1995)</i>                                  | Reported only by NEIC                                                        |
| MJMA           | JMA magnitude                                                    | <i>Tsuboi (1954)</i>                                               | Reported only by JMA                                                         |
| ML (MI)        | Local (Richter) magnitude                                        | <i>Richter (1935); Hutton and Boore (1987)</i>                     |                                                                              |
| MLSn           | Local magnitude calculated off $S_n$ phases                      | Balfour et al. (2008)                                              | Reported by PGC only for earthquakes west of the Cascadia subduction zone    |

|         |                                                                |                                                              |                                                                                                                                 |
|---------|----------------------------------------------------------------|--------------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------|
| MLv     | Local (Richter) magnitude computed from the vertical component |                                                              | Reported only by DJA and BKK                                                                                                    |
| MN (Mn) | Lg-wave magnitude                                              | <i>Nuttli (1973); IASPEI (2005)</i>                          | Also reported as mbLg from MDD                                                                                                  |
| MS (Ms) | Surface wave magnitude                                         | <i>Gutenberg (1945a); Vaněk et al. (1962); IASPEI (2005)</i> | Classical surface wave magnitude computed from station between 20°-160°                                                         |
| Ms1     | Surface wave magnitude including                               | IDC (1999) and references therein                            | Reported only by IDC; it includes also stations below 20° distance                                                              |
| mslmx   | Maximum likelihood surface wave magnitude                      | Ringdal (1976); IDC (1999) and references therein            | Reported only by IDC                                                                                                            |
| Ms7     | Surface wave magnitude                                         | <i>Bormann et al. (2007)</i>                                 | Reported only by BJI and measured in the distance range 3°-177° on records of a Chinese-made seismograph with WWSSN-LP response |
| MW (Mw) | Moment magnitude                                               | <i>Kanamori (1977); Dziewonski et al. (1981)</i>             |                                                                                                                                 |
| Mw(mB)  | Proxy Mw based on mB                                           | <i>Bormann and Saul (2008)</i>                               | Reported only by DJA and BKK                                                                                                    |
| Mwp     | Moment magnitude from P-waves                                  | <i>Tsuboi et al. (1995)</i>                                  | Reported only by DJA and BKK and used for rapid response purposes                                                               |

Not as perfect is the agreement between Ms(CENC) and Ms(20), measured at different components, period ranges, record responses and with a constant difference in the calibration function ( $\Delta M = 0.25$  m.u.). And the average differences between other combinations of similar types of magnitude measurement depend on magnitude, ranging from small to large magnitudes for mB(CENC) and mB(BB) from  $\Delta M = -0.3$  to  $0.0$  m.u., for Ms(CENC) and Ms(BB) from  $\Delta M = 0.0$  to  $+0.4$  m.u., and for mb(CENC), which is measured only within the first 6 s after the P-wave arrival, the difference to standard mb may reach at the largest earthquakes even  $-1$  m.u. It is just this kind of procedure-dependent data incompatibilities in international data exchange, which the measurement standards aim to eliminate, thus reducing data scatter and improving the global compatibility of magnitude estimates.

This experience from China shows that it is highly desirable that seismic stations/agencies reporting data to the ISC or to other regional or global seismological data centers compare - in conjunction with the introduction of the new IASPEI standard magnitudes - for a sufficiently large and representative set of seismic records from events in a wide range of magnitudes and distances the results of their traditional magnitude procedures with those according to the new IASPEI standard procedures. A suitable representative test data set can be downloaded via IS 3.4, which gives also guidelines for carrying out such a test by using the SEISAN software. If the average absolute differences turn out to be within  $0.1$  m.u. then the traditional procedure yields in fact IASPEI standard compatible results.

Sometimes this may be the case even if some procedural details, such as the applied filter responses, differ significantly. An example is given in IS 3.3 (Figures 10 and 14) comparing

Ms<sub>20</sub> measured on either SRO-LP or WWSSN-LP responses. Analysis software which has an SRO-LP simulation filter implemented, such as Seismic Handler Motif (SHM), could then be used as an even better alternative for Ms<sub>20</sub> determination, because the measured A and T are within acceptable limits of measurement accuracy identical with those measured on WWSSN-LP records and can thus be reported with the standard amplitude nomenclature IAMs<sub>20</sub>. But because of the generally better SNR of SRO-LP records one is able to measure even more surface-wave magnitudes, especially from weaker events. Moreover, also automatic A and T measurements are less error-prone because of the less noisy and thus simpler waveforms in SRO-LP filtered records (see Figure 13 b in IS 3.3). This, however, is not a rule. In other cases of magnitude measurements in other frequency ranges and on other seismic phases already smaller changes in recommended measurement parameters may result in much larger magnitude differences.

In any event, the ISC as the most representative, complete and accurate final depositary of seismological parameter data should be informed by all its data contributing stations/agencies about the details of their procedures for recording, measuring and calibrating their magnitude data. In order to assure sufficient completeness and compatibility of provided information, the IASPEI/CoSOI WG has developed a questionnaire which is annexed to both IASPEI (2013) and (as Annex 2) to IS 3.4. Based on such information and the results of comparative measurements with traditional agency and new IASPEI standard procedures the number of agency specific entries in Table 2 is likely to shrink significantly with the global introduction of the IASPEI (2013) measurement standards.

## 6 Proposal for a more generalized unique nomenclature for amplitude and magnitude data

In order to assure future IASPEI-authorized standards annotation and reporting of measurements for amplitude-based seismic magnitudes, the WG on magnitude measurements had agreed already a decade ago on the suitability of a more general systematic nomenclature for presenting amplitude and magnitude data in a unique manner (see IS 3.2 in NMSOP-1, in Bormann, 2002 and 2009). It is based on recommendations of the Sub-Commission on Magnitudes of the Commission on Practice made at the Joint General Assembly of the IASPEI/IAVCEI Meeting in Durham 1977. These recommendations aimed at measuring and reporting magnitudes in an unambiguous manner for all seismic phases and from all component readings for which magnitude calibration functions according Richter (1935, 1958), Gutenberg and Richter (1956) and Vaněk et al. (1962) were available (see first bullet-point paragraph on page 4). The reason that Gutenberg and Richter (1956) proposed besides P waves also PP and S for magnitude determination and also amplitude measurements from vertical and horizontal components were that

- medium to long-period PP is usually the largest longitudinal wave in the core shadow of P beyond 100°, up to 180°, allowing more reliable mB event estimates, based on more stations than with P waves alone;
- medium to long-period S is usually the largest secondary body-wave phase after P with amplitudes much larger than that of P and with a radiation pattern about 45° off that of P (see section 3.4.1 of Chapter 3). This again assured more independent magnitude data and, when averaged with those from P wave readings, data that are much less affected by the radiation pattern of the seismic source, which may in the

case of only a few stations with P-wave readings and insufficient azimuthal coverage produce rather biased magnitude estimates;

- teleseismic S-waves have generally their largest amplitudes in the horizontal components, and horizontal PP amplitudes are also relatively large or even larger than those of P, and, because of different take-off angle, also source mechanism dependent. Moreover, at Gutenberg and Richter's times stable long-period vertical component recordings were not yet as commonly available as nowadays.

During recent decades the virtual global seismic network of high-resolution stable broadband recordings has grown tremendously. Therefore, an insufficient number of stations with potential amplitude readings and the need for independent amplitude readings from PP and S has become much less of a problem for a reliable magnitude determination than in the past. However, S-wave amplitude, period and/or ground velocity readings and reporting to the international data centers are still grossly neglected at many observatories/networks (see Chapter 1) although they are highly desirable for improving the shear-wave attenuation model of the Earth down to shorter periods and thus with higher spatial resolution and with larger penetration depths than possible with surface waves. But when measuring S amplitudes they could be used for additional S-wave mB estimates as well with the advantage of reducing possible P-wave station mB biases in the case of inappropriate azimuthal station coverage. But then it would be desirable to clearly differentiate between mB based on P, S and/or PP.

For the original proposal of a more systematic unique magnitude nomenclature see Willmore (1979; p. 124)) and its extension in IS 3.2 of NMSOP-1 (Bormann, 2002 and 2009. An example along these lines has been given for a strong Mongolia earthquake in Tab. 3.1 of Chapter 3.

In view of the now reached agreement on standard measurement procedures and a unique nomenclature for 6 widely determined types of magnitude as well as the general trend to measure related amplitudes, with the exception for ML, on vertical component records only, this proposal may at present appear to be obsolete. In section 5 of this IS it has also been shown that agencies or stations reporting magnitudes to international data centers or publishing them in their bulletins may define their own ISF format compatible magnitude names. They could be unique and unambiguous provided that they are accompanied, as in Table 2, by references to publications/documents which describe in detail the procedures for recording, filtering, measuring, and calibrating these magnitudes, in a similar way as the IASPEI standard magnitudes have been defined in IASPEI (2013) or IS 3.3. Moreover, it has to be assured that other data producers that apply the same procedures use also the same nomenclature. If handled this way then it is indeed the best way to keep data centers and external data users fully informed. This avoids mixing incompatible data and biased conclusions. Precondition is, however, that all data producers really make their agency-specific ways of magnitude determination, hopefully together with some information on how there magnitudes relate to other well established and/or IASPEI standard magnitudes of the same type, publicly available via their national or regional/global data centers (for questionnaire see references in section 5).

Nevertheless, we will give below in Table 3 a few examples for another way of designing unique magnitude names, also for other than the now common phases, components, record responses and calibration functions. It is based on the earlier version presented in I.S 3.2 of NMSOP-1. However, the current upgrade allows also to differentiate between magnitudes

based on displacement amplitudes  $A$  and broadband velocity amplitudes  $V$  and the instrument code is no longer put in brackets. Moreover, since  $V$  stands now for velocity the vertical component code is now  $Z$ , as in the Gutenberg-Richter (1956) calibrations diagrams.

Further, one has to know that Willmore (1979) presents in Fig. 1.1 the following standard response classes for analog seismographs: Type A (short-period, A1-A4), B (long-period, B1-B3), C (displacement broadband Kirnos), D (velocity broadband 1-100 s), and E (classical Galitzin) besides the classical Wood-Anderson (WA) response. These responses still resemble widely used responses of modern digital records or of common synthetic filter outputs derived therefrom (e.g., A1 = Benioff-SP, A2 = WWSSN-SP, A3 = French APX system, A4 = Russian SKM-3 seismograph; B1 = WWSSN-LP, B2 = French system, B3 = high-gain HGLP system; C = Kirnos seismograph). The latter has two widely used versions: C1 = SK - corner period of 10 s and C2 = SKD – 25 s, with C1 being a standard filter at CENC for Chinese mB and Ms measurement and C2 being a standard filter in SHM and Pitsa software because of the superiority of C2 records in seismic phase recognition (see Chapter 11). And the Willmore (1979) standard response D covers the main period range of modern velocity broadband seismographs, which is between 1 and 100 s. Therefore, we use these class symbols in order to illustrate the filter/seismograph type specification in the proposed generalized magnitude nomenclature. It may be replaced or complemented by any other unique specified filter/seismograph response symbol. Along these lines we propose in Table 3 just a few magnitude names for readings of maximum amplitudes of different seismic phases made on horizontal or vertical component records of such basic standard types of seismographs/filters and calibration functions and compare them with respective “generic” magnitude names. As a general unique magnitude name we propose **MXYFC** with

- **M** being the general symbol for magnitude.;
- **X** the symbol name of the seismic phase on which the amplitude and periods for the respective magnitude have been measured, e.g., P, S, PP, Lg, PKP;
- **Y** the component on which amplitude and period have been measured, e.g., N for the north-south, E for the east-west, Z for the vertical and H – for the vectorially combined horizontal component;
- **F** the symbol for the filter/seismograph response on the record of which the amplitude and period have been measured;
- **C** the symbol for the calibration function which has been applied to calculate the respective magnitude based on the measure ground motion amplitudes and periods.

However, in the case that the considered magnitudes have been calculated on the basis of other measurement parameters or procedures such as duration, spectra, waveform fitting and inversion, tsunami run-up or wave height measurements, the MXYFC nomenclature is not appropriate and should better be replaced by MAG (MAgency) or MAU (MAuthor) with reference being available for full procedural details.

Along these lines a few examples are presented in Table 3, using the following additional abbreviations:

A-E seismograph/filter type classes according to their definition in Willmore (1979) and outlined above, with a number added if there are several common types within a given class. E.g., one could give the NEIC PDE filter and the IDC/CTBTO filter used for short-period parameter readings, which differ from A1-A4 (see Figure 3 in IS 3.3) the numbers A5 and A6, respectively.

|          |                                                                                                                                                                                                                                                                                                                                                                       |
|----------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| WA       | Wood-Anderson horizontal-component seismograph with displacement-proportional response in the period range 0.1 to 0.8 s;                                                                                                                                                                                                                                              |
| RO       | stands for the long-period Seismological Research Observatory (SRO-LP) response (see Figure 10 in IS 3.3);                                                                                                                                                                                                                                                            |
| $\Delta$ | epicentral distance as commonly used in calibration functions;                                                                                                                                                                                                                                                                                                        |
| h        | hypocentral depth;                                                                                                                                                                                                                                                                                                                                                    |
| L        | largest (relatively) long-period surface wave                                                                                                                                                                                                                                                                                                                         |
| Q        | at the end of the magnitude name stands for the respective Gutenberg and Richter (1956) calibration function $Q(\Delta, h)$ for either PZ, PH, PPZ, PPH, SH (see Figures 1a-c and Table 7 in DS 3.1 and for standard PZ also in tabulated form for all source depths with program description in PD 3.1);                                                             |
| $\sigma$ | at the end of the magnitude name stands for the Prague-Moscow (Vaněk et al., 1962, or Karnik et al., 1962) calibration function $\sigma(\Delta)$ for surface-wave readings of both LZ and LH; recommended as standard by IASPEI and used at both the NEIC and the ISC (in tabulated form as Table 4 in DS 3.1; as standard formula see Chapter 3, Eqs. (3.35 or 3.36) |
| C        | Stands for any other specific calibration function, e.g., RI for Richter (1958), HB for Hutton and Boore (1987) or VC for the Veith and Clawson (1972) calibration function $P(\Delta, h)$ used at the IDC/DTBTO (for details see DS 3.1).                                                                                                                            |

For magnitudes that have been determined from records of seismographs with other response characteristics than the standards A to D or WA and/or by using calibration functions other than  $\sigma(\Delta)$ ,  $Q(\Delta, h)$  or local scales properly linked to the original Richter  $M_L$  (ML) scale, this should be specified by giving filter type F and C in brackets, i.e.,  $M(F; C)$ , or by adding a complementary comment line with the name of the relevant author/institution or with a link to proper reference and documentation.

Amplitude data for identified seismic phases should be specified and reported to data centers in the following general format for non-IASPEI standard “amplitude phase names”: **AXYF** or **VXYF**:

with A = displacement amplitude in nm or V = velocity amplitude in nm/s  
and X, Y and F as defined above, e.g.:

- ASHB2 – horizontal component S-wave displacement amplitude on a WWSSN-LP record;
- APPZC2 – vertical component PP-wave displacement amplitude on a Kirnos SKD record;
- VPZD – vertical component P-wave velocity amplitude on type D record.

In the case that phase name, component, filter type and calibration function can be uniquely described by just one letter, then the general MXYFC nomenclature would perfectly match with the ISF 5-letter magnitude name format. This, however, is not possible for more than single letter phase names such as Lg, PP or PKP, in the case of using more specific responses within a standard response class (such as A1 or C2) and/or calibration functions that require more than one letter to describe them unambiguously. A few such examples have been added to Table 3 too, typed in *Italics*.



This dilemma remains as long as the most relevant current seismic standard data formats limit the length of magnitude names to 5 letters. A future extension might therefore be considered by IASPEI/CoSOI. As long as this is not the case, however, we can only recommend, that the component information and any specification within a considered response class is conveyed by the related magnitude “amplitude phase name”, as shown above. The type of calibration function used for calculating a specific type of magnitude from a compatible set of measured parameter data, however, has then to be documented in the metadata and/or in related generally accessible documentation as it has been done for the current IASPEI standard magnitudes. The same holds if some agency decides to measure amplitudes and periods not at the maximum of the respective phase waveform but within an arbitrarily fixed measurement time window after the phase first arrival, such as the 5.5 s window set by the International Data Centre (IDC) of the Comprehensive Test-Ban Treaty Organization (CTBTO). Such a combination of magnitude and amplitude phase name nomenclature or M(Agency) specification would limit also for non-standard magnitudes their names to match the ISF 5-letters magnitude code, e.g.:

- MPA in conjunction with APZA5 (previous NEIC PDE mb)
- MPPD in conjunction with VPPHD;
- MPPC in conjunction with APPZC2;
- MLgA in conjunction with ALgZA2;
- MPIDC with reference to procedure definition (e.g., section 17.4.2 of Chapter 17).

**Table 3** Preliminary proposal for specific magnitudes not measured and/or scaled according the recommended IASPEI (2013) standard and the equivalent “generic” magnitude names. The ambiguity of generic magnitude names is obvious.

| Specific                                   | Generic | Description                                                                                                                                                                                                                                                                                         |
|--------------------------------------------|---------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>MPZAQ</b>                               | mb      | P-wave magnitude from short-period narrowband vertical component recordings of type A calibrated with $Q(\Delta, h)$ for PZ; <b>equivalent to standard mb if A2 = WWSSN-SP response</b> and standard measurement criteria are observed.                                                             |
| <b>MPHCQ</b>                               | mB      | P-wave magnitude from medium-period (more broadband) recordings calibrated with $Q(\Delta, h)$ for PH.                                                                                                                                                                                              |
| <b>MPZDQ</b>                               | mB      | P-wave magnitude from velocity broadband recordings calibrated with $Q(\Delta, h)$ for PZ; <b>equivalent to mB_BB</b> if standard measurement criteria are observed.                                                                                                                                |
| <b>MSHCQ</b>                               | mB      | S-wave magnitude from medium-period (more broadband) recordings calibrated with $Q(\Delta, h)$ for SH.                                                                                                                                                                                              |
| <b>MLZD<math>\sigma</math></b>             | Ms      | Surface-wave magnitude from L readings in vertical component broadband velocity records of type D, calibrated with the IASPEI standard “Prague-Moscow” formula $\sigma(\Delta)$ (cf. Eq. 3.35, or 3.36 in Chapter 3); <b>equivalent to Ms_BB</b> if amplitudes are read in the period range 3-60 s. |
| Mw(Author/Agency),<br>e.g.<br><b>MwCMT</b> | Mw      | Non-saturating moment magnitude based on the zero-frequency plateau of the displacement spectrum or other related estimates via signal analysis/inversion in the time domain, e.g., of (Global) Centroid Moment Tensor Project.                                                                     |

|                                                                          |    |                                                                                                                                                                                                                                                                                                     |
|--------------------------------------------------------------------------|----|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Me(Author/Agency)<br>e.g.,<br><b>MeGS</b> ,<br><b>MeGFZ</b>              | Me | Energy magnitude as defined/derived by specific author(s) or institution(s), e.g.<br>Me of the US Geological Survey (Choy and Boatwright, 1995)<br>Me of the GFZ German Research Centre for Geosciences (Di Giacomo et al., 2008)                                                                   |
| Mt(Author), e.g.<br><b>MtABE</b>                                         | Mt | Tsunami magnitude as defined/derived by specific author(s) or institutions, e.g. Mt according to Abe (1989)                                                                                                                                                                                         |
| <i>MPZA6VC</i>                                                           | mb | P-wave magnitude measured on short-period vertical component recordings with the CTBTO/IDC narrowband velocity band-pass filter (here abridged as A6), within 5.5 s after the P-wave first onset and calibrated with Veith and Clawson (VC; 1972) calibration curve $P(\Delta, h)$ .                |
| <i>MPPHCQ</i>                                                            | mB | PP-wave magnitude from medium-period (more broadband) recordings calibrated with $Q(\Delta, h)$ for PPH.                                                                                                                                                                                            |
| <i>MLZB1<math>\sigma</math></i><br>or<br><i>MLZRO<math>\sigma</math></i> | Ms | Rayleigh surface-wave magnitude from L readings in vertical component records of type <b>B1 = WWSSN-LP</b> , or SRO-LP, calibrated with the IASPEI standard $\sigma(\Delta)$ (cf. Eq. 3.35, or 3.36 in Chapter 3).<br><b>Equivalent with standard Ms_20</b> if measured at periods between 18-22 s. |

In the case that amplitudes have been measured on phases, components and with procedures according to the respective IASPEI magnitude standards (see section 2) then the IASPEI amplitude phase name nomenclature should be used instead, e.g., IAMB\_Lg or IVMs\_BB. For non-standard agency or author specific magnitudes, however, one should avoid any confusion with traditional generic or new standard magnitudes and generally use capital M as the unique magnitude symbol, followed by phase, component, filter and calibration function symbol as in Tab. 3, or, if this exceeds the 5-letters code, by the proposed abridge version MXF in conjunction with the respective more detailed amplitude phase names A(orV)XFn and the calibration function C specified in the metadata.

## 7 Author or agency specific magnitude nomenclature

Classical magnitudes based on not strictly standardized procedures still form the majority of available magnitude data for quite some time. This particularly applies to the three “generic” types of magnitudes, ML, mb and Ms. The majority of countries, using ML, have not yet been able to derive their own local/regional calibration functions that are properly scaled and tested with respect to the IASPEI standard. mb data, measured up to now (2013) and contained in national and international data bases, are still more or less a mixture of data measured with slightly different instrument responses within differently fixed or flexible measurement time windows. And also Ms data may differ by up to several tenths of magnitude units depending on the specific Ms procedure applied. Therefore, it is important that the new standards now clearly differentiate between Ms\_20 and Ms\_BB or any other type of Ms. **Any future use of the undifferentiated generic magnitude symbol Ms would be highly counter-productive**, not so much in publications to a broader public when accuracy matters less but **in any scientific publication and data base with relevance for future research**.

Regrettably, such a clear distinction between old (still mixed and thus more noisy or even incompatible) and new (standardized) ML and mb data is not yet possible because the IASPEI Working Group was unable to agree on specified non-generic IASPEI standard names for ML and mb. This persisting ambiguity between old generic and new standard magnitude data symbols may confuse in future users of both old and new data with respect to their compatibility, unless the distinction can be made on the basis of the related standard amplitude phase names IAML and IAMB for standard ML and mb in contrast to AML and Amb for non-standard versions of these two magnitudes.

For ML data that are used for relevant studies only within the considered regions and less frequent, if ever, for interregional comparison of compatibility (an exception see section 3.2.9.6 in Chapter 3), the generic ML nomenclature may be less of a problem as long as no interregional and border crossing seismic hazard assessment is based on such data. In order to reduce uncertainties in this respect we have presented in Table 2 of DS 3.1 MI formulas for 16 different regions world-wide together with information on their calibration to either the traditional Richter (1935) scale and Wood-Anderson records or to the new IASPEI standard formula (3.15) in Chapter 3 when using synthetic WA records. Moreover, in both NMSOP-1 (2002 and 2009) and NMSOP-2 (2012/13) we deliberately **use the more general symbol MI for local magnitudes. It may be changed to ML when the related formula for the region under consideration has been properly scaled to the IASPEI standard formula and procedure.** If not, it should be considered and also denoted as an area- or agency-specific formula MIAG if it is likely to produce average absolute differences  $> 0.1$  m.u. as compared to formula (3.15) in Chapter 3.

The multitude of currently used agency-specific magnitudes may be very confusing for users. Systematic inter-comparison between these formulas and the new IASPEI standards offers the chance to streamline, homogenize and reduce their number by agreeing on the circumstances under which these magnitudes have the same values as their IASPEI counterparts. In fact, in most cases a specific magnitude scale will agree on average with another magnitude scale at some appropriate value, e.g. local scales at some measured input value at a suitable hypocentral distance between about 17 km (Hutton and Boore, 1987) to 100 km (Richter, 1935) (see Fig. 3.30 in Chapter 3), the teleseismic scales mB, Ms and Mw at a common average magnitude value, which is around 6.5 (Fig. 3.70 in Chapter 3; see Utsu, 2002), or the modern moment and energy magnitude scales, Me and Mw, which yield identical values at a constant ratio of released seismic energy to seismic moment (see Chapter 3, section 3.3.3 and Bormann and Di Giacomo, 2011). At other than these scaling values, however, they may differ significantly, depending on local/regional differences in wave attenuation, the difference in the periods and bandwidth at which these magnitudes are measured and/or the difference in the  $E_s/M_0$  ratio of the actual rupture process.

The required uniqueness of magnitude data can principally be assured and documented by a rather short specific magnitude names within the 5-letters ISF code, as in Table 2, yet necessitating a defining documentation in the metadata, which will – hopefully in future – also be complemented by some statistics of the authors/agencies about average absolute or mean/median and standard deviations or trend differences with respect to standard magnitudes, moment magnitudes or other magnitudes of the same type. However, **if it can be proven that agency-specific deviations in procedural details from the recommended standard procedures result in magnitude values that differ in terms of absolute average deviations not more than 0.1 m.u. from the respective standard values then they can be considered and reported as standard magnitudes as well.** Examples are given in section

3.2.3.2 of Chapter 3, in section 5 of IS 3.3 with Figure 14 and in section 5 above with the discussion in conjunction with Table 2. Guidelines with record data for such comparative measurements are given in IS 3.4 together with questionnaires for reporting the results of such a comparison as well as the agency-specific procedures for magnitude measurements to the International Seismological Center (ISC).

New procedures for magnitude measurement are continuously proposed, such as non-saturating broadband P-wave magnitude estimators which take additionally rupture duration into account [e.g., mBc by Bormann and Wylegalla (2005) and Bormann and Saul (2009); the unspecified M by Hara (2007); Mw<sub>pd</sub> by Lomax and Michelini (2009)], W-phase-based estimates of Mw and moment tensor solution estimates based on the analysis of W phases [see Kanamori and Rivera (2008), Hayes et al. (2009); Chapter 3, section 3.2.8.6] or magnitude estimates based on strong-motion data. Generally, new or modified classical procedures of magnitude determination should be clearly distinguishable from any common “generic” or standardized magnitude and be given unique specific names. Some general rules are recommended:

- If authors or institutions have given to their new magnitudes already unique names with 5 or less letters (such as mBc or Mw<sub>pd</sub>) these names should be preserved in national or international data bases, bulletins or publications in agreement with the ISF 5-letters magnitude code and links be given to relevant publications for user information;
- If such magnitudes are unspecified, then a code should be added that permits the unambiguous identification of the author, agency and type of magnitude, preferably within the ISF 5-letters restriction, e.g. MHARA or MwWPH for the W-PHase derived Mw, otherwise in an extended form as the related amplitude phase name or as accompanying metadata;
- If author/agency proposed names exceed the ISF 5-letters code one could also try to shorten it in a still comprehensible and unambiguous way, as, e.g., Ms(VMAX) was shortened into Ms\_VX in the NEIC Hydra parameter data plots;
- If possible these modifications should be agreed upon between the data centers and authors/agencies so as to assure their unique use also in publications.

## **8 The role of the ISC in implementing IASPEI standard magnitude measurements and assuring unique amplitude and magnitude nomenclature**

The WG on Magnitude measurements was set up by the IASPEI Commission on Seismic Observation and Interpretation (CoSOI) in response to a IASPEI resolution submitted in 2001 by the ISC Governing Council under the chairmanship of A. Dziewonski. Its mandate has been outlined on pages 6 and 7 above. The goals a-d have been achieved with all related documentation on the procedures, standard nomenclature, comparison with non-standard magnitudes, questionnaire on agency-specific procedures etc. being now available, including publications in peer-reviewed journals and related NMSOP-2 items [e.g., Liu et al. (2005; 2006); Bormann et al. (2007; 2009); Bormann and Saul (2008); NMSOP-2, Chapter 3, section 3.2, IS 3.3 and 3.4, DS 3.1; <http://nmsop.gfz-potsdam.de>.]

With respect to **goal a)** we can now be sure that  $mB\_BB$  is very close to the original Gutenberg (1945b) and Gutenberg and Richter (1956) body-wave magnitude  $m = mB$ , deviating in its original range of definition between  $5.5 < mB < 8.5$  less than 0.1 m.u. (see Bormann et al., 2009). The same applies when comparing  $Ms\_20$  with the classical Gutenberg (1945a)  $Ms$  and the traditional NEIC vertical component surface-wave magnitude measured around 20 s (Bormann et al., 2009; Lienkaemper, 1984).

However,  $Ms\_BB$  is the only surface-wave magnitude that agrees **perfectly** with the  $Ms$  formula proposed by Vaněk et al. (1962) which is the official IASPEI standard  $Ms$  formula since 1967. Being perfectly scaled to the original Gutenberg  $Ms$  (around 20s) that is almost exclusively based on event magnitudes  $\geq 6 - 8+$ ,  $Ms\_BB$  yields however on average for magnitudes  $3 < Ms\_20 < 6$  between 0.4 and 0.1 m.u. larger values than  $Ms\_20$ . The simple reason is that the Rayleigh wave maxima for these smaller earthquakes, which are preferentially recorded at local to regional distances, rarely occur at periods around 20 s (see Bormann et al., 2009 and Figs. 3.35 and 3.40 in Chapter 3). This also explains the systematic distance-dependent errors found by Herak and Herak (1985) and Rezapour and Pearce (1993) when  $Ms$ , calculated with the IASPEI standard formula, is based exclusively on amplitudes measured in the period range between 18 and 22s.  $Ms\_BB$  also reduces by about  $\frac{1}{2}$  the systematic bias of  $Ms\_20$  when compared with  $Mw$  for  $Mw < \approx 6.5$  (Kanamori and Anderson, 1975; Ekström and Dziewonski, 1988; Di Giacomo et al., 2013) as well as the difference between  $Ms$  with  $mb$  and  $ML$  for  $Mw < 6$  (see Chapter 3, Fig. 3.70). Since  $Ms\_BB$  may be measured down to epicentral distances of  $2^\circ$  and measured at periods as small as 3 s, it agrees in fact rather well with  $ML$  at regional distances (see Figure 5, upper right in Bormann et al., 2007, comparing  $ML$  with traditional Chinese  $Ms$  that resembles  $Ms\_BB$ ).

$ML$  (Richter, 1935, 1958),  $mb$  (Engdahl and Gunst, 1966), and  $mb(Lg)$  (Nuttli, 1973, 1985, 1986, 1988) are classical US derived magnitudes. The US members in the WG have taken care of assuring that the related new standards are compatible with these original definitions within about 0.1 m.u.

With respect to **goal b)** it has been proven by implementing the new standards both in fully automatic procedures at the NEIC, in the GFZ-GEOFON SeisComp3 real-time procedures (for  $mb$  and  $mB\_BB$ ) as well as in comprehensive off-line interactive data analyses at the China Earthquake Network Center (CENC) and at the Collm (CLL) observatory in Germany that the new standards allow to make best use of the advantages of digital data and processing. Moreover, the new broadband standard magnitudes  $mB\_BB$  and  $Ms\_BB$ , that are based on just the single measurement parameter  $V_{max}$  instead of  $A_{max}$  and  $T$ , are even more robust and have a smaller data scatter than the band-limited  $mb$  and  $Ms\_20$  (see section 3.2.3.2 in Chapter 3 and Table 5 in IS 3.3).

With respect to **goal c)** the new nomenclatures for standard magnitudes and related amplitudes are now available but have been implemented as of May 2013 only in the NEIC Hydra event data plot and in the SEISAN analysis software record plots for magnitude measurements. They are not yet available in the bulletins of the ISC, the NEIC or other agencies and stations.

Concerning **goal d)** the ISC had circulated already years ago the first versions of the IASPEI (2005) WG recommendations and of the questionnaire on current agency-specific magnitude procedures. First answers to the latter by a few stations/agencies, which had also been posted on the ISC website, revealed however some inconsistencies and ambiguities which required a

revision of the questionnaire which is now annexed to the IASPEI (2013) WG recommendations as well as to IS 3.4. This necessitates a second effort by the ISC to disseminate to all its data contributors and membership agencies these latest versions together with an urgent request for consideration, implementation and feedback in the context of outlining beforehand to them the future ISC policy in implementing these IASPEI recommendations.

Only when pioneered by the ISC as the singular true global seismological parameter data agency with the self-demand to assure - in close collaboration with IASPEI and CoSOI - the best achievable degree of data completeness, quality and unambiguity, also the other main goals e-h) can be achieved in the years ahead in the very best long-term interests of the global seismological community. This, however, also requires a majority support by the ISC Governing Council, the initiator of the IASPEI resolution and thus of the WG establishment. And this position should be independent on the particular national interests, limiting conditions or current priority tasks under which some of the ISC members or cooperating associates have to run at present their daily operations.

## Acknowledgments

The author owes great thanks to his fellow WG members for the many years of fruitful cooperation, and to J. Dewey, D. Storchak and D. Di Giacomo in particular for their critical reading the draft version of this Information Sheet. Their useful comments and editorial recommendations have helped to improve and substantiate the manuscript. J. Saul helped to clarify the representation of magnitude data in the QuakeML format.

## References

- Abe, K. (1981). Magnitudes of large shallow earthquakes from 1904 to 1980. *Phys. Earth Planet. Interiors*, **27**, 72-92.
- Abe, K. (1982). Magnitude, seismic moment and apparent stress for major deep earthquakes. *J. Phys. Earth*, **30**, 321-330.
- Abe, K. (1984). Complements to “Magnitudes of large shallow earthquakes from 1904 to 1980”. *Phys. Earth Planet. Interiors*, **34**, 17-23.
- Abe, K., and Kanamori, H. (1979). Temporal variation of the activity of intermediate and deep focus earthquakes. *J. Geophys. Res.*, **84**, B7, 3589-3595.
- Abe, K. (1989). Quantification of tsunamigenic earthquakes by the  $M_t$  scale. *Tectonophysics*, **166**, 27-34.
- Abe, K., and Kanamori, H. (1980). Magnitudes of great shallow earthquakes from 1953 to 1977. *Tectonophysics*, **62**, 191-203.
- Bonner, J. L., Russel, D. R., Harkrider, D. G., Reiter, D. T., and Herrmann, R. B. (2006). Development of a time-domain, variable-period surface-wave magnitude measurement procedure for application at regional and teleseismic distances, part II: Application and Ms-mb performance. *Bull. Seism. Soc. Am.* **96**, 2, 678-696, doi: 10.1785/0120050056.
- Bormann, P. (2002; 2009). IS 3.2: Proposal for unique magnitude nomenclature. In: Bormann, P. (Ed.) (2002). *New Manual of Seismological Observatory Practice (NMSOP-1)*, IASPEI, GFZ German Research Centre for Geosciences, Potsdam (printed version) or electronic

- version (2009), IS 3.2, 6 pp., DOI: 10.2312/GFZ.NMSOP\_r1\_IS\_3.2., accessible via <http://nmsop.gfz-potsdam.de>.
- Bormann, P., and K. Wylegalla (2005). Quick estimator of the size of great earthquakes. *EOS, Transactions, American Geophysical Union*, **86** (46), 464.
- Bormann, P., and J. Saul (2008). The new IASPEI standard broadband magnitude  $m_B$ , *Seismol. Res. Lett.* **79** (5), 699-706.
- Bormann, P., and Saul, J. (2009). A fast, non-saturating magnitude estimator for great earthquakes, *Seism. Res. Lett.* **80** (5), 808-816; doi: 10.1785/gssrl.80.5.808.
- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bormann, P., Liu, R., Ren, X., Gutdeutsch, R., Kaiser, D., Castellaro, S. (2007). Chinese national network magnitudes, their relation to NEIC magnitudes, and recommendations for new IASPEI magnitude standards, *Bull. Seism. Soc. Am.* **97**, 114-127.
- Bormann, P., Liu, R., Xu, Z., Ren, K., Zhang, L., Wendt S. (2009). First application of the new IASPEI teleseismic magnitude standards to data of the China National Seismographic Network, *Bull. Seism. Soc. Am.*, **99** (3), 1868-1891; doi: 10.1785/0120080010
- Brune, J.N. (1970). Tectonic stress and the spectra of seismic shear waves from earthquakes, *J. Geophys. Res.*, **75**, 4997-5009.
- Brune, J.N. (1971). Correction, *J. Geophys. Res.*, **76**, 5002.
- Choy, G. L., and Boatwright, J. L. (1995). Global patterns of radiated seismic energy and apparent stress. *J. Geophys. Res.*, **100**, B9, 18,205-18,228.
- Choy, G. L., and Kirby, S. H. (2004). Apparent stress, fault maturity and seismic hazard for normal-fault earthquakes at subduction zones, *Geophys. J. Int.*, **159**, 991-1012.
- Di Giacomo, D., Grosser, H., Parolai, S., Bormann P., and Wang, R. (2008). Rapid determination of  $M_e$  for strong to great shallow earthquakes. *Geophys. Res. Lett.*, **35**, L10308; doi:10.1929/2008GL033505.
- Di Giacomo, D., Parolai, S., Bormann P., Grosser, H., Saul, J., Wang, R., and Zschau, J. (2010). Suitability of rapid energy magnitude estimations for emergency response purposes. *Geophys. J. Int.*, **180**, 361-374; doi: 10.1111/j.1365-246X.2009.04416.x.
- Di Giacomo, D., Bondár, I., Storchak, D. A., Engdahl, E. R., Bormann, P., and Harris, J. (2013). ISC-GEM: Global Instrumental Earthquake Catalogue (1900-2009), III. Re-computed  $M_S$  and  $m_b$ , proxy  $M_W$ , final magnitude composition and completeness assessment. *Phys. Earth Planet. Int.* (submitted).
- Dziewonski, A. M., Chou, T. A., and Woodhouse, J.H. (1981). Determination of earthquake source parameters from waveform data for studies of global and regional seismicity. *J. Geophys. Res.*, **86**, 2825-2852.
- Ekström, G., and Dziewonski, A. M. (1988). Evidence of bias in estimations of earthquake size. *Nature*, **332**, 319-323.
- Engdahl, E. R., and Gunst, R. H. (1966). Use of a high speed computer for the preliminary determination of earthquake hypocenters. *Bull. Seism. Soc. Am.*, **56**, 325-336.
- Euchner, F., and Schorlemmer, D. (Eds.) (2013). QuakeML – an XML representation of seismological data. Basic event description. Version 1.2 (14 Febr. 2013).

<https://quake.ethz.ch/quakeml/docs/REC?action=AttachFile&do=view&target=QuakeML-BED-20130214a.pdf>

- Granville, J. P., Kim, W.-Y., and Richards, P. G. (2002). An assessment of seismic body-wave magnitudes published by the Prototype International Data Centre. *Seism. Res. Lett.*, **73**(6), 893-906.
- Granville, J. P., Richards, . G., Kim, W.-Y., and Sykes, L. R. (2005). Understanding the differences between three teleseismic  $m_b$  scales. *Bull. Seism. Soc. Am.*, **95**(5), 1809-1824.
- Gutenberg, B. (1945a). Amplitudes of surface waves and magnitudes of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 3-12.
- Gutenberg, B. (1945b). Amplitudes of P, PP, and S and magnitude of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 57-69.
- Gutenberg, B. (1945c). Magnitude determination of deep-focus earthquakes. *Bull. Seism. Soc. Am.*, **35**, 117-130.
- Gutenberg, B. (1956). The energy of earthquakes. *Q. J. Geol. Soc. London*, **112**, 1-14.
- Gutenberg, B., and Richter, C. F. (1954). Seismicity of the Earth and associated Phenomena. 2<sup>nd</sup> edition, Princeton University Press, 310 pp.
- Gutenberg, B., and Richter, C. F. (1956). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Habermann, R. E. (1995). Opinion. *Seism. Res. Lett.*, **66**, 5, 3.
- Hara, T. (2007). Measurement of the duration of high-frequency energy radiation and its application to determination of the magnitudes of large shallow earthquakes. *Earth Planets Space* **59**, 227-231.
- Hayes, G. P., Rivera, L., and Kanamori, H. (2009). Source inversion of the W-phase: Real-time implementation and extension to low magnitudes. *Seism. Res. Lett.*, **80**, 5, 817-822; doi: 10.1/85/gssrl.80.5.817.
- Herak, M., and Herak, D. (1993). Distance dependence of  $M_S$  and calibrating function for 20 s Rayleigh waves. *Bull. Seism. Soc. Am.*, **83**, 6, 1881-1892.
- Hutton, L. K., and Boore, D. M. (1987). The  $M_L$  scale in southern California. *Bull. Seism. Soc. Am.*, **77**, 2074-2094.
- Hutton, L. K., and Jones, L. M. (1993). Local magnitudes and apparent variations in seismicity rates in Southern California. *Bull. Seism. Soc. Am.*, **83**, 2, 313-329.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf).
- IASPEI (2013) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)  
<http://www.iaspei.org/commissions/CSOI.html#wgmm>
- ISC (2013). Summary of the Bulletin of the International Seismological Centre, 2010, January-June, **47**(1-6), 187 pp.
- Kanamori, H. (1977). The energy release in great earthquakes. *J. Geophys. Res.*, **82**, 2981-2987.
- Kanamori, H., and Anderson, D. L. (1975). Theoretical basis of some empirical relations in seismology. *Bull. Seism. Soc. Am.*, **65**, 1073-1095.



- Kanamori, H., and Rivera, L. (2008). Source inversion of W phase: speeding up seismic tsunami warning. *Geophys. J. Int.*, **175**, 222-238.
- Karnik, V., N. V. Kondorskaya, Ju. Riznitchenko, E. F. Savarensky, S. L. Sobolev, N. V. Shebalin, J. Vanek, and A. Zatopek (1962). Standardization of the earthquake magnitude scale. *Studia geoph. et geod.* **6**, 41-48.
- Lienkaemper, J. J. (1984). Comparison of two surface-wave magnitude scales: M of Gutenberg and Richter (1954) and Ms of "Preliminary Determination of Epicenters". *Bull. Seism. Soc. Am.*, **74**(6), 2357-2378.
- Lomax, A., and A. Michelini (2009).  $M_{wpd}$ : A duration-amplitude procedure for rapid determination of earthquake magnitude and tsunamigenic potential from P waveforms, *Geophys. J. Int.*, **176**, 200-214; doi: 10.1111/j.1365-246X.2008.03974.x.
- Liu, R., Chen, Y., Bormann, P., Ren, X., Hou, J., Zou, L., and Yang, H. (2005). Comparison between earthquake magnitudes determined by China seismograph network and US seismograph network (I). Body wave magnitude. *Acta Seismologica Sinica*, **18**, 6, 627-631.
- Liu, R., Chen, Y., Bormann, P., Ren, X., Hou, J., Zou, L., and Yang, H. (2006). Comparison between earthquake magnitudes determined by China seismograph network and US seismograph network (II). Surface wave magnitude. *Acta Seismologica Sinica*, **19**, 1, 1-7.
- Murphy, J. R., and Barker, B. W. (2003). Revised distance and depth corrections for use in estimation of short-period P-wave magnitudes. *Bull. Seism. Soc. Am.*, **93**, 1746-1764.
- Nuttli, O. W. (1973). Seismic wave attenuation and magnitude relations for eastern North America. *J. Geophys. Res.*, **78**, 876-885.
- Nuttli, O. W. (1985). Average seismic source-parameter relations for plate-margin earthquakes. *Tectonophysics*, **118**, 161-174.
- Nuttli, O. W. (1986). Yield estimates of Nevada Test Site explosions obtained from seismic Lg waves, *J. Geophys. Res.*, **91**, 2137-2151.
- Nuttli, O. W. (1988). Lg magnitudes and yield estimates for underground Novaya Zemlya nuclear explosions, *Bull. Seism. Soc. Am.* **78**, 873-884.
- Okal, E. A., and Talandier, J. (1989).  $M_m$ : A variable-period mantle magnitude. *J. Geophys. Res.*, **94**, 4169-4193.
- Okal, E. A., and Talandier, J. (1990).  $M_m$ : Extension to Love waves of the concept of a variable-period mantle magnitude. *Pure Appl. Geophys.*, **134**, 355-384.
- Rezapour, M., and Pearce, R. G. (1998). Bias in surface-wave magnitude  $M_S$  due to inadequate distance corrections. *Bull. Seism. Soc. Am.*, **88**, 1, 43-61.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bull. Seism. Soc. Am.*, **25**, 1-32.
- Richter, C. F. (1958). Elementary seismology. *W. H. Freeman and Company*, San Francisco and London, viii + 768 pp.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2011). Seismic phase names: IASPEI standards. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, Vol. 2, 1162-1173; doi: 10.1007/978-90-481-8702-7.
- Tsuboi, C. (1954). Determination of the Gutenberg-Richter's magnitude of earthquakes occurring in and near Japan. *Zisin. (J. Seism. Soc. Japan)*, Ser. II, **7**, 185-193.
- Tsuboi, S., Abe, K., Takano, K., and Yamanaka, Y. (1995). Rapid determination of  $M_w$  from broadband P waveforms. *Bull. Seism. Soc. Am.*, **85**, 2, 606-613.

- Uhrhammer, R. A., and Collins, E. R. (1990). Synthesis of Wood-Anderson seismograms from broadband digital records. *Bull. Seism. Soc. Am.*, **80**, 702-716.
- Utsu, T. (2002). Relationships between magnitude scales. In: *International Handbook of Earthquake and Engineering Seismology, Part A*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 733–746.
- Vaněk, J., Zapotek, A., Karnik, V., Kondorskaya, N.V., Riznichenko, Yu.V., Savarensky, E.F., Solov'yov, S.L., and Shebalin, N.V. (1962). Standardization of magnitude scales. *Izvestiya Akad. SSSR., Ser. Geofiz.*, **2**, 153-158 (108-111 in the English translation).
- Veith, K. F., and Clawson, G. E. (1972). Magnitude from short-period P-wave data. *Bull. Seism. Soc. Am.*, **62**, 435-452.
- Willmore, P.L. (1979). Manual of Seismological Observatory Practice. World Data Center A for Solid Earth Geophysics, Report SE-20, September 1979, Boulder, Colorado, 165 pp.; <http://www.seismo.com/msop/msop79/inst/inst1.html>.

| Topic   | The new IASPEI standards for determining magnitudes from digital data and their relation to classical magnitudes                                                                                                                                                                                                                           |
|---------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, Dept. 2, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a><br><b>James W. Dewey</b> , U. S. Geological Survey, MS 966, Denver, CO 80225-0046; Phone: 303 273 8419, E-mail: <a href="mailto:jdewey@usgs.gov">jdewey@usgs.gov</a> |
| Version | March 2012; DOI: 10.2312/GFZ.NMSOP-2_IS_3.3                                                                                                                                                                                                                                                                                                |

**Note:** Figure and Table numbers followed in this text by ??? relate to the NMSOP editions 2002 (printed) and 2009 (website). They will be changed when the revised Chapters 3 and 11 of NMSOP-2 become available.

|                                                                                                                                          | page      |
|------------------------------------------------------------------------------------------------------------------------------------------|-----------|
| <b>1 Why there is a need for measurement standards of magnitudes</b>                                                                     | <b>2</b>  |
| <b>2 Aim of the proposed magnitude standards</b>                                                                                         | <b>3</b>  |
| <b>3 Summary recommendations on determining earthquake magnitudes from digital data</b>                                                  | <b>3</b>  |
| 3.1 Poles and zeroes of Wood-Anderson, WWSSN-SP, and WWSSN-LP seismographs                                                               | 4         |
| 3.2 IASPEI standard procedures for widely used magnitude types                                                                           | 6         |
| 3.3 Magnitude nomenclature                                                                                                               | 11        |
| 3.4 Nomenclature for amplitudes, periods and amplitude-measurement times                                                                 | 13        |
| 3.5 Agency-specific circumstances that may lead to modification of standard procedures                                                   | 13        |
| <b>4 Relation of IASPEI standard procedures to original definitions and to procedures for measuring magnitudes at different agencies</b> | <b>14</b> |
| 4.1 ML                                                                                                                                   | 14        |
| 4.2 mB                                                                                                                                   | 15        |
| 4.3 mb                                                                                                                                   | 16        |
| 4.4 Ms_20                                                                                                                                | 20        |
| 4.5 Ms_BB                                                                                                                                | 22        |
| 4.6 mb_Lg                                                                                                                                | 25        |
| 4.7 Mw                                                                                                                                   | 26        |
| <b>5 Deviations from the recommended measurement standards</b>                                                                           | <b>26</b> |
| 5.1 Variations in filter parameters acceptable for determining standard ML, mb and Ms_20                                                 | 27        |
| 5.1.1 ML                                                                                                                                 | 27        |
| 5.1.2 mb                                                                                                                                 | 28        |
| 5.1.3 Ms_20                                                                                                                              | 29        |
| 5.2 The influence of the measurement time-window on mb and mB estimates                                                                  | 32        |
| 5.3 The influence of the calibration function on the magnitude estimate                                                                  | 34        |
| 5.3.1 ML calibration function                                                                                                            | 34        |
| 5.3.2 mB_BB and mb calibration function                                                                                                  | 35        |
| 5.3.3 Ms_BB and Ms_20 calibration function                                                                                               | 36        |
| 5.3.4 How amplitude and measurement time window should be measured                                                                       | 37        |
| <b>6 Summary</b>                                                                                                                         | <b>39</b> |
| <b>Acknowledgments</b>                                                                                                                   | <b>39</b> |
| <b>References</b>                                                                                                                        | <b>40</b> |

## 1 Why there is a need for measurement standards of magnitudes

In October 2005, the Commission on Seismic Observation and Interpretation of the International Association of Seismology and Physics of the Earth's Interior (IASPEI) adopted the summary recommendations made by the IASPEI Working Group on Magnitudes on new measurement standards for widely used local, regional and teleseismic magnitude scales (IASPEI, 2005). These recommendations have recently been refined and detailed (IASPEI, 2013) and a final scientific report, to be published in a reputable international journal, is currently under preparation.

The Working Group has been established by the IASPEI General Assembly 2001 in Hanoi, following a request of the Governing Council of the International Seismological Center (ISC). The latter was concerned about significant discrepancies between short-period body-wave magnitudes  $m_b$  determined by the US Geological Survey's National Earthquake Information Center (USGS/NEIC) and the Prototype International Data Centre (PIDC), predecessor to the International Data Centre (IDC) of the Comprehensive Test-Ban Treaty Organization (CTBTO) at Vienna. Since then, significant causes of these discrepancies have been identified as coming from differences in the applied short-period filter response, in the length of the measurement time window (Granville et al., 2002 and 2005; Bormann et al., 2007 and 2009) and in the different calibration functions used (Gutenberg and Richter, 1956a, and Veith and Clawson, 1972, respectively; in the following abbreviated by G-R and V-C) for correcting the effects of epicentral distance and source depth on the measured amplitudes (Murphy and Barker, 2003; Granville et al., 2005). The average discrepancy between  $m_b$ (NEIC) and  $m_b$ (IDC) was found to be approximately +0.4 magnitude units (m.u.) (Granville et al., 2002). This difference, however, strongly increases with magnitude and is larger than 1 m.u. for really great earthquakes (e.g., 1.3 m.u. for the 2004 Mw9.3 Sumatra-Andaman earthquake; see Bormann et al., 2007). Further, the differences are smaller for shallow earthquakes than for deep earthquakes, because for source depths  $h > 300$  km the V-C corrections are on average 0.4 m.u. (up to 0.6 m.u. for the deepest earthquakes) smaller than the G-R corrections. Despite these systematic discrepancies, the USGS/NEIC and the IDC publish their body-wave magnitude with the same symbol  $m_b$ .

Similarly, the symbol  $M_s$  is commonly used both for surface-wave magnitudes based on amplitude measurements around 20 sec period and those measured in a much wider period range. For both  $M_s$  measured near 20s and  $M_s$  measured over broader period ranges, multiple attenuation functions have been proposed. These may produce magnitudes that differ systematically (sometimes up to about 0.6 m.u.), especially for smaller earthquakes recorded only at distances  $< 30^\circ$  (Bormann et al., 2009).

Publishing magnitudes derived by different measurement procedures with identical magnitude symbols may not only confuse and mislead users of such data, it also jeopardizes the authenticity, compatibility and long-term continuity of magnitude data and thus their use in earthquake statistics and related applications. For example, when merging  $m_b$ (NEIC) and  $m_b$ (IDC) data at the ISC for calculating event averages and standard deviations, systematic differences and increased data scatter may significantly bias such estimates for magnitudes larger than 5. Therefore, IASPEI realized a general need to identify such discrepancies in measurement practices of widely determined different types of magnitudes, to analyze, quantify and understand the reasons for such discrepancies and to propose international standards for making such measurements.

## 2 Aim of the proposed magnitude standards

IASPEI recommended that measurement standards for magnitudes should, as far as possible:

- agree with magnitudes of the same type that have been measured for decades from analog seismograms according to original definitions;
- promote the best possible use of the advantages of digital data and processing;
- minimize bulletin magnitude biases that result from procedure-dependent single-station or network magnitude biases;
- increase the number of seismological stations and networks with well-defined procedures;
- increase essentially the accuracy, representativeness, homogeneity and long-term global compatibility of magnitude data and their usefulness for seismic hazard assessment and research.

## 3 Summary recommendations on determining earthquake magnitudes from digital data

The Working Group on Magnitudes (for short, *Magnitude WG*) of the IASPEI Commission on Seismological Observation and Interpretation was established to recommend standard procedures for making measurements from digital data to be used in calculating several widely used types of earthquake magnitude. The recommended procedures from the *Magnitude WG* have been approved by the IASPEI Commission on Seismological Observations and Interpretations. We henceforth refer to the proposed procedures as the IASPEI standard procedures for magnitude determination. The *Magnitude WG* is planning to publish an article, in the open literature, that explains the rationale for the IASPEI standard procedures. Current members of the *Magnitude WG* are, in alphabetical order: P. Bormann (co-chair), J. M. Dewey (co-chair), I. Gabsatarova, S. Gregersen, A. Gusev, W. Kim, R. Liu, H. Patton, B. Presgrave, J. Saul, R. A. Uhrhammer, S. Wendt.

The IASPEI standard procedures address the measurement of amplitudes and periods from digital data for use in calculating the generic magnitude types  $ML$ ,  $Ms$ ,  $mb$ ,  $mB$ , and  $mb\_Lg$ . For  $Ms$ , standard procedures are proposed for two different traditions --  $Ms$  measured from waves with periods near 20s [here denoted  $Ms_{20}$ ] and  $Ms$  measured from waves in a much broader period-range [here denoted  $Ms_{BB}$ ]. For the generic intermediate-period/broadband body-wave magnitude  $mB$ , we propose a procedure based on the maximum amplitude of the  $P$ -wave measured on a velocity-proportional trace: we denote the resulting magnitude  $mB_{BB}$ . The IASPEI standard procedures also specify a standard equation for  $Mw$  from among several slightly different equations. Abbreviated descriptions of the procedures are described below. More detailed descriptions, and discussion of acceptable alternatives to specific steps in the application of individual procedures, will be discussed in the planned journal article and in the last section of this Information Sheet.

Some of the IASPEI standard procedures require the use of broadband (BB) records that are proportional to ground motion velocity at least within the period range that has been recommended for the measurement of the respective magnitudes. Other procedures require

filtering BB records so that they replicate the response of classical standard seismographs, such as the Wood-Anderson (WA) seismograph or the short-period (SP) and long-period (LP) seismographs that were used in the World-Wide Standardized Seismograph Network (WWSSN) (next section).

### 3.1 Poles and zeroes for Wood-Anderson, WWSSN-SP, and WWSSN-LP seismographs

Poles and zeroes corresponding to the displacement transfer functions for representative “average” classical analog standard seismographs WWSSN-SP, WWSSN-LP and Wood Anderson with their equivalent eigenperiod and damping of the seismometer and galvanometer and coupling factor  $\sigma^2$  (where applicable) are given in Table 1 (courtesy of Charles R. Hutt, U. S. Geological Survey, who also provided much of the immediately following documentation). Note that the number of zeroes determines the slope of the response at the low-frequency end and that this slope changes towards higher frequencies at the frequencies determined by the poles. At the frequency of a conjugate complex pole the slope is reduced by two orders, while at a single pole with real part only, the slope is reduced by one order only. The left number in brackets is the value of the real part; the right number that of the imaginary part. The normalized amplitude responses corresponding to the poles and zeroes of Table 1 are shown in Figure 1.

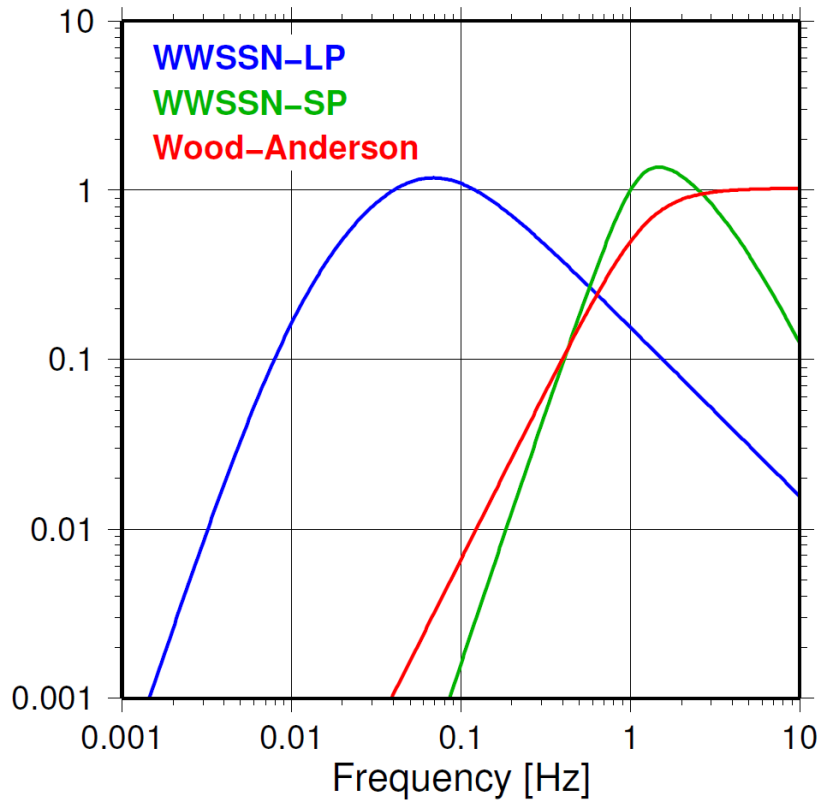
At magnifications of 50,000 and higher, transfer functions for the WWSSN short-period seismograph differed somewhat for different magnification settings, due to galvanometer-seismometer reaction. The response shown in Table 1 and recommended for use with *mb* is appropriate to a “100,000 magnification” short-period WWSSN seismograph.

Transfer functions for the WWSSN long-period seismograph also differed somewhat for different magnification settings, due to galvanometer-seismometer reaction. The response shown in Table 1 is appropriate to a “1500 magnification” long-period vertical WWSSN seismograph. The transfer functions of the WWSSN-LP horizontal seismograph also differed from that of the WWSSN-LP vertical seismograph. One should note that the coupling which exists in electromagnetic seismographs with galvanometric recording between the seismometer and galvanometer has been taken into account in the listed poles in such a way as to make the coupling factor effectively zero. Accordingly, the resulting free periods and damping values given for the equivalent seismometer ( $T_s$  and  $h_s$ ) and galvanometer ( $T_g$  and  $h_g$ ) in Table 1 are somewhat different from the nominal values of the uncoupled system components of classical seismographs and the respective coupling factors are reported as zero. The nominal free periods of the uncoupled components are  $T_s = 1.00$  s and  $T_g = 0.75$  s for WWSSN-SP and  $T_s = 15.0$  s and  $T_g = 100.0$  s for WWSSN-LP.

The frequency response of the Wood-Anderson seismograph (Table 1) is based on the paper of Uhrhammer and Collins (1990).

**Table 1** Zeroes and poles corresponding to the displacement transfer function of the WWSSN-SP seismograph, WWSSN-LP seismograph, and the Wood-Anderson (WA) seismograph. Zeroes and poles are represented in angular frequency (radians per second).  $T_s$  = seismometer free period;  $h_s$  = seismometer damping constant;  $T_g$  = galvanometer free period;  $h_g$  = galvanometer damping constant;  $\sigma^2$  = coupling factor. In this representation, free periods and damping constants have been adjusted slightly so that response can be correctly modeled with  $\sigma^2 = 0$ . The displacement-response functions implied by the poles and zeroes are commonly normalized with respect to the following frequencies ( $f_n$ ) using the associated normalization factors ( $A_0$ ) as follows: WWSSN-SP,  $f_n = 1$  Hz,  $A_0 = 532.14$ ; WWSSN-LP,  $f_n = 0.04$  Hz,  $A_0 = 0.97866$ ; WA,  $f_n = 4$  Hz,  $A_0 = 1.0028$ .

| Seismograph | Zeros                                  | Poles                                                                                                                                                                                                           | $T_s/s$ | $h_s$  | $T_g/s$ | $h_g$  |
|-------------|----------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|---------|--------|---------|--------|
| WWSSN-SP    | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0) | (-3.72500, -6.22000) (=p <sub>1</sub> )<br>(-3.72500, 6.22000) (=p <sub>2</sub> )<br>(-5.61200, 0.00000) (=p <sub>3</sub> )<br>(-13.2400, 0.00000) (=p <sub>4</sub> )<br>(-21.0800, 0.00000) (=p <sub>5</sub> ) | 0.867   | 0.5138 | 0.729   | 1.0935 |
| WWSSN-LP    | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0) | (-0.40180, 0.08559)<br>(-0.40180, 0.08559)<br>(-0.04841, 0.00000)<br>(-0.08816, 0.00000)                                                                                                                        | 15.29   | 0.978  | 96.18   | 1.045  |
| WA          | (0.0, 0.0)<br>(0.0, 0.0)               | (-5.49779, -5.60886)<br>(-5.49779, 5.60886)                                                                                                                                                                     | 0.8     | 0.7    |         |        |



**Figure 1** Displacement amplitude-frequency responses of the classical seismographs: WWSSN-SP, WWSSN-LP and Wood-Anderson (WA), normalized to the frequencies ( $f_n$ ) in the caption of Table 1.

### 3.2 IASPEI standard procedures for widely used magnitude types

The amplitudes used in the magnitude formulas below are in most circumstances to be measured as one-half the maximum deflection of the seismogram trace, peak-to-adjacent-trough or trough-to-adjacent-peak, where peak and trough are separated by one crossing of the zero-line: the measurement is sometimes described as “one-half peak-to-peak amplitude.” None of the magnitude formulas presented in this article are intended to be used with the full peak-to-trough deflection as the amplitude. The periods are to be measured as twice the time-intervals separating the peak and adjacent-trough from which the amplitudes are measured. The amplitude-phase arrival-times are to be measured as the time of the zero-crossing between the peak and adjacent-trough from which the amplitudes are measured. The issue of amplitude and period measuring procedures, and circumstances under which alternative procedures are acceptable or preferable, is discussed further in Section 5.

Modern digital seismogram analysis programs commonly measure amplitudes in units of nm (for displacement) or nm/s (for velocity), respectively, and not, as assumed by the classical magnitude formulas, in  $\mu\text{m}$ . The commonly known classical calibration relationships have been modified to be consistent with displacements measured in nm.

***ML*** – local magnitude consistent with the magnitude of Richter (1935)

For crustal earthquakes in regions with attenuative properties similar to those of Southern California, the proposed standard equation is

$$ML = \log_{10}(A) + 1.11 \log_{10}R + 0.00189R - 2.09, \quad (1)$$

where:

$A$  = maximum **trace** amplitude in **nm** that is measured on output from a **horizontal-component** instrument that is filtered so that the response of the seismograph/filter system replicates that of a **Wood-Anderson standard seismograph** but with a static magnification of 1 (see Table 1 and Figure 1);

$R$  = **hypocentral distance in km**, typically less than 1000 km.

Equation (1) is an expansion of that of Hutton and Boore (1987). The constant term in equation (1), -2.09, is based on an experimentally determined static magnification of the Wood-Anderson of 2080, rather than the theoretical magnification of 2800 that was specified by the seismograph's manufacturer. The formulation of equation (1) reflects the intent of the *Magnitude WG* that reported *ML* amplitude data not be affected by uncertainty in the static magnification of the Wood-Anderson seismograph.

For seismographic stations containing two horizontal components, amplitudes are measured independently from each horizontal component, and each amplitude is treated as a single datum. There is no effort to measure the two observations at the same time, and there is no attempt to compute a vector average.



For crustal earthquakes in regions with attenuative properties that are **different** than those of coastal California, and for measuring magnitudes with vertical-component seismographs, the standard equation is of the form:

$$ML = \log_{10}(A) + C(R) + D, \quad (2)$$

where  $A$  and  $R$  are as defined in equation (1), except that  $A$  may be measured from a **vertical-component** instrument, and where  $C(R)$  and  $D$  have been **calibrated** to adjust for the different regional attenuation and to adjust for any systematic differences between amplitudes measured on horizontal seismographs and those measured on vertical seismographs.

***Ms\_20 – teleseismic surface-wave magnitudes at period of ~20 s***

$$Ms_{20} = \log_{10}(A/T) + 1.66\log_{10}\Delta + 0.3, \quad (3)$$

where:

$A$  = **vertical-component** ground displacement in **nm** measured from the maximum trace-amplitude of a surface-wave phase having a period between **18 s and 22 s** on a waveform that has been filtered so that the frequency response of the seismograph/filter system replicates that of a World-Wide Standardized Seismograph Network (WWSSN) **long-period seismograph** (see Table 1), with  $A$  being determined by dividing the maximum trace amplitude by the magnification of the simulated WWSSN-LP response at period  $T$ ;

$T$  = period in seconds ( $18 \text{ s} \leq T \leq 22 \text{ s}$ );

$\Delta$  = **epicentral** distance in degrees,  $20^\circ \leq \Delta \leq 160^\circ$ ;

Equation (3) is formally equivalent to the  $Ms$  equation proposed by Vaněk et al. (1962) but is here applied to vertical motion measurements in a narrow range of periods.

Some agencies compute  $Ms_{20}$  only for shallow-focus earthquakes (typically those whose confidence-intervals on focal-depth would allow them to be shallower than 50 or 60 km).  $Ms_{20}$  would be expected to significantly under-represent the energy of intermediate- and deep-focus earthquakes, due to their less effective generation of surface waves, unless an adjustment is made to account for their large focal-depths, e.g., according to Herak et al. (2001).

***Ms\_BB – surface-wave magnitudes from broad-band instruments***

$$Ms_{BB} = \log_{10}(V_{max}/2\pi) + 1.66 \log_{10}\Delta + 0.3, \quad (4)$$

where:

$V_{max}$  = ground **velocity in nm/s** associated with the maximum trace-amplitude in the surface-wave train, as recorded on a **vertical-component** seismogram that is **proportional to velocity**, where the period of the surface-wave,  $T$ , should satisfy the condition  $3 \text{ s} < T < 60 \text{ s}$ , and where  $T$  should be preserved together with  $V_{max}$  in bulletin data-bases;

$\Delta$  = **epicentral** distance in degrees,  $2^\circ \leq \Delta \leq 160^\circ$

As with  $M_s_{20}$ , some agencies compute  $M_s_{BB}$  only for shallow-focus earthquakes.

Equation (4) is based on the  $M_s$  equation proposed by Vaněk et al. (1962), but is here applied to vertical motion measurements and is used with the  $\log_{10}(V_{max}/2\pi)$  term replacing the  $\log_{10}(A/T)_{max}$  term of the original.

***mb*** – short-period body-wave magnitude

$$mb = \log_{10}(A/T) + Q(\Delta, h) - 3.0, \quad (5)$$

where:

$A$  = P-wave ground amplitude in **nm** calculated from the maximum trace-amplitude in the **entire P-phase train** (time spanned by P, pP, sP, and possibly PcP and their codas, and ending preferably before PP);

$T$  = period in seconds,  $T < 3 \text{ s}$ ; of the maximum P-wave trace amplitude.

$Q(\Delta, h)$  = attenuation function for **PZ** (P-waves recorded on vertical component seismographs) established by **Gutenberg and Richter (1956a)** in the tabulated or algorithmic form as used by the U.S. Geological Survey/National Earthquake Information Center (USGS/NEIC) (**Table 2**);

$\Delta$  = **epicentral** distance in degrees,  $20^\circ \leq \Delta \leq 100^\circ$ ;

$h$  = focal depth in **km**;

and where both  $T$  and the maximum trace amplitude are measured on output from a **vertical-component** instrument that is filtered so that the frequency response of the seismograph/filter system replicates that of a WWSSN **short-period** seismograph (**Table 1**), with  $A$  being determined by dividing the maximum trace amplitude by the magnification of the simulated WWSSN-SP response at period  $T$ .

***mB\_BB*** – broadband body-wave magnitude

$$mB_{BB} = \log_{10}(V_{max}/2\pi) + Q(\Delta, h) - 3.0, \quad (6)$$

where:

$V_{max}$  = ground **velocity in nm/s** associated with the maximum trace-amplitude in **the entire P-phase train** (time spanned by P, pP, sP, and possibly PcP and their codas, but ending *preferably* before PP (*Note: see discussion in the final section*), as recorded on a vertical-component seismogram that is **proportional to velocity**, where the period of the measured phase,  $T$ , should satisfy the condition  $0.2 \text{ s} < T < 30 \text{ s}$ , and where  $T$  should be preserved together with  $V_{max}$  in bulletin data-bases;

$Q(\Delta, h)$  = attenuation function for PZ established by Gutenberg and Richter (1956a), (Table 2);

$\Delta$  = epicentral distance in degrees,  $20^\circ \leq \Delta \leq 100^\circ$ ;

$h$  = focal depth in **km**.

Equation (6) differs from the equation for  $m_B$  of Gutenberg and Richter (1956a) by virtue of the  $\log_{10}(V_{max}/2\pi)$  term, which replaces the classical  $\log_{10}(A/T)_{max}$  term.

**Table 2** Attenuation (Q) function to be used with  $mb$  (equation 5) and  $mB_{BB}$  (equation 6). This version of  $Q(\Delta, \text{focal-depth})$  was digitized in the 1960's from Figure 5 of Gutenberg and Richter (1956a) and is used at the USGS/NEIC. This is a comma-delimited file, intended to be copied into applications, with the first two rows being header rows. The first column is epicentral distance  $D (= \Delta)$  in degrees, and the following columns give Q for distance  $D$  and depths 0, 25, ...700 km. For  $D$  and focal-depth lying between tabulated values, Q is obtained by linear interpolation from the four neighboring tabulated values.

```

D, Focal Depth (km),,,,,,,,,,,,,,
,0.0, 25, 50, 75,100,150,200,250,300,350,400,450,500,550,600,650,700
20,6.1,6.1,6.1,6.1,6.1,6.2,6.3,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.2,6.0
21,6.1,6.2,6.1,6.1,6.1,6.2,6.3,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.2,6.0
22,6.2,6.2,6.2,6.2,6.1,6.2,6.3,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.3,6.1
23,6.3,6.3,6.2,6.2,6.1,6.2,6.4,6.3,6.2,6.1,6.2,6.3,6.4,6.4,6.4,6.3,6.1
24,6.4,6.3,6.3,6.2,6.2,6.3,6.4,6.3,6.2,6.1,6.2,6.3,6.3,6.4,6.4,6.4,6.1
25,6.5,6.4,6.3,6.2,6.2,6.3,6.4,6.3,6.2,6.1,6.2,6.3,6.3,6.4,6.4,6.4,6.2
26,6.5,6.4,6.3,6.3,6.3,6.4,6.5,6.4,6.2,6.1,6.2,6.2,6.3,6.4,6.4,6.4,6.2
27,6.5,6.4,6.4,6.3,6.3,6.4,6.5,6.4,6.2,6.1,6.2,6.2,6.3,6.4,6.4,6.4,6.3
28,6.6,6.5,6.4,6.4,6.4,6.5,6.5,6.4,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.3
29,6.6,6.5,6.4,6.4,6.4,6.5,6.5,6.4,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.3
30,6.6,6.6,6.5,6.5,6.5,6.5,6.5,6.4,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.3
31,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.3
32,6.7,6.7,6.6,6.6,6.5,6.6,6.4,6.4,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.4
33,6.7,6.7,6.6,6.6,6.6,6.5,6.4,6.4,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.4
34,6.7,6.7,6.7,6.7,6.6,6.5,6.4,6.4,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.4,6.3
35,6.6,6.7,6.7,6.7,6.7,6.5,6.4,6.3,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.3,6.3
36,6.6,6.7,6.7,6.7,6.7,6.5,6.4,6.3,6.3,6.1,6.1,6.2,6.3,6.4,6.4,6.3,6.3
37,6.5,6.6,6.7,6.7,6.7,6.5,6.4,6.3,6.2,6.1,6.1,6.2,6.3,6.4,6.4,6.3,6.3
38,6.5,6.6,6.7,6.7,6.7,6.5,6.4,6.3,6.2,6.1,6.1,6.2,6.3,6.4,6.3,6.3,6.3
39,6.4,6.5,6.6,6.7,6.6,6.5,6.4,6.3,6.1,6.0,6.1,6.2,6.3,6.4,6.3,6.3,6.3
40,6.4,6.5,6.6,6.7,6.6,6.5,6.3,6.2,6.1,6.0,6.1,6.2,6.3,6.4,6.3,6.2,6.3
41,6.5,6.5,6.5,6.6,6.6,6.4,6.3,6.2,6.0,6.0,6.1,6.2,6.3,6.3,6.3,6.2,6.3
42,6.5,6.5,6.5,6.6,6.6,6.4,6.3,6.2,6.0,6.0,6.1,6.2,6.3,6.3,6.3,6.2,6.3
43,6.5,6.5,6.5,6.6,6.6,6.4,6.3,6.1,6.0,6.0,6.1,6.2,6.3,6.3,6.3,6.2,6.3
44,6.6,6.6,6.5,6.6,6.6,6.4,6.3,6.1,6.1,6.0,6.1,6.2,6.3,6.3,6.3,6.2,6.2
45,6.7,6.7,6.6,6.6,6.6,6.4,6.2,6.1,6.1,6.0,6.1,6.2,6.3,6.3,6.3,6.2,6.2

```

**D, Focal Depth (km)(continuation),,,,,,,,,,,,,,**  
**,0.0, 25, 50, 75,100,150,200,250,300,350,400,450,500,550,600,650,700**  
46,6.8,6.7,6.7,6.7,6.6,6.4,6.2,6.1,6.1,6.0,6.1,6.2,6.3,6.3,6.3,6.2,6.2  
47,6.9,6.8,6.7,6.7,6.6,6.4,6.2,6.1,6.1,6.0,6.1,6.2,6.3,6.3,6.3,6.2,6.2  
48,6.9,6.8,6.8,6.7,6.6,6.5,6.2,6.1,6.1,6.0,6.1,6.2,6.2,6.3,6.3,6.2,6.2  
49,6.8,6.8,6.8,6.8,6.7,6.5,6.2,6.2,6.1,6.1,6.1,6.2,6.2,6.3,6.3,6.2,6.2  
50,6.7,6.8,6.8,6.8,6.8,6.5,6.3,6.2,6.1,6.1,6.1,6.1,6.2,6.3,6.3,6.1,6.1  
51,6.7,6.7,6.8,6.8,6.8,6.5,6.3,6.2,6.2,6.1,6.1,6.1,6.2,6.2,6.2,6.1,6.1  
52,6.7,6.7,6.8,6.8,6.8,6.5,6.4,6.2,6.2,6.1,6.1,6.1,6.1,6.2,6.2,6.1,6.1  
53,6.7,6.7,6.8,6.8,6.8,6.6,6.4,6.2,6.2,6.1,6.1,6.1,6.1,6.1,6.2,6.1,6.1  
54,6.8,6.8,6.8,6.8,6.8,6.6,6.4,6.3,6.2,6.1,6.1,6.1,6.1,6.1,6.1,6.1,6.0  
55,6.8,6.8,6.8,6.8,6.8,6.6,6.5,6.3,6.2,6.2,6.1,6.1,6.1,6.1,6.1,6.0,6.0  
56,6.8,6.8,6.8,6.8,6.8,6.7,6.5,6.3,6.2,6.2,6.1,6.1,6.1,6.1,6.1,6.0,6.0  
57,6.8,6.8,6.8,6.9,6.8,6.7,6.5,6.4,6.2,6.2,6.2,6.2,6.1,6.1,6.0,6.0,6.0  
58,6.8,6.8,6.9,6.9,6.8,6.7,6.5,6.4,6.3,6.2,6.2,6.2,6.1,6.1,6.0,6.0,6.0  
59,6.9,6.9,6.9,6.9,6.9,6.7,6.5,6.4,6.3,6.2,6.2,6.2,6.2,6.1,6.0,6.0,6.0  
60,6.9,6.9,6.9,6.9,6.9,6.7,6.5,6.4,6.3,6.3,6.2,6.2,6.2,6.1,6.0,6.0,6.0  
61,6.9,6.9,6.9,6.9,6.8,6.7,6.5,6.4,6.3,6.3,6.3,6.3,6.2,6.2,6.1,6.0,6.0  
62,7.0,6.9,6.9,6.9,6.8,6.7,6.6,6.4,6.4,6.3,6.3,6.3,6.3,6.2,6.1,6.1,6.0  
63,7.0,6.9,6.9,6.8,6.7,6.7,6.6,6.5,6.4,6.4,6.4,6.3,6.3,6.2,6.2,6.1,6.0  
64,7.0,6.9,6.8,6.7,6.7,6.7,6.6,6.5,6.5,6.4,6.4,6.4,6.4,6.3,6.2,6.1,6.1  
65,7.0,6.9,6.8,6.7,6.7,6.7,6.6,6.5,6.5,6.5,6.4,6.4,6.4,6.3,6.2,6.1,6.1  
66,7.0,6.9,6.8,6.7,6.7,6.7,6.5,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.2,6.2,6.1  
67,7.0,6.9,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.3,6.2,6.1  
68,7.0,6.9,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.3,6.2,6.2  
69,7.0,6.9,6.7,6.7,6.6,6.6,6.5,6.5,6.5,6.5,6.4,6.4,6.4,6.3,6.3,6.2,6.2  
70,6.9,6.9,6.7,6.7,6.6,6.6,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.3,6.3,6.2,6.2  
71,6.9,6.9,6.7,6.7,6.6,6.6,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.3,6.3,6.3,6.2  
72,6.9,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.3,6.3,6.3,6.2  
73,6.9,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.3,6.3,6.3,6.3  
74,6.8,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4,6.4,6.3,6.3,6.3,6.3,6.3  
75,6.8,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.5,6.4,6.3,6.2,6.3,6.3,6.3  
76,6.9,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.5,6.4,6.3,6.2,6.3,6.3,6.3  
77,6.9,6.8,6.8,6.7,6.6,6.5,6.5,6.5,6.5,6.6,6.5,6.4,6.2,6.2,6.2,6.3,6.3  
78,6.9,6.8,6.8,6.7,6.6,6.5,6.5,6.5,6.5,6.6,6.5,6.4,6.2,6.2,6.2,6.3,6.3  
79,6.8,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.6,6.6,6.5,6.4,6.2,6.2,6.2,6.3,6.3  
80,6.7,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.6,6.6,6.5,6.4,6.2,6.2,6.2,6.3,6.3  
81,6.8,6.8,6.7,6.7,6.6,6.5,6.5,6.5,6.6,6.6,6.5,6.4,6.3,6.3,6.3,6.3,6.3  
82,6.9,6.8,6.8,6.7,6.6,6.5,6.5,6.5,6.6,6.6,6.5,6.4,6.3,6.3,6.3,6.3,6.3  
83,7.0,6.9,6.8,6.7,6.7,6.6,6.5,6.5,6.6,6.6,6.5,6.5,6.3,6.3,6.3,6.4,6.3  
84,7.0,7.0,6.8,6.8,6.7,6.6,6.5,6.6,6.6,6.6,6.5,6.5,6.4,6.4,6.4,6.4,6.3  
85,7.0,7.0,6.9,6.8,6.7,6.6,6.5,6.6,6.6,6.6,6.6,6.5,6.4,6.4,6.4,6.4,6.4  
86,6.9,7.0,7.0,6.8,6.8,6.6,6.6,6.6,6.6,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4  
87,7.0,7.0,7.0,6.9,6.8,6.7,6.6,6.6,6.7,6.7,6.6,6.5,6.5,6.5,6.5,6.5,6.4  
88,7.1,7.1,7.0,6.9,6.8,6.8,6.6,6.6,6.7,6.7,6.6,6.6,6.6,6.6,6.6,6.5,6.4  
89,7.0,7.1,7.1,7.0,6.9,6.8,6.7,6.7,6.7,6.7,6.6,6.6,6.6,6.6,6.7,6.7,6.6,6.5  
90,7.0,7.0,7.1,7.0,6.9,6.8,6.7,6.7,6.7,6.7,6.6,6.7,6.7,6.7,6.7,6.7,6.7,6.5  
91,7.1,7.1,7.2,7.1,7.0,6.9,6.8,6.7,6.7,6.7,6.7,6.7,6.7,6.7,6.8,6.8,6.7,6.6  
92,7.1,7.2,7.2,7.2,7.1,6.9,6.8,6.8,6.7,6.8,6.7,6.8,6.8,6.8,6.8,6.8,6.8,6.7  
93,7.2,7.2,7.2,7.2,7.1,7.0,6.9,6.8,6.8,6.8,6.8,6.8,6.8,6.9,6.8,6.9,6.7  
94,7.1,7.2,7.2,7.2,7.2,7.0,6.9,6.9,6.9,6.9,6.9,6.9,6.9,6.9,7.0,6.9,6.8  
95,7.2,7.2,7.2,7.2,7.2,7.1,7.0,7.0,6.9,6.9,6.9,6.9,6.9,7.0,7.0,7.0,6.9  
96,7.3,7.2,7.3,7.3,7.3,7.2,7.1,7.0,7.0,7.0,7.0,7.0,7.0,7.0,7.0,7.0,7.0,6.9  
97,7.4,7.3,7.3,7.3,7.3,7.2,7.1,7.1,7.0,7.0,7.0,7.0,7.1,7.1,7.1,7.0,7.0  
98,7.5,7.3,7.3,7.3,7.3,7.3,7.2,7.1,7.1,7.1,7.1,7.1,7.1,7.1,7.1,7.1,7.0  
99,7.5,7.3,7.3,7.3,7.4,7.3,7.2,7.2,7.2,7.1,7.1,7.2,7.2,7.2,7.2,7.1,7.0  
100,7.3,7.3,7.3,7.4,7.4,7.3,7.2,7.2,7.2,7.2,7.2,7.2,7.2,7.2,7.2,7.2,7.1

-----

***mb\_Lg*** – regional magnitude based on the amplitude of *Lg* measured in a narrow period range around 1 s

$$mb\_Lg = \log_{10}(A) + 0.833\log_{10}[r] + 0.4343\gamma(r - 10) - 0.87 \quad (7)$$

where:

*A* = “sustained ground-motion amplitude” in **nm**, defined as the third largest amplitude in the time window corresponding to group velocities of 3.6 to 3.2 km/s, in the period (*T*) range 0.7 s to 1.3 s;

*r* = **epicentral distance** in **km**

$\gamma$  = coefficient of attenuation in  $\text{km}^{-1}$ .  $\gamma$  is related to the quality factor *Q* through the equation  $\gamma = \pi/(Q \cdot U \cdot T)$ , where *U* is group velocity and *T* is the wave period of the *Lg* wave.  $\gamma$  is a strong function of crustal structure and should be determined specifically for the region in which the *mb\_Lg* is to be used.

*A* and *T* are measured on output from a **vertical-component** instrument that is filtered so that the frequency response of the seismograph/filter system replicates that of a WWSSN **short-period** seismograph. Arrival times with respect to the origin of the seismic disturbance are used, along with epicentral distance, to compute group velocity *U*.

***Mw*** – moment magnitude

$$Mw = (\log_{10}M_0 - 9.1)/1.5,$$

where *M<sub>0</sub>* = scalar moment in **N·m**, determined from waveform modeling or from the long-period asymptote of spectra.

or its CGS equivalent (*M<sub>0</sub>* in **dyne·cm**),

$$Mw = (\log_{10}M_0 - 16.1)/1.5.$$

The order of operations in the right-hand sides of equations for *Mw*, subtraction prior to multiplication by 2/3, avoids an ambiguity that arises if multiplication is performed prior to subtraction, which in a certain percentage of cases leads to *Mw* being different according to whether the moment is expressed in CGS or SI units (Utsu, 2002). In this context it may be worth mentioning, that P. Bormann and G. Choy have agreed on the same way of standardized writing of the energy magnitude relationship  $Me = (\log E_S - 4.4)/1.5$  instead of the expanded and rounded-off version  $Me = (2/3)\log E_S - 2.9$ .

### 3.3 Magnitude nomenclature

The “*ML*, *Ms\_20*, *Ms\_BB*, *mb*, *mB\_BB*, *Mw*, *mb\_Lg*” nomenclature used in this section of IS\_3.3 is defined to facilitate data transmission in the IASPEI Seismic Format (ISF) (International Seismological Centre, <http://www.isc.ac.uk/standards/isf>, last accessed March

2012). In the ISF format, magnitude nomenclature is restricted to five characters. The recommended magnitude nomenclature is intended to be consistent with nomenclature that has been used by editorial boards of the Bulletin of the Seismological Society of America, except that the recommended nomenclature does not subscript characters and uses a hyphen in place of parentheses in order to reduce character count. The representation of the magnitude names in italics is for the purpose of distinguishing nomenclature from regular text in the present section of this Information Sheet; we do not consider italics an element of nomenclature. The recommended nomenclature differs from other widely used current nomenclatures. Relationships between nomenclatures are illustrated in Table 3.

In Table 3 “**BSSA editorial style**” is the nomenclature that seems to us most representative of style recommended in recent decades by editors of the Bulletin of the Seismological Society of America. “**NMSOP generic**” and “**NMSOP specific**” are respectively the “generic” and “specific” nomenclature used in the first edition of the IASPEI New Manual of Seismological Observatory Practice (Bormann, 2002a, 2002b). In the specific nomenclature the general magnitude symbol *M* is followed by the symbol of the seismic phase and then by the symbol of the component (V – vertical; H – horizontal) on which amplitude is measured. Then the first letter(s) in brackets stand(s) for the instrument type on which magnitude is measured: e.g., A = WWSSN-SP, B = WWSSN-LP, D = velocity broadband, WA = Wood Anderson. Further symbols/names stand for specific calibration functions used, e.g., “CF” for California and “Author” for author- or agency specific calibration function. “USGS Search” is the nomenclature used by the USGS/NEIC in its Earthquake Catalog Search (<http://neic.usgs.gov/neis/epic/>, last accessed March 2012). “USGS EDR” is the nomenclature used by the USGS/NEIC in its on-line machine-readable Earthquake Data Reports (<ftp://hazards.cr.usgs.gov/edr/>, last accessed March 2012). “ISC” is the nomenclature historically used by the International Seismological Centre. USGS/NEIC and ISC nomenclature has changed as advances in information technology have permitted greater flexibility in nomenclature.

**Table 3** A sample of alternative nomenclatures for the magnitude types that are considered in this report. For detailed explanation see text.

| This report | BSSA editorial style                   | NMSOP generic | NMSOP specific          | USGS Search | USGS EDR | ISC                  |
|-------------|----------------------------------------|---------------|-------------------------|-------------|----------|----------------------|
| ML          | <i>M<sub>L</sub></i>                   | Ml            | MH(WA;CF),<br>MV(WA;CF) | ML          | ML       | <i>M<sub>L</sub></i> |
| Ms_20       | <i>M<sub>S</sub></i> (20)              | Ms            | MLV(B)                  | Ms          | MSZ      | Ms                   |
| Ms_BB       | <i>M<sub>S</sub></i> (BB)              | Ms            | MLV(D)                  | --          | --       | Ms                   |
| mb          | <i>m<sub>b</sub></i>                   | mb            | MPV(A)                  | mb          | MB       | Mb                   |
| mB_BB       | <i>m<sub>B</sub></i> (BB)              | mB            | MPV(D)                  | --          | --       | mB                   |
| mb_Lg       | <i>m<sub>b</sub></i> (L <sub>g</sub> ) | mbLg          | MLgV(A;Author)          | Lg          | LG       | MN                   |

In conjunction with the full-fledged implementation of the new IASPEI standards, the WG encourages seismological agencies to work toward harmonizing their nomenclatures in order

to minimize misunderstanding on the side of data users and to better guide users' search for standard magnitude data.

### 3.4 Nomenclature for amplitudes, periods, and amplitude-measurement times

The ISF (IASPEI Seismic Format) data-exchange format that was discussed in the previous section allows for data lines (the ARRIVAL data type) to convey station-specific magnitude data. The amplitude, period, and amplitude-measurement time that is associated with a single magnitude measurement is transmitted on a single line, along with the calculated station-magnitude value. A “phase name” (see **Table 4**) must generally be associated with the data.

**Table 4** ISF “phase names” to be used for transmittal of amplitudes, periods, and amplitude-measurement times for the standard magnitude types considered in this paper. “I” stands for “International” or “IASPEI”, “A” for displacement amplitude, and “V” for velocity amplitude.

| Magnitude type | Phase Name |
|----------------|------------|
| ML             | IAML       |
| Ms_20          | IAMs_20    |
| Ms_BB          | IVMs_BB    |
| mb             | IAmb       |
| mB_BB          | IVmB_BB    |
| mb_Lg          | IAmb_Lg    |

### 3.5 Agency-specific circumstances that may lead to modification of standard procedures

The IASPEI Standard Procedures are proposed for implementation by seismological agencies in general. Adoption of the Standard Procedures by an agency will make that agency's magnitudes more useful to global seismological research and will enable the agency to directly compare its own results with IASPEI standard magnitudes produced by other agencies. The *Magnitude WG* recognizes, however, that agency-specific circumstances may make it desirable for individual agencies to use procedures that differ from the Standard Procedures. In this case, agency documentation and agency-produced bulletins should contain sufficient information to allow a user to understand how the agency's magnitudes can be related to magnitudes produced by the Standard Procedures.

For situations in which data demand modifications to the attenuation equations that are specified in the Standard Procedures, we recommend the practice that has been followed in the historic development of magnitude scales: the current standard equations are to be used as baselines for defining otherwise arbitrary constants in the improved equations. An agency

magnitude-type that is commonly, or in certain magnitude ranges, biased by more than 0.1 magnitude unit with respect to the magnitude produced by an IASPEI Standard Procedure should be identified by nomenclature that is distinct from the IASPEI magnitude nomenclature.

**Note:** The preceding material of this section (3) was reproduced with minor modification from IASPEI (2013). A following subsection in IASPEI (2011 and 2013), entitled “**DOCUMENT-ATION OF STATION/AGENCY MAGNITUDE PROCEDURES,**” has been attached in NMSOP-2 as Annex 2 to IS 3.4 “Guidelines for using the IASPEI standard magnitude reference data set”. This permits seismic stations/agencies to respond to this questionnaire in conjunction with the invitation to analyze the standard magnitude reference data set. In fact, this is the first practical step to educate and train station/agency personnel in the application of the IASPEI standard procedures in comparison with their traditional practices of magnitude determination.

#### **4 Relation of IASPEI standard procedures to original magnitude definition and to procedures for measuring magnitudes at different agencies**

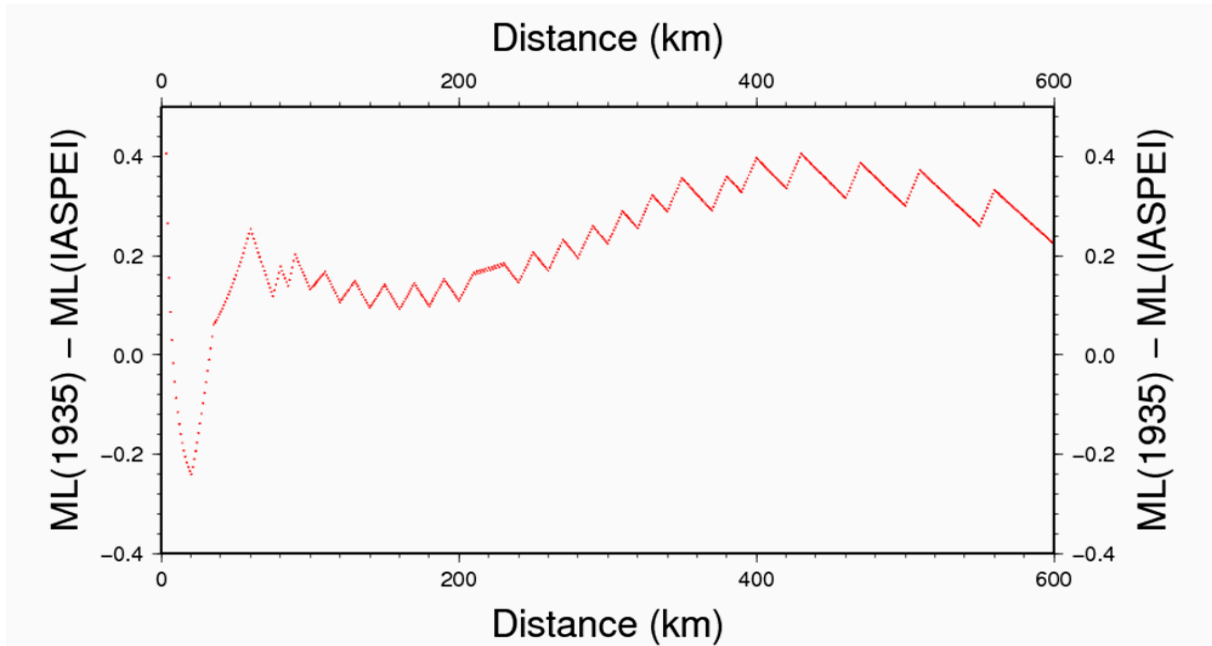
As summarized in section 2, the IASPEI recommended measurement standards should, as far as possible, produce magnitudes that are consistent with magnitudes of the same type that have been measured for decades from analog seismograms according to original definitions. This aims at assuring continuity of magnitude estimates published in older and younger earthquake catalogs. We outline below to what extent this has been achieved or, if not, what are the reasons and advantages for deviations.

##### **4.1 ML**

The local magnitude scale was developed by Richter (1935) for earthquakes in Southern California. The tabulated calibration values according to Richter (1958) are presented in DS 3.1. The more recent Hutton and Boore (1987) formula for Southern California, on which the standard ML(IASPEI) formula (1) is based, agrees well for hypocentral distances  $R$  between  $50 \text{ km} < R < 200 \text{ km}$  with the Richter calibration values, but produces higher values (up to 0.6 m.u.) for nearby stations ( $R < 50 \text{ km}$ ;) and lower (up to 0.24 m.u.) values for more distant stations up to 600 km (see also Fig. 3.30 in Chapter 3). The large effect at close distances is due to the fact that (according to Hutton and Boore) Richter assumed a geometric spreading with  $1/r^2$  whereas the observed spreading is  $\approx 1/r$ .

However, ML(IASPEI) differs from the Hutton and Boore (1987) ML by accounting for the difference in static magnification assumed by the manufacturer of the WA instruments (2800) and the one empirically determined by Uhrhammer and Collins (1990), which is about 2080. If ML(H-B) is measured on an original WA record then this difference does not matter. However, when compared with ML(H-B) on a WA simulation record one would expect ML(IASPEI) to be larger than ML(H-B) by  $\log(2800/2080) = 0.13 \text{ m.u.} = \text{constant}$ . Both effects together (i.e., the use of the Hutton and Boore calibration function and the correction for a constant difference in static magnifications) would produce in the epicentral distance range  $0 \text{ km} < D \leq 600 \text{ km}$  differences between ML(Richter 1935 synthesized) and ML(IASPEI) as depicted in Figure 2.





**Figure 2** Difference between ML(1935) and ML(IASPEI) as a function of epicentral distance when using, for ML(1935), the Richter (1958) tabulated calibration values (therefore the saw-tooth-pattern in the above figure) and **simulated** WA records assuming a static magnification of 2800 and, for ML(IASPEI), the calibration formula (1) with amplitudes measured on simulated WA records assuming a constant, empirically determined average static magnification of 2080 by Uhrhammer and Collins (1990). This accounts for a constant difference between these two magnitude values of  $\log(2800/2080) = 0.13$  m.u. (courtesy of K. Stammler, Federal Institute for Geosciences and Natural Resources, BGR, 2012).

However, as pointed out already by Kim (1998), the difference between the old and the empirically determined WA response **is not constant, but frequency-dependent and on average for frequencies at which ML is typically measured, about 0.1 m.u., i.e., 0.03 m.u. less than assumed in Figure 2**. According to D. Bindi (2011, personal communication) the difference varies from 0.065 m.u. at the WA eigenperiod of 0.8 s to 0.13 m.u. at about 8 times longer and shorter periods, respectively, because the damping parameter of the recalibrated WA is not 0.8 (as originally assumed) but 0.7 (see Table 1). For more details see *Comment 1.1* in DS 3.1.

#### 4.2 mB (mB\_BB)

mB is the original Gutenberg (1945b and c) body-wave magnitude. It was measured on relatively broadband medium-period instruments at periods between 2 s and 20 s (mostly between 5 s and 10 s) (see Abe and Kanamori 1980; Abe 1981 and 1984). The IASPEI recommendation to measure the maximum amplitude within the whole P-wave train, including also depth phases of P, follows the practice of Gutenberg. Gutenberg and Richter (1956a) published calibration functions for vertical and horizontal component amplitudes of P and PP waves as well as for horizontal component amplitudes of S waves amplitudes. The IASPEI standard uses only vertical component P-wave readings. For this reason, and to

reflect other modifications discussed in the following paragraphs, the IASPEI standard nomenclature is specified as  $mB_{BB}$ .

Gutenberg measured the vast majority of  $mB$  for earthquakes with magnitudes larger than 6.5. With current velocity BB instruments of high dynamic range it is possible to determine  $mB_{BB}$  down to about magnitude 4 (at best) or 5 (see Bormann et al. 2007 and 2009). The current instruments also permit reliable measurements to be made at periods above 20s. The new standards have therefore broadened the period range within which  $mB_{BB}$  may be measured beyond the earlier *de facto* range, to  $0.2 \text{ s} < T < 30 \text{ s}$ .

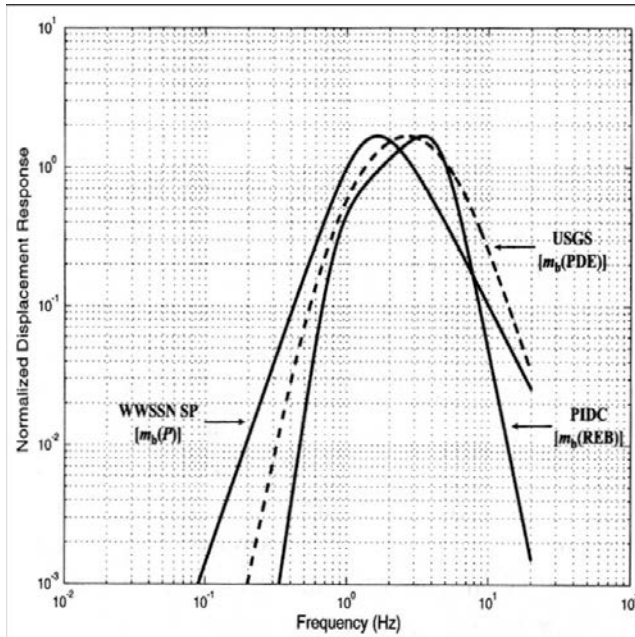
The other change is that Gutenberg measured the  $(A/T)_{\max}$  on displacement records. The standard, however, recommends to measure  $V_{\max}$  on velocity broadband records instead and use  $(V_{\max}/2\pi)$  as a proxy for  $(A/T)_{\max}$ . As a matter of fact, analysts tend to measure on displacement records  $A_{\max}/T$  instead of  $(A/T)_{\max}$ , yet the former often occurs at longer periods with the tendency to underestimate  $(A/T)_{\max}$  and thus  $mB$ , especially for smaller earthquakes. This has been confirmed by comparing traditional measurements of  $mB$  and related  $T$  at the Chinese Earthquake Network Center (CENC), based on Kirnos-type of displacement broadband records, with  $mB(BB)$  and related  $T$  measured on velocity broadband records. Despite these differences in old and new measurement practice the agreement between  $mB(\text{CENC})$  and  $mB_{BB}$  is on average 1:1 for magnitudes above 6 (the Gutenberg magnitude range); around  $mB_{BB} = 5$ , however, the classical  $mB(\text{CENC})$  is about 0.2 m.u. lower (see related Figures Chapter 3 or in Bormann et al., 2009).

In summary: Standard  $mB_{BB}$  agrees for values  $> 6.0$  well with Gutenberg's  $mB$  measured on displacement-proportional medium-period instruments. However,  $mB_{BB}$  is applicable in a wider range of periods and magnitudes than was usually used with traditional  $mB$ , is measured directly from  $V_{\max}$ , and is thus more closely and with less scatter related to released seismic energy than the traditional  $mB$ .

Therefore, routine measurement of broadband  $mB_{BB}$  should become a must in future, at least for earthquakes with magnitudes above 5.5. Although no knowledge of period is required for calculating  $mB_{BB}$  it should be measured and reported, because the period at which  $V_{\max}$  is observed is, on average, closely related to the corner period of the radiated source spectrum, increases with magnitude and decreases with increasing stress-drop and rupture velocity (see Chapter 3). While for smaller earthquakes recorded with low SNR in broadband records the measured periods may be strongly biased by the periods of dominating microseismic noise, Figure 3 reveals for magnitudes above 6 the expected exponential increase of period with magnitude.

### 4.3 mb

$mb$  was introduced into global observatory practice with the rapid increase in the number of sensitive short-period seismographs associated with the deployment of the World-wide Standard Seismograph Network (WWSSN), with the installation of arrays of short-period seismographs, and with regional deployments of short-period (SP) seismographs. They were mostly of the type of the WWSSN-SP response, of the later follow-up PDE-response used at the USGS/NEIC, the response used at the International Data Center (IDC, former Preliminary IDC = PIDC) of the CTBTO (see Figure 3) or covering the SP frequency range between about 1 Hz and 4 Hz in a similar way with relative bandwidth of only one to two octaves.



**Figure 3** Normalized displacement responses of a typical WWSSN short-period seismograph, of the band-pass filtered record output used by the USGS prior to 2009 for its automatic PDE processing procedure, and the filter response according to the PIDC (now IDC) procedure for short-period data analysis. All responses are normalized to have the same peak gain (modified to same abscissa-ordinate scaling as in Figure 1 from Granville et al., 2005, p. 1813, Fig. 4, © Seismological Society of America).

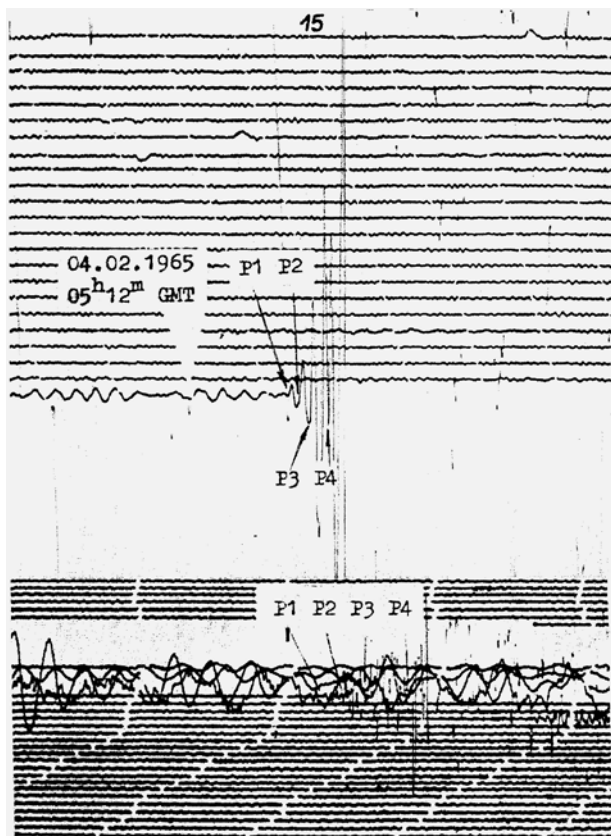
The high sensitivity of such band-limited short-period seismographs at quiet sites allowed for a dramatic lowering of the size threshold for which teleseismic body-wave magnitudes could be assigned. On the other hand it became clear that for strong earthquakes, mb measurements tend to be systematically less than other types of magnitude, and the effect is dramatic (more than 1 m.u.) for great earthquakes (e.g., Figure 4 in Kanamori, 1983). This effect is sometimes termed “magnitude saturation” (more correctly, a relative insensitivity to increase in the moment of the earthquake). It has two components, a spectral and a rupture-duration related one. Spectral “saturation” begins when the corner frequency of the radiated seismic source spectrum (e.g., Aki, 1967; Geller, 1976; Bormann et al., 2009; and Chapter 3), respectively of the recorded input wave spectrum, is lower than the corner frequency of the seismometer response. With reference to mb and Figure 3 this means that “saturation” is expected to begin earliest for mb(IDC) and latest for mb(WWSSN-SP). Yet mB\_BB, measured for periods up to 30 s, saturates accordingly much later than any mb. However, the IDC response allows to measure mb still reliably for smaller events than one can measure on WWSSN-SP records, because of its steeper roll-off and thus improved SNR towards lower frequencies and its peak magnification at 2 times higher frequency.

These comparisons give a feeling for the average influence of different SP seismometer responses on mb magnitude estimates. For the cases investigated they are significant ( $> 0.1$  m.u.) for the PIDC = IDC response. Another case will be discussed in section 5.1.2.

Yet, besides differences in seismograph responses used at different agencies and related differences in spectral “saturation” the application of different measurement time-window after the P-wave first onset may become even the dominating factor on the measured mb values. This is surely the case when the rupture duration of earthquakes and especially the time after the P onset at which the largest spectral amplitudes are released significantly exceed the measurement time-window within which  $(A/T)$  is measured. Bormann et al. (2009) derived a simple formula for calculating the average rupture duration  $T_R$  in its dependence on the largest event magnitude  $M$ :  **$\log T_R = 0.6M - 2.8$** . For individual earthquakes, however, the rupture duration may be up to about 2-3 times longer or shorter than estimated by this simple formula, due to differences in stress drop and rupture velocity.

In the first years of the WWSSN the United States Coast and Geodetic Survey (USCGS) instructed data analysts to measure  $A_{\max}$  in the P-wave train within the first 5 half-cycles. Somewhat later this was extended to within the first 5 s after the P-wave onset. The latter practice is still kept by the IDC of the CTBTO (5.5 s after the first P-onset; see Chapter 17). The reason for this time-window setting was that the WWSSN had as one of its main goals not only the best possible detectability of weak events but also the discrimination between natural earthquakes and underground nuclear explosions. And this discrimination capability was enhanced by measuring within such a short early P-wave time-window, within which explosions surely release their maximum of energy. This, however, is usually not the case for earthquakes with magnitudes larger than 6 for which rupture durations - on average - last for more than 6 s. This, probably, led the IASPEI Commission on Practice to propose in its first report of 1972 to extend the measurement time window for P-wave amplitudes to “15 or 25 s”.

Yet at some stations routine practice prior to 1972 was to always measure the maximum amplitude within the whole P-wave train. The first author, for example, introduced this practice at the seismological observatory Moxa (MOX), Germany, in 1965. Figure 4 shows a record example from 1965 (Bormann, 1969) and Tab. 3.1 of Chapter 3 gives a data example from the bulletin 1967 (Bormann and Stelzner, 1972).



**Figure 4** Displacement records of a 1965 Aleutian Islands earthquake at station MOX, Germany. **Below:** Records of type A (SP; 4 octave bandwidth); **Above:** Record of type B (medium period; 8 octave bandwidth). Note the multiple rupture process with  $P_{\max}$  arriving only 40 s after P1.  $mb1 = 5.8$  and  $mb4 = 7.0$  as compared to  $mB1 = 6.4$  and  $mB4 = 7.9$ . When compared with  $mB1$  and  $mB4$ , the spectral underestimation for the more short-period  $mb1$  is  $-0.6$  m.u., for the more long-period  $mb4 = -0.9$  m.u. (Bormann, 1969).

The main difference between the traditional Moxa practice and the currently proposed IASPEI  $mb$  procedure was that instead of the WWSSN-SP response the Moxa procedure used a more broadband SP type A record, displacement-proportional between 0.8 and 10 Hz. Additionally, in the case of a multiple rupture process, for all major subsequent P onsets, from the initial one up to  $P_{\max}$ , the amplitudes, periods and related magnitudes were measured both in the SP and medium-period (Kirnos SKD) records.

Experience thus collected led Bormann and Khalturin (1975) to state: „...that the extension of **the time interval for the measurement of (A/T)<sub>max</sub> up to 15 or 25 sec., ... is not sufficient** in all practical cases, **especially not for the strongest earthquakes...**” and to recommend in such cases an extension of the measurement to at least 1 min after the first P onset. This recommendation was accepted at the IASPEI meeting in 1976 and included in the Willmore (1979) edition of the Manual of Seismological Observatory Practice. This became also common practice at the USGS/NEIC. Yet, a few extraordinarily large earthquakes confirmed that the rupture duration may last even for several minutes and P<sub>max</sub> arrive even well after 60 s (see record examples in Bormann and Saul, 2008; 2009a). Therefore, Houston and Kanamori (1986) proposed to measure m<sub>b</sub> from the maximum amplitude over the whole P waveform recorded on WWSSN-SP records, without setting a fixed time-window limit, as now recommended by the new IASPEI standard.

A consequence of the different practices of m<sub>b</sub> measurement, both with respect to the frequency responses and measurement time windows used, has been rather different correlation relationships between m<sub>b</sub> and M<sub>s</sub> (see Chapter 3, section “Relationships between magnitude scales”) as well as between the m<sub>b</sub> values of different agencies. Bormann et al. (2007; Figure 2, courtesy of S. Wendt, 2003) showed that the difference between m<sub>b</sub>(PDE) and m<sub>b</sub>(PIDC) is on average +0.06 m.u. for m<sub>b</sub>(PDE) < 4.0, +0.38 m.u. for m<sub>b</sub>(PDE) between 4.0-4.9, +0.48 m.u. for m<sub>b</sub>(PDE) between 5.0-5.9 and 0.61 for m<sub>b</sub>(PDE) > 5.9, and that it reached +1.5 m.u. for the great Mw9.3 Sumatra 2004 earthquake with m<sub>b</sub>(PDE) = 7.2. The IASPEI (2005) recommended m<sub>b</sub> procedure yielded m<sub>b</sub> = 7.5 (Bormann et al., 2007) and thus a difference with respect to m<sub>b</sub>(IDC) of 1.8 m.u.

In contrast, the difference between m<sub>b</sub>(PDE), measured by the USGS prior to 2009 with the response shown in Figure 3, and the m<sub>b</sub>(IASPEI\_CENC) is much less, where CENC stands for China Earthquake Network Center, which tested extensively the application of the IASPEI standard procedures for magnitude measurement. According to Bormann et al. (2009) the orthogonal regression relationship is

$$m_b(\text{PDE}) = 0.90 m_b(\text{IASPEI\_CENC}) + 0.59. \quad (8)$$

Relationship (8) yields average differences, m<sub>b</sub>(PDE) – m<sub>b</sub>(IASPEI\_CENC), between +0.14 m.u. and -0.11 m.u. for m<sub>b</sub>(IASPEI\_CENC) between 4.5 and 7.0, but differences may reach 0.5 m.u. for still larger m<sub>b</sub> values because of the 60 s limit of the old USGS m<sub>b</sub> measurement time-window.

In summary: since standard m<sub>b</sub> is measured on simulated records with the same frequency response as early WWSSN records, a significant spectral saturation effect for larger magnitudes remains. However, by measuring A<sub>max</sub> in a variable, rupture-duration dependent measurement time-window, the further reduction of m<sub>b</sub> by using a too short fixed time-window is now avoided. Thus, as shown by Houston and Kanamori (1986) and Bormann et al. (2009), m<sub>b</sub> values up to about 7.5 can now be measured, giving a more realistic picture of the high-frequency energy release and thus the shaking potential of great earthquakes. For earthquakes with magnitudes below 5-6, however, routine measurement time-windows within the first 5 to 10 s after the P-wave introduce, as a rule, no significant bias. Slight differences of actual WWSSN-SP responses and related simulation filters from the adopted standard may result in average m<sub>b</sub> differences of about 0.03 m.u. (see, e.g., Figure 11) or, when rounded to the nearest 0.1 m.u., in rounding differences of maximum 0.1 m.u. for about 30% of the events.

#### 4.4 Ms\_20

Ms\_20 replicates the procedure by which Ms(PDE) has been computed for almost four decades by the USGS/NEIC. Ms\_20 is supposed to be closest to the original definition of the surface-wave magnitude by Gutenberg (1945a). This is empirically true yet requires some clarification about significant differences in measurement procedures and applied calibration function. Gutenberg's Ms formula reads, when A is measured in  $\mu\text{m}$ :

$$M_s = \log A + 1.656 \log \Delta + 1.818. \quad (9)$$

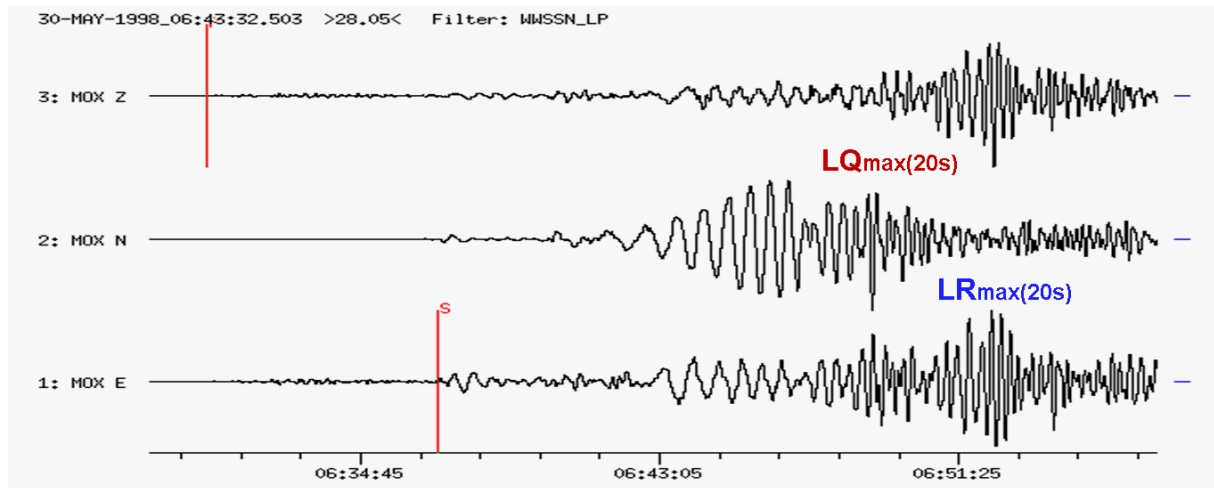
It is applicable between  $20^\circ \leq \Delta < 130^\circ$  with differences  $< 0.05$  m.u. as compared to respective tabulated values in Gutenberg (1945a) (see also DS 3.1 of this Manual). The latter, however, account better than the simple formula for the energy focusing of surface waves towards the antipodes. The differences between Ms values calculated for larger distances according to (9) or by using the tabulated calibration values are 0.07 m.u. at  $140^\circ$ , 0.12 m.u. at  $160^\circ$  and 0.55 m.u. at  $180^\circ$ .

Note, that formula (9) uses only displacement amplitudes A. Gutenberg himself did not note in his notebooks, at which period A had been measured, because he had defined his Ms scale for “periods of about 20 s”. As a matter of fact, surface-wave maxima in records of earthquakes with magnitudes  $> 7$  and those with dominatingly oceanic travel paths tend to have periods around 20 s. This was the case for most of the records analyzed at Pasadena in these early years. Moreover, Gutenberg (1945a) and Gutenberg and Richter (1956b) argued that a measurement period of about 20s, in preference to significantly shorter periods, would for earthquakes of focal-depth less than 40 km suppress significant underestimation of Ms that might result from strong depth-dependence of surface-wave excitation at the shorter periods. Yet, comparison with the original bulletins also of other stations used by Gutenberg as data sources for calculating the Gutenberg-Richter  $M_{GR}$  magnitudes in “Seismicity of the Earth” (Gutenberg and Richter, 1954) showed that periods as low as 12 s and as high as 23 s were sometimes used, with values outside of the period range 18 s to 22 s not being rare (Abe, 1981; Lienkaemper, 1984).

Further, the Gutenberg formula was derived for scaling horizontal component surface-wave readings, but Gutenberg did not determine the true maximum vector sum of the surface-wave amplitudes in the N-S and E-W component measured at the same time (or within  $1/4^{\text{th}}$  to  $1/2^{\text{th}}$  of the measured period, as commonly practiced or assumed). Rather, Gutenberg “vectorially” combined the largest surface-wave amplitudes measured in the two horizontal components although they may arrive at rather different times and belong to different types of surface waves. This may, in the extreme, result in 0.15 m.u. too large magnitude estimates, if both Rayleigh (LR) and Love waves (LQ) have comparably large amplitudes and the backazimuth of wave approach is  $0^\circ + n90^\circ$  with  $n = 0, 1, 2$ , or 3. An example is shown in Figure 5.

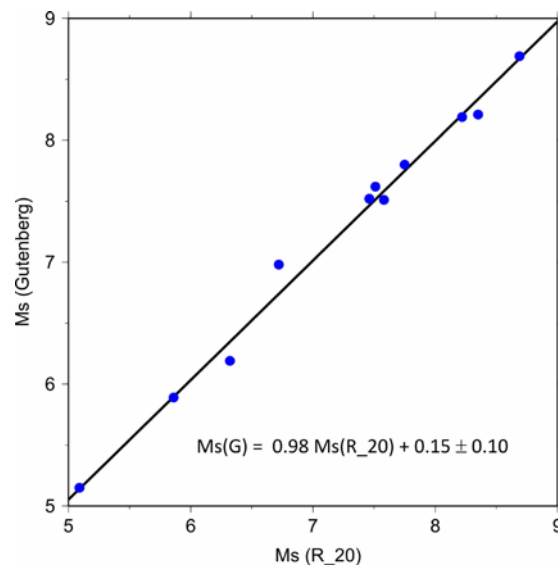
The U.S. Coast and Geodetic Survey, predecessor to the USGS/NEIC, began publishing Ms\_20 values in 1968. The U.S. agency formalized the Gutenberg tendency to prefer waves with periods near 20s (the USGS/NEIC uses periods in the 18s – 22s range) but adopted an IASPEI recommendation that the magnitude formula be that of Vaněk et al. (1962)

$$M_s = \log_{10}(A/T)_{\text{max}} + 1.66 \log \Delta + 3.3, \quad (A \text{ in } \mu\text{m}). \quad (10)$$



**Figure 5** The difference between  $M_s$  derived from these records according to the Gutenberg “vectorial” combination  $A_{Hmax} = (LR_E^2 + LQ_N^2)^{1/2}$  and the true  $A_{Hmax}$  of LR is +0.12 m.u. This is about the largest difference possible (0.15 m.u.) and depends on the BAZ (here  $85^\circ$ ).

The U.S. agency procedure also suggested that contributing observatories measure horizontal component surface-wave amplitudes measured at simultaneous arrivals (Lienkaemper, 1984), thus allowing to calculate the true vectorial  $A_{Hmax}$ . The latter change, together with some other factors explained by Lienkaemper (1984), makes  $M_{s\_20}$  empirically compatible with the classical Gutenberg-Richter (1954) magnitudes in “Seismicity of the Earth”. This is confirmed by Figure 6. For more detailed discussion see Chapter 3.



**Figure 6** Comparison of  $M_s$ (Gutenberg, 1945a) and  $M_{s\_20}$ (IASPEI) when vectorially combining for the former the largest N-S and E-W component amplitude readings and for the latter the largest radial (R) surface-wave component amplitude (courtesy of S. Wendt, 2011).

The introduction of stable long-period vertical component (V) seismometers into global seismological practice in the 1960s introduced another change in  $M_s$  measurement. According to Hunter (1972) the differences between  $M_s(V)$  and  $M_s(H)$  are negligible, as confirmed by Bormann and Wylegalla (1975). According to their orthogonal regression relationship

$$MLV = 0.97MLH + 0.19, \quad (11)$$

the two magnitudes differ on average in the range from 4 to 8.5 less than 0.07 m.u.

From May 1975 onwards, the USGS decided to calculate their  $M_s$  exclusively from vertical component readings. Since that time,  $M_s(\text{NEIC})$  is in fact identical with the IASPEI confirmed standard magnitude  $M_{s\_20}$ .

An important feature of the IASPEI  $M_{s\_20}$  formula, equation (3), is that the original formula (10) was derived using data collected over a broader distance range and over a much broader range of periods than the ranges to which the formula is restricted in equation (3). A number of studies, cited in 5.3.3, have suggested that the restriction of equation (3) to periods near 20s produces a distance-dependent bias within the  $20^\circ$  --  $160^\circ$  distance range to which equation (3) is restricted by the  $M_{s\_20}$  formula. Alternative formulas have been proposed to reduce this bias and to allow extension of the  $M_{s\_20}$  procedure to distances less than  $20^\circ$ .

$M_s$  based on surface waves near 20s are computed by the CTBTO/IDC with an alternative formula (equation 18 of Rezapour and Pearce, 1998) in place of the IASPEI formula, and using data from distances  $2^\circ$  --  $100^\circ$ , instead of  $20^\circ$  --  $160^\circ$  (Stevens and McLaughlin, 2001). The  $M_{s\_20}(\text{IDC})$  values are published with some delay in the online bulletin of the ISC (International Seismological Centre, <http://www.isc.ac.uk/iscbulletin/search/bulletin/>, last accessed March 2012). The IDC computes  $M_{s\_20}(\text{IDC})$  for nearly an order of magnitude more events than the USGS/NEIC. For events in common, the  $M_{s\_20}(\text{IDC})$  values tend to be about 0.1 magnitude units smaller than  $M_{s\_20}$  produced by the USGS/NEIC, although for stations in common the amplitudes computed at the IDC and USGS/NEIC agree to within several hundredths of a magnitude unit on average (Stevens and McLaughlin, 2001).

#### 4.5 $M_{s\_BB}$

Besides the U.S. preferred practice for calculating  $M_{s\_20}$ , there has long existed another tradition of  $M_s$  determination in a much wider range of periods between about 2 and 25 s. Even in pre-WWII bulletins of many stations world-wide that were equipped with classical mechanical or electromagnetic medium-period broadband seismographs of Wiechert, Mainka, Bosch-Omori, Galitzin, or similar, one regularly finds readings of surface-wave maxima at periods outside the range of 18-22 s. Since the 1950s, first-rate stations deployed all over the former Soviet Union (FSU) territory and later also in most FSU allied countries of Eastern Europe, as well as those in Mongolia, China, and Cuba have been equipped with displacement proportional broadband seismographs of type Kirnos. The earlier version (SK) had a passband between about 0.1 to 10 s, the later version (SKD) between 0.1 and 20 s (see Figure 10). Using records of these instruments, already Solov'ev (1955) presented empirical evidence that the ratio  $(A/T)$  is a stable quantitative characteristic of surface wave maxima, independent of period in a wide range from local to teleseismic distances.

$M_{s\_BB}$  is calibrated with formula (4), only slightly modified from the original formula (10), for inserting directly measured velocity amplitudes in nm/s. The original formula (10) was published by a team of Czech and Russian authors (Vaněk et al., 1962; Karnik et al., 1962) and accepted by IASPEI in 1967 as the standard formula for  $M_s$  calculations.



Although formula (10) was originally proposed for the distance range  $2^\circ$  to  $160^\circ$  subsequent work indicates that it is more appropriate to use for distances larger than  $140^\circ$  the tabulated values published by Kondorskaya et al. (1981), which are reproduced in Table 4 of DS 3.1 in this Manual. The reason is the same as the reason that Lienkaemper (1984) used at distances  $>125^\circ$  instead of Gutenberg's (1945a) formula (9) the tabulated calibration values published in Richter (1958) (see Table 3 in DS 3.1), because they better account for the energy focusing effect towards the antipodes.

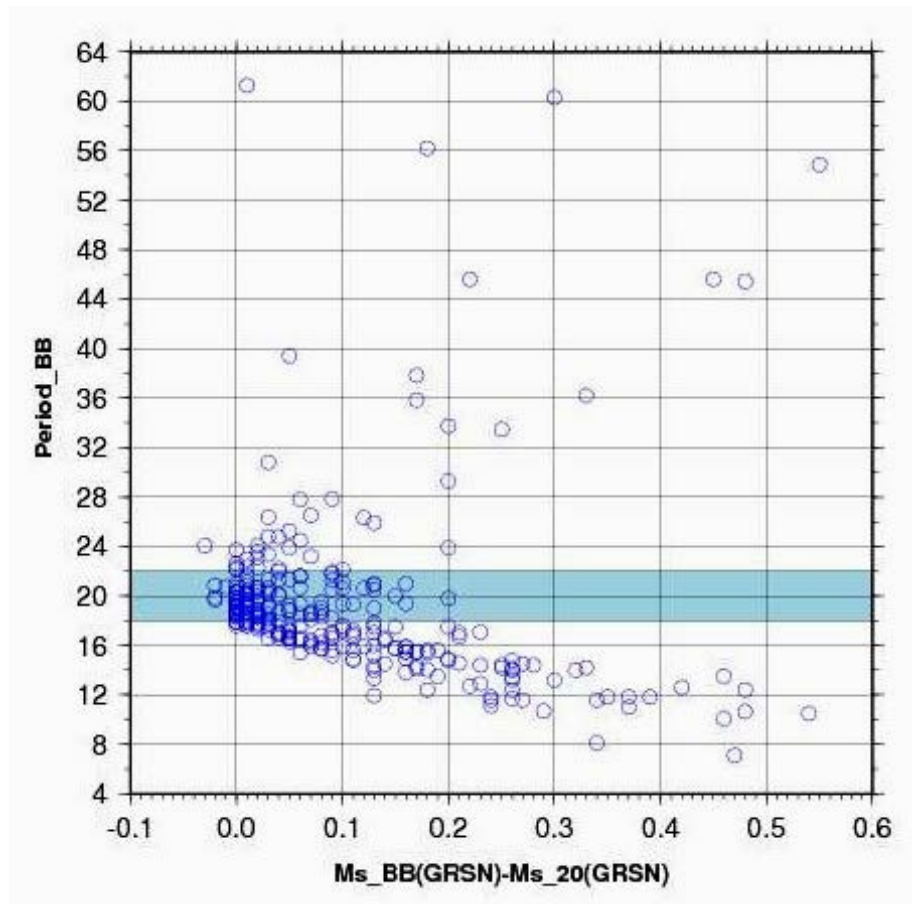
The Czech and Russian team that developed equation (10) (Vaněk et al., 1962; Karnik et al., 1962) measured the surface-wave maxima in a wide period range between some 2 s and 25 s. In Chapter 3 it is shown that  $V_{max}$  recorded in modern very broadband records may occur, although rarely, at periods well beyond 25 s, up to about 60 s. Accordingly the period range for  $M_s\_BB$  measurements has been extended now so that it is  $3\text{ s} < T < 60\text{ s}$ . The Czech and Russian team showed that the prevailing periods depend on distance (Vaněk et al., 1962). Their related Table (see Table 6 in DS 3.1) has been reproduced in the Wilmore (1979) Manual of Seismological Observatory Practice. The Wilmore (1979) Manual also includes the recommendation: "When the period differs **significantly** from the values in Table 3.2.2.1, it may be advisable not to use the data for magnitude determination." The issue of use of the  $M_s\_BB$  formula beyond the distance-dependent period ranges indicated in Table 6 of DS 3.1 still deserves special study.

The IASPEI proposed procedure for  $M_s\_BB$  determination has been extensively tested at the CENC, where it was applied to some 10,000 station readings of globally distributed earthquakes recorded by the broadband China National Seismic Network (CNSN) in the distance range between  $2^\circ$  and  $100^\circ$  (Bormann et al., 2009). Despite the large scatter of station-site period readings the general trend and range of periods published by Vaněk et al. (1962) has been confirmed down to local-regional distances and periods as short as 2 to 5 s. Additionally a trend of period increase with magnitude was evident (see also related Figures and discussions in Chapter 3). It is notable that the variance of  $M_s\_BB$  resulting from this test was somewhat smaller than the variance of  $M_s\_20$  (Table 5).

**Table 5** Average standard deviations of station magnitude estimates from average network event magnitudes for different scales. Data were taken from Bormann et al. (2007 and 2009). GRSN = German Regional Seismic Network, data by courtesy of S. Wendt; CNSN = China National Seismic Network. Note that CNSN SD values for  $M_s\_20$  and  $M_s\_BB$  relate to readings of equivalent Chinese magnitude standards of  $M_s7$  and  $M_s$ , respectively.

| Magnitude | Average SD (GRSN) | Average SD (CNSN) |
|-----------|-------------------|-------------------|
| mb        | 0.21              | 0.23              |
| mB_BB     | 0.15              | 0.21              |
| $M_s\_20$ | 0.12              | 0.26              |
| $M_s\_BB$ | 0.10              | 0.23              |

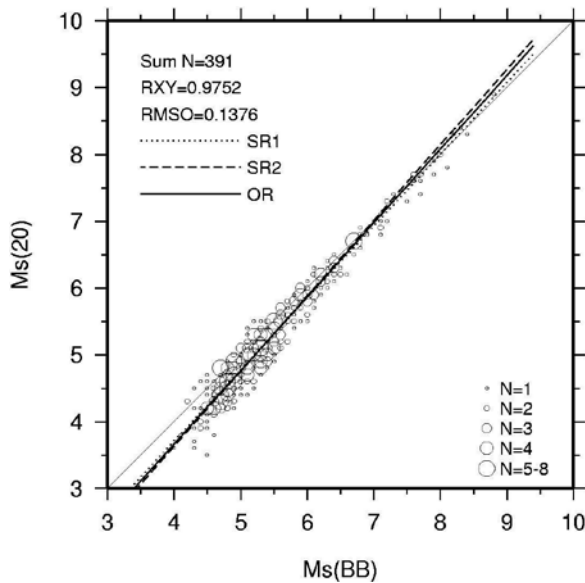
Although  $M_s\_BB$  and  $M_s\_20$  agree well whenever  $T$  at  $V_{max}$ , respectively  $A_{max}$ , is close to 20 s, differences may reach 0.5+ m.u. when  $T$  differs (depending on magnitude, travel-path and distance) significantly from 18-22 s (Figure 7).



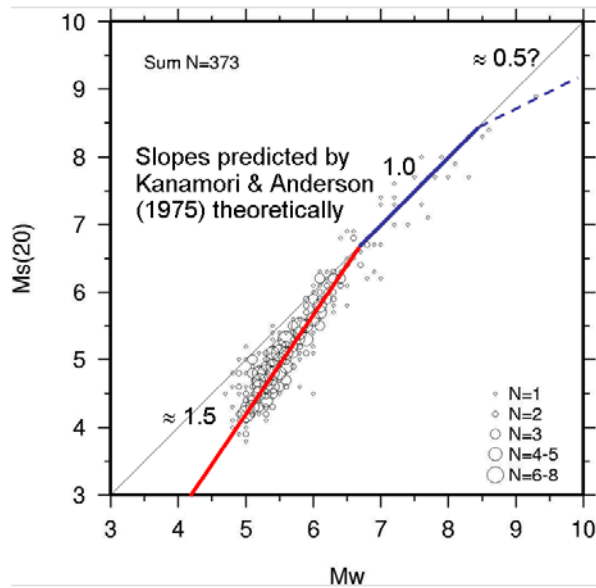
**Figure 7** Difference between  $M_s\_BB$  and  $M_s\_20$  in relation to the period at which  $V_{max}$  was measured on the velocity broadband records. Data are based on GRSN station record readings (courtesy of S. Wendt, 2008).

In the study of Bormann et al. (2009) most of the  $V_{max}$  values of Rayleigh surface waves were observed outside the period range 18–22 s. This holds for 79% of the analyzed CNSN records and for 62% of the analyzed GRSN records (see Figure 7), although the latter include many more oceanic travel paths and rarely any record at epicenter distances below  $13^\circ$ , which are much more frequent in Chinese network records and typical for rather short-period Rayleigh-wave maxima on continents.

Since the two magnitudes are calculated with the same numerical attenuation relation, but surface-wave amplitudes from which  $M_s\_BB$  can be calculated are commonly larger at periods less than 18 s, the difference  $M_s\_BB - M_s\_20$  for earthquakes of  $M_s\_BB < 6.5$  tends to be 0.1 to 0.2 magnitude units, increasing with decreasing magnitude. But  $M_s\_20$  and  $M_s\_BB$  agree on average rather well for magnitudes above 6.5 (Figure 8). Similarly, both magnitudes agree on average rather well with  $M_w$  for magnitudes between 6.5 and 8+, but for  $M_w < 6.5$  the discrepancy between  $M_s\_20$  and  $M_w$  is about twice as large as that between  $M_s\_BB$  and  $M_w$ . The magnitude “bias” between  $M_s\_20$  and  $M_w$  had already been predicted by Kanamori and Anderson (1975) on the basis of theoretical dimension analysis and was empirically confirmed by Ekström and Dziewonski (1988). Recent data by Bormann et al. (2009) fully agree with the theoretically predicted trends for  $M_s\_20$  (Figure 9), but also showed that  $M_s\_BB$  reduces the  $M_s\_20$  “bias” for  $M_w < 6.5$  by half.



**Figure 8** Standard regressions SR1 and SR2 as well as orthogonal regression between  $M_s_{20}$  and  $M_s_{BB}$ .  $R_{XY}$  – correlation coefficient,  $RMSO$  – orthogonal root-mean-square error. Cut-out from Figure 11 in Bormann et al. (2009), Bull. Seism. Soc. Am., **99** (3), p. 1881, © Seismological Society of America.



**Figure 9** Comparison between CNSN data of standard  $M_s_{20}$  with Global Centroid Moment Tensor solutions  $M_w$  and the theoretically predicted trends by Kanamori and Anderson (1975). Since, according to Figure 14,  $M_s_{BB}$  tends to be larger than  $M_s_{20}$  for  $M_s_{BB} = M_w < 6.5$ , its bias with respect to  $M_w$  is halved in this range.

However, what is or is not considered a bias of a given scale in relation to another is often a matter of perspective or preference given to certain more physically based seismic parameters such as seismic moment or released seismic energy. For example, the reason that  $M_s_{20}$  scales as 1.5  $M_w$  for small earthquakes is that  $M_s_{20}$  scales then as  $\log M_0$  for these earthquakes. So  $M_s_{20}$  behaves in a very predictable way with respect to the moment of small earthquakes, and many seismologists find this desirable. Developers of the  $M_s(V_{max})$  magnitude (Bonner et al., 2006) specifically emphasize the proportionality of  $M_s(V_{max})$  to  $\log M_0$  as evidence that  $M_s(V_{max})$  is a valuable magnitude. Future comparison of  $M_s_{BB}$  and  $M_s(V_{max})$  data, which are now routinely calculated at the USGS/NEIC, will better reveal the benefits or drawbacks, compatibility or complementarity of these two surface-wave magnitudes based on variable period data that are obtained by very different methodologies. In summary: of the two currently adopted IASPEI  $M_s$  procedures,  $M_s_{BB}$  agrees best with the original definition and intention of the  $M_s$  equation derived by Vaněk et al. (1962).  $M_s_{BB}$  is simply measured on unfiltered velocity broadband records, and several studies have shown that it has lower standard deviations than  $M_s_{20}$ (IASPEI). It is also applicable to many more surface wave records in a wider (down to local) distance range than  $M_s_{20}$ (IASPEI). Moreover, for magnitudes below 6.5  $M_s_{BB}$  is closer to  $M_w$  than  $M_s_{20}$ , irrespective of whether or not this is considered desirable.

#### 4.6 $mb_{Lg}$

Nuttli (1973) originally developed  $mb_{Lg}$  for eastern North America, and he proposed  $mb_{Lg}$  equations for that region that would be reasonable approximations to equations having the form of equations (7) and that are linear in  $\log(\text{epicentral distance})$ . The approximate

equations of Nuttli(1973) implicitly incorporate an eastern North America attenuation function, and the equations are used by many agencies to assign  $mb\_Lg$  to eastern North American earthquakes. The approximate equations of Nuttli (1973) yield  $mb\_Lg$  that are about 0.1 magnitude units smaller than those of the proposed standard procedure, when  $\gamma = .00063 \text{ km}^{-1}$  in equations (7). We propose the procedure of Nuttli (1986) as standard, in preference to the procedure of Nuttli(1973), because the former is transportable to regions with attenuation different than that of eastern North America (Nuttli, 1986; Patton, 2001).

#### 4.7 $M_w$

The recommended standard formulas for calculating  $M_w$  via seismic moment  $M_0$  follows directly from the way how it has been systematically derived from the basic assumptions made about the ratio between seismic energy and scalar seismic moment release, average rigidity and stress drop in the source volume and the Gutenberg-Richter relationship between  $\log E_s$  and  $mB$  on the one hand and  $mB$  and  $M_s$  on the other hand (see Bormann and Di Giacomo, 2011).

### 5 Deviations from the recommended measurement standards

The measurement standards specify the frequency response, the period range of measurement, the measurement time-window, the calibration function to be used and how amplitude measurements should be made. Inevitably, different seismological centers will see the desirability to adjust the procedures to address special circumstances. Paradoxically, a chief value in defining standard procedures is to codify a baseline procedure from which the desirability of future improvements to the procedures may be examined. In the following we illustrate examples of the types of *changes in standard procedures that may be acceptable without necessitating that the magnitudes computed with the changed procedure be assigned a different nomenclature*. An alternative procedure may also be preferable to a standard procedure in a particular region if it is more consistent with regional attenuation properties. In this last situation, however, magnitudes computed with the alternative procedure should usually be described with a different nomenclature than magnitudes computed with the standard procedure. For some guidance see IS 3.2.

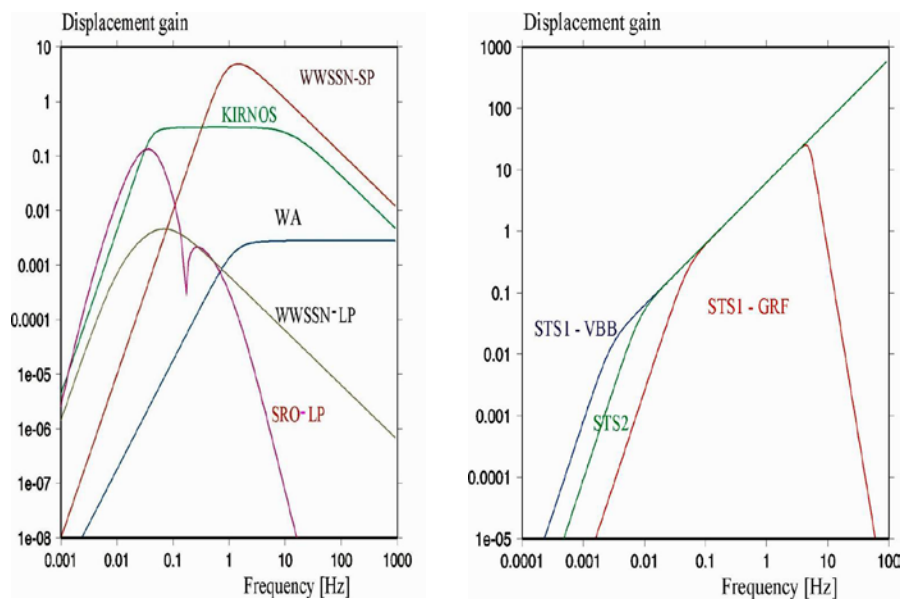
The first-order criterion to judge whether any alternative procedure is equivalent to a standard procedure is whether or not the alternative procedure yields in a wide magnitude range results that deviate on average less than 0.1 m.u. from those resulting from the strict application of the proposed measurement standards. Alternative proposed measurement parameters will for many purposes be acceptable as substitutes for the standard procedures if it can be confirmed by a representative comparative data set (see IS 3.4), that they do not significantly bias the magnitudes beyond this limit. From a practical standpoint, a changed procedure may be preferable to a standard procedure if, in addition to producing results that are not biased with respect to a standard procedure, it produces results with less scatter than the standard procedure or if it allows unbiased magnitudes to be determined for distance ranges or seismic-noise conditions in which the standard procedure is not appropriate.

The measurement of amplitudes and periods from digital signal by an automatic computer algorithm poses problems that are not addressed by the IASPEI standard procedures. Automatic pickers may require modification of the procedures in order to minimize picking of amplitudes in noisy or prior-shock-contaminated seismograms that would be rejected as

unsuitable by a seismologist who is making measurements with interactive processing software. In cases when a standard procedure is modified to suppress measurements made from noise, the suitability of the modification should be tested with respect to magnitudes that are measured interactively.

### 5.1 Variations in filter parameters acceptable for determining standard ML, mb, and Ms<sub>20</sub>

Figure 10 depicts the responses of seismographs represented in Figure 1 along with responses of several additional seismographs from which magnitude measurements have commonly been made. Unlike Figure 1, where gains are normalized to unity, the relative values of gains in Figure 10 are the ones implemented in the data acquisition and processing system at the Central Seismological Observatory SZGRF of the BGR in Germany (<http://www.szgrf.bgr.de/>; last accessed 11 March 2012) and the Seismic Handler Motif (SHM) software. We will in the following discuss their suitability for determining standard magnitudes.



**Figure 10 Left:** Relative displacement gain of simulation filter responses implemented in the Seismic Handler (SHM) software for analyzing records of the Gräfenberg (GRF) array and of stations of the German Regional Seismic Network (GRSN); **right:** Responses of different generations of velocity broadband seismographs STS1(GRF) (old version used at the GRF array), STS1 (VBB) (advanced version as used in the IRIS global network) and STS2. Figure modified from Chapter 11 to same ordinate-abscissa scale.

#### 5.1.1 ML

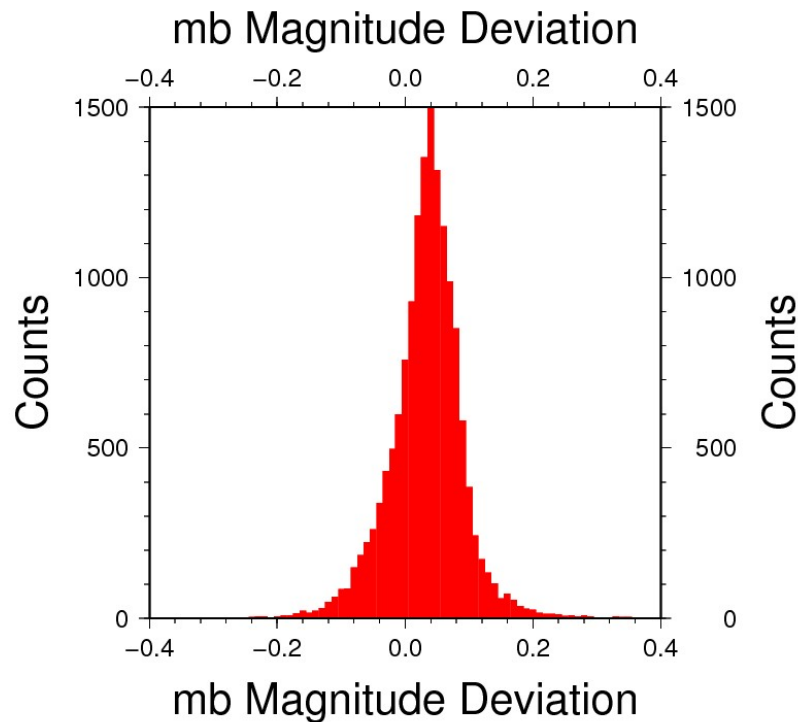
According to Figures 1 and 10 (left-hand panel) the response of the Wood-Anderson (WA) seismograph decays with decreasing frequency below 0.5 Hz only with the second order. Accordingly, sea or ocean storm microseisms with frequencies between  $0.1 \text{ Hz} < f < 1 \text{ Hz}$  are

not very efficiently suppressed and the signal-to-noise ratio in WA records may become very low for small earthquakes. By increasing the corner frequency of the WA high-pass simulation filter, and/or by steepening its roll-off towards lower frequencies, the SNR may be significantly improved and enable the measurement of ML correctly down to very small earthquakes, provided that the corner frequency of the signal spectrum is still well within the plateau of the modified WA response. Uhrhammer et al. (2011) have used this strategy to enhance SNR for purposes of measuring ML in California.

### 5.1.2 mb

In section 4.3 we have already extensively discussed the effect of different short-period responses, presented in Figure 3, on mb estimates. With reference to the average seismic source spectrum presented in Figure 1 of Bormann et al. (2009) (see also Chapter 3) we concluded that for magnitudes between about 3.5 and 5.5, the corner frequency of the radiated source spectra and thus the maximum of radiated seismic energy is expected to vary on average between some 0.5 and 4 Hz. When both the relative bandwidth as well as the response peak frequencies within this range differ significantly from the WWSSN-SP standard response then this will result in mb estimates that may differ more than 0.1 m.u. from the IASPEI standard mb as it is the case for mb(IDC). However, the PIDC response (see Figure 3) allows to measure mb still reliably for smaller events than one can measure on WWSSN-SP records, because of its steeper roll-off and thus improved SNR towards lower frequencies and its peak magnification at 2 times higher frequency. Thus it may be an acceptable substitute for the IASPEI standard WWSSN-SP response for  $mb < 4$ . A rigorous comparative test is not yet available but would be worthwhile to undertake.

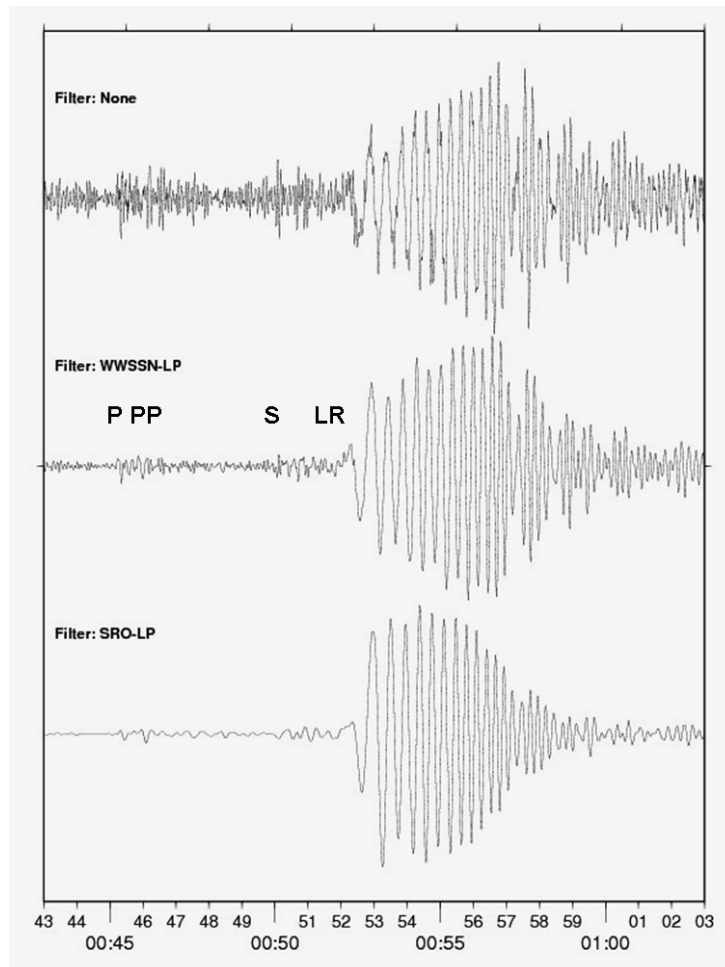
However, also transfer functions for classical WWSSN short-period seismograph differed somewhat for different magnification settings. In addition, different WWSSN-SP simulation filters may have been implemented in different analysis software. For example, the Seismic Handler Motif (SHM) software, developed by Klaus Stammli, BGR Hannover, assumes beyond the response peak only a first-order decay of gain with higher frequencies (see left panel of Figure 10 and poles and zeros in Tab. 11.3 of Chapter 11), in contrast to the decay with second order for frequencies  $f > 3$  Hz of the IASPEI adopted WWSSN-SP standard response (Figure 1 and Table 1 in this IS). This effectively means a slight reduction of the relative bandwidth RBW and thus of the measured amplitudes (see Chapter 4). In conjunction with the upgrading of the SHM software to accommodate the new IASPEI standard responses Klaus Stammli filtered about 15,000 P waveforms both with the old SHM and the new IASPEI WWSSN-SP response. He found that mb(IASPEI) would on average be 0.033 m.u. smaller than mb measured on old SHM WWSSN-SP simulated records (with a standard deviation of  $\pm 0.06$  m.u.) (see Figure 11). Thus, filter-wise the older mb data reported by the BGR to international data centers would still be within the tolerance limit of 0.1 m.u. and thus be acceptable as standard mb values, provided that the variable measurement-time window and peak-to-trough/2 rules have been applied as well. The latter is generally the case in the interactive SHM mode.



**Figure 11** Difference between mb values calculated with amplitudes and periods measured on simulated WWSSN-SP records filtered with slightly different responses: mb(BGR) with the WWSSN-SP response in the left panel of Figure 10 and the poles and zeros in Table 11.3 ?? of Chapter 11 and mb(IASPEI) with the respective response curve and parameters in Figure 1 and Table 1 of this IS 3.3.

### 5.1.3 Ms<sub>20</sub>

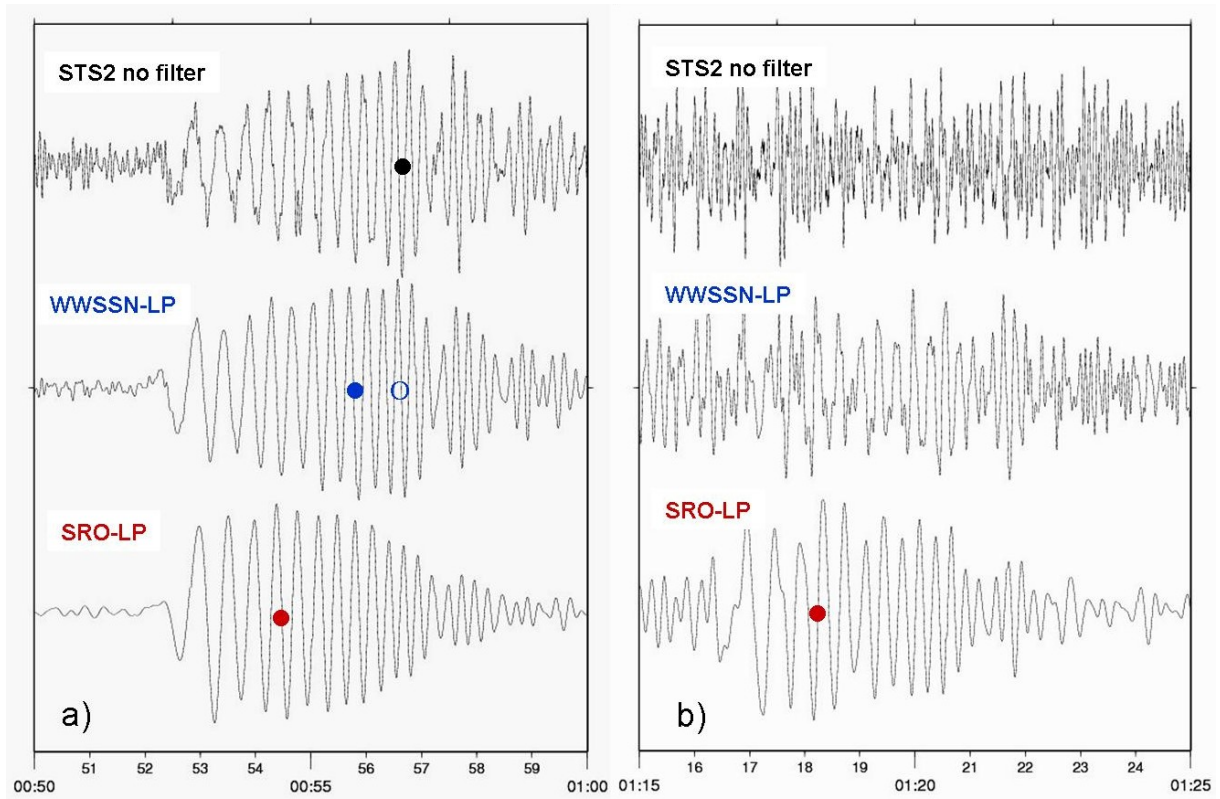
According to the standard procedure Ms<sub>20</sub> should be measured on records with the WWSSN-LP response. Yet the instrument's gain decays with increasing frequency only by first order. Therefore, ocean microseisms with periods between approximately 4 to 8 s may result in very low SNR for measurement of surface-waves with periods around 20 s from weak earthquakes. This limits the lower magnitude threshold down to which the mb-Ms<sub>20</sub> criterion for discriminating between natural earthquakes and underground nuclear explosions can be applied (see Fig. 11.22 ??? in Chapter 11 of this Manual). Therefore, a former global U.S. network of Seismic Research Observatories (SRO), operating between 1974 and 1993, used a modified SRO-LP response with much steeper flanks and an additional 6 s microseism notch-filter. This resulted in a rather narrow high-gain 20 s band-pass (see left-hand panel of Figure 10). The difference in shape as well as in the poles and zeros with respect to the WWSSN-LP response is striking (see Table 11.3 ??? in Chapter 11 of this Manual). When compared with WWSSN-LP and velocity broadband records, SRO-LP records look rather smooth and less noisy (Figure 12).



**Figure 12** Comparison of an unfiltered STS2 velocity broadband record of an earthquake in the North Atlantic Ocean at station CLL ( $D = 26.8^\circ$ ,  $mb = 5.7$  (GSR)) with the respective WWSSN-LP and SRO-LP filtered records (courtesy of S. Wendt, 2012).

One might expect from so different responses and record appearance significant differences in long-period magnitude estimates. This, however, is not the case for the band-limited  $M_s_{20}$ . Its measurement periods between 18 and 22 s are comparably well covered by the peak-response ranges of both WWSSN-LP and SRO-LP. Therefore, the average relationship between  $M_s_{20}$  measured on WWSSN-LP and SRO-LP records is indeed 1:1 over the large range of 4 magnitude units with a standard deviation of only 0.03 m.u. (Figure 14), although the respective surface-wave record trains might look quite different (Figure 13a). The data scatter in Figure 14 would only increase somewhat, if WWSSN-LP records with rather low SNR and thus significant amplitude reading errors would be included into the statistics as well. But for surface-wave trains with  $SNR \leq 1$  in unfiltered velocity and/or WWSSN-LP records it might be possible to determine a reasonably good  $M_s$  estimate only from SRO-LP records (Figure 13b).



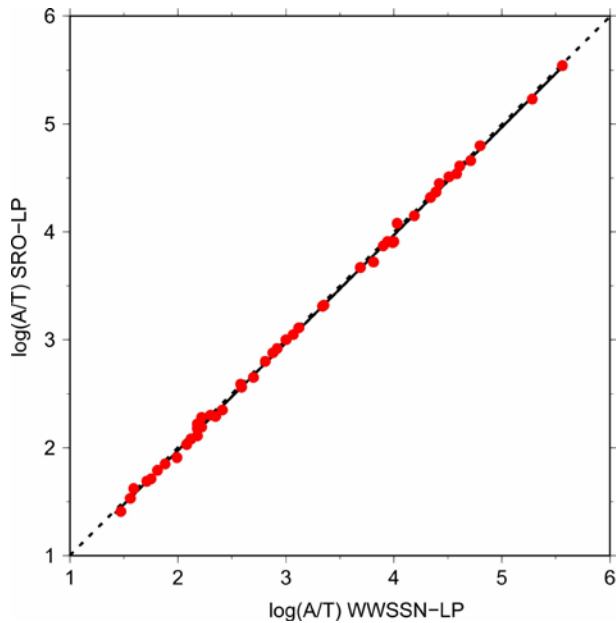


**Fig. 13** Unfiltered velocity-proportional STS2 broadband records and related WWSSN-LP and SRO-LP filtered records of the surface-wave train of two earthquakes in the North Atlantic Ocean recorded at station CLL, Germany:

- a)** 06 Oct. 2011, OT = 00:39:31.0, epicentral distance  $D = 26.8^\circ$ ,  $m_b = 5.7$ ,  
 $M_s_{BB} = 5.6$ ,  $M_s_{20}$  (WWSSN-LP and SRO-LP) = 5.5;
- b)** 06 Oct. 2011, OT = 01:33:33.2, epicentral distance  $D = 26.7^\circ$ ,  $m_b = 4.5$ ,  
 $M_s_{20}$  (SRO-LP) = 4.5.

The solid dots mark the amplitude measurement times for  $M_s_{BB}$  (black) and  $M_s_{20}$  (blue and red). The open blue circle marks the maximum trace amplitude on the WWSSN-LP record at  $T = 15.7$  s, which would correspond with  $V_{max}$  on the unfiltered STS2. (Acknowledgment: the figure has been compiled and complemented based on records that were kindly provided by S. Wendt (2012).

From Fig. 13a) one recognizes that the trace  $A_{max}$  is measured earliest on the SRO-LP record at  $T = 20.3$  s and about 80 s later on the WWSSN-LP record within the  $M_s_{20}$  period window at  $T = 18.9$  s. But the magnitude values are the same, although both being smaller by 0.1 m.u. than  $M_s_{BB}$ , which is measured at  $T = 15.9$  s, about two minutes later than on the SRO-LP record. One should note, however, that in fact also the maximum amplitude in the WWSSN-LP surface-wave train does not occur within the 18–22 s period window, but rather at  $T = 15.7$  s, i.e., close to the period of 15 s at which the WWSSN-LP response has its largest magnification. And when determining  $\log(A/T)_{max}$  with  $T = 15.7$  s then one gets the same  $M_s = 5.6$  as directly with  $M_s_{BB}$ . Thus, limiting  $M_s$  determination on WWSSN-LP records strictly to the narrow period range of 18–22 s is rather arbitrarily fixed to the 20 s spectral amplitude maximum and not related to the maximum record trace amplitude. Yet the 20 s spectral amplitudes are best recorded by a proper SRO\_LP filter centered at 20 s.



**Figure 14** The 1:1 relationship between vertical component maximum Rayleigh wave  $\log(A/T)$  values measured at periods between 18 and 22 s on SRO-LP and WWSSN-LP records using the IASPEI standard magnitude reference data set (see IS 3.4 of this Manual; courtesy of Siegfried Wendt, 2011).

Figure 14 confirms that  $M_s_{20}(\text{SRO-LP})$  is fully compatible with standard  $M_s_{20}$ . Accordingly, its amplitudes should be reported as  $IAMs_{20}$ . Since SRO-LP records have in any event a better SNR than WWSSN-LP records this would even justify to declare SRO-LP as the best suited response for standard  $M_s_{20}$  measurements. However, up to now, this filter is less frequently implemented than WWSSN-LP in seismic waveform analysis programs. An exception is the program Seismic Handler (SHM) by Klaus Stammer which can be downloaded from <http://seismic-handler.org/portal>. Chapter 11 and DS 11.1 to 11.3 present many SRO-LP filtered records in comparison to other standard filtered and unfiltered BB records.

In summary: For  $M_s_{20}$  determination with the predefined 18 to 22 s period range for measuring A the exact implementation of the WWSSN-LP standard response parameters is not critical, provided that these periods fall within the passband range of any alternative response. Note, however, that neither SRO-LP nor WWSSN-LP narrowband records are suitable to measure correct broadband  $M_s_{BB}$ , which necessitates records with a velocity proportional response which covers the whole range of periods for which this magnitude may be calculated ( $T = 3$  to 60 s), yet at least up to about 40 s. The velocity maximum  $V_{\max}$  of Rayleigh waves occurs only seldom at longer periods. Examples of suitable velocity broadband responses are given in the right-hand panel of Figure 10 with STS2 and STS1-VBB. These frequency ranges are also covered by several more recent broadband systems from the UK, USA and China (see DS 5.1 in this Manual).

## 5.2 The influence of the measurement time-window on mb and mB estimates

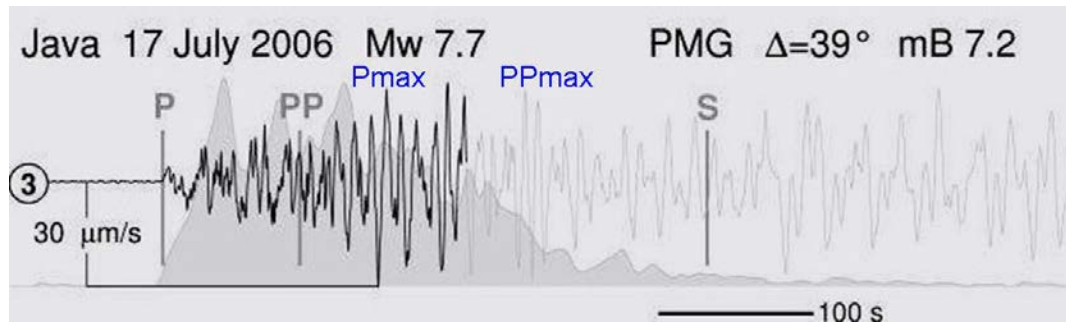
In the above definition of the mb and mB standards it is stated that these body-wave magnitudes should be “...calculated from the maximum trace-amplitude in the **entire P-phase train** (time spanned by P, pP, sP, and possibly PcP and their codas, and ending preferably before PP).

Gutenberg has derived calibration functions for P and PP waves but not for PcP and depth phases. PcP amplitudes are generally smaller than P amplitudes, especially broadband ones

(see, e.g., record examples for event 6 in DS 11.2). So there is no magnitude bias likely from measuring PcP instead of P.

Depth phases were considered by Gutenberg as part of the primary phases with commonly smaller (and then for magnitude estimates irrelevant) amplitudes because of reflection and/or conversion losses as well as extra attenuation along the additional segment of travel. For shallow earthquakes a clear discrimination between superimposed direct and depth phases is difficult anyway and usually beyond the possibilities of routine seismogram analysis. Yet, depth phases may become even larger than their primary phases, as confirmed by both observations (e.g., Figs. 2.62-2.65 in Chapter 2, and Fig. 11.16 ??? in Chapter 11) and synthetic modeling (e.g., Langston and Helmberger, 1975).. The great variability of amplitudes of both P and its depth phases is a source-mechanism dependent function of take-off angle and azimuth and therefore varies with distance and azimuth of the recording station with respect to the source radiation pattern. Procedures such as that of Choy and Boatwright (1995) for calculating released seismic energy  $E_s$  and energy magnitude  $M_e$  assume that the P-wave train consists of direct P plus depth phases and they apply source-mechanism dependent corrections. Since the take-off angles of direct P, pP and sP differ strongly, the inclusion of the depth phases into the mb and mB estimates has an averaging stabilizing effect which reduces possible bias due to strongly variable P-wave radiation when only a limited number of observations with insufficient azimuthal coverage is available.

Also broadband PP amplitudes are usually smaller than those of direct P and especially depleted in higher frequencies because of their longer overall ray path with additional segments through the stronger attenuating upper mantle (see, e.g., record examples of events No. 6 and 7 in DS 11.2). Moreover, onsets of PP are usually well enough separated from P, easy to identify by means of their differential travel-time difference (see differential travel-time curves in Figure 4 of EX 11.2), have their own calibration values (see DS 3.1 in this Manual) and thus are not a biasing factor for P-wave based mb or mB estimates. However, the rupture duration of great earthquakes may be so long that significant amounts of P-wave energy are still arriving at the theoretically expected first arrival time of PP. Should then the measurement time-window **preferably end before PP**, as recommended by the standards? Not necessarily, if good reason speaks against it and can be proven by measurement. Figure 15 gives an example for the application of a simple technique, which allows to constrain the appropriate time window in which to search for the maximum P amplitude. It measures the duration  $d$  of high-frequency radiation, which is mainly generated at the progressing rupture front. Thus  $d$  is also a rough estimate of the rupture duration  $T_R$  but tends to be longer than  $T_R$  because of the deliberate inclusion of the depth phases of P and their codas. The technique (for details see (Bormann and Saul, 2008) resembles those of Hara (2007a and b), Lomax et al. (2007), Lomax and Michelini (2009). It computes the envelopes of individual, velocity-proportional P-wave records filtered in a frequency band between 1 and 3 Hz. From the average of all envelopes, aligned by P onset, the time lag is determined at which the amplitude falls below 40% (energy-wise to 16%) of its maximum, thus giving a robust estimate of the approximate duration  $d$  of the whole P-wave train. It allows to constrain the time window for mb and mB measurements, even in the presence of later phases like PP, which contribute to the high-frequency signal much less than direct P and its depth phases. This is especially useful in automated setups, but simple visual time window selection on the SP-filtered record by an experienced analyst before measuring  $A_{max}$  for mb calculation yields practically the same result.



**Figure 15** Vertical-component broadband record of the Mw7.7 Java earthquake of 2006 at station PMG. The grey shaded area is the envelope of the positive amplitudes of the velocity-proportional high-frequency (1-3 Hz) filtered P-wave record (plotted at larger scale). Vmax of P is searched within the time-window from the P onset and the time at which the envelope amplitude falls below 40% of the maximum envelope amplitude. This part of the record trace is shown in black. The Pmax, to be measured for mB determination occurs some 140 s after the first P onset, with a period around 10 s, some 50 s later than the theoretically calculated first onset of PP. In contrast, according to the SP envelope, Pmax in the SP record, to be measured for mb, appears already 40 s after the P onset. (Amended cut-out from Figure 3 in Bormann and Saul (2008), *Seism. Res. Lett.*, **79**(5), p. 700, © Seismological Society of America.)

Kanamori (2006), investigating the seismic energy release of the great Mw9.3 2004 Sumatra earthquake, also extended the integration window over the whole P-wave train into the PP range with an estimated bias in  $E_s$  that would correspond to less than 0.1 m.u. in Me.

### 5.3 The influence of the calibration function on the magnitude estimates

The *Magnitude WG* had been entrusted solely with proposing unique measurement standards but not with the development or proposal of new calibration functions. For most magnitude types, the *WG* adopted a strategy of identifying classical, already widely used, attenuation functions as the standard function, with expectation that these functions would be improved upon in the future. The *WG* endorsed the commonly used practice of using the classical attenuation functions as reference base lines for scaling improved global and regionally specific attenuation functions. Some of such alternative functions exist already but in most cases no agreement has so far been reached for their universal acceptance and use. Below we briefly comment on the current and some of the potential alternative future calibration relationships in order to understand their main differences and possible effects in the case of change.

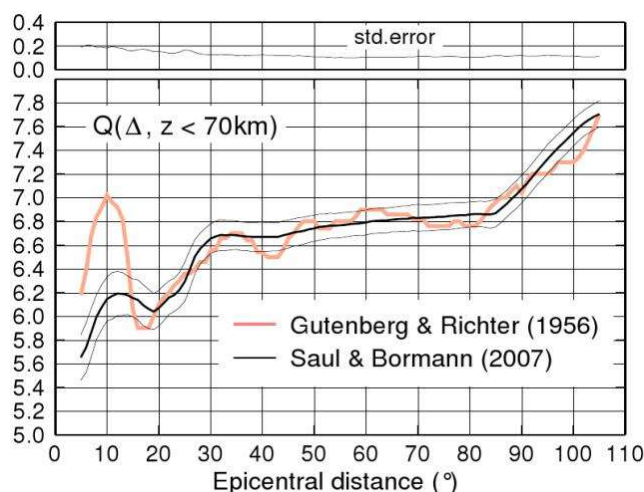
#### 5.3.1 ML calibration function

The new IASPEI ML standard replaces the old Richter calibration function by the expanded Hutton and Boore (1987) relationship, modified for the Uhrhammer and Collins (1990) recalibration of the Wood Anderson seismograph. The difference from the original Richter formula and its tabulated calibration values has been plotted in Figure 2 (with comments), and the benefits of scaling other regional calibration functions to the H-B formula better at shorter distances than 100 km have already been discussed in section 4.1 of this IS (see also Fig. 3.30

in Chapter 3). Yet, ML differs from most of the other magnitudes in that most ML values are already calculated with region-specific formulas, and they are appropriately designated by unique formulas. The most important principle to establish with ML is that the standard amplitudes are measured through the WA filter and that these amplitudes should be reported in bulletins.

### 5.3.2 mB\_BB and mb calibration functions

Both standard magnitudes are still calibrated with the original revised Gutenberg-Richter (1956a) Q-values for vertical component medium-period P waves (QPZ). The original diagrams and tabulated values are given in DS 3.1, the tabulated values resulting from the USGS scan of the QPZ diagram are presented in Table 2 of this IS for the distance range 20° to 100° as a function of hypocenter depth. The original QPZ( $\Delta$ , h) values were given for the distance range 5° to 110°. However, the values in the beginning of the Earth's core shadow zone of P beyond 100° and in the strongly upper mantle affected distance range between 5° and 20° are considered not yet reliable enough for global application. These problems have already been pointed out much earlier by Evernden (1967) and Booth et al. (1974). A detailed discussion of this issue goes far beyond the scope of this IS and still requires many dedicated investigations which will greatly benefit from the global introduction into observatory practice of the proposed measurement standards. Figure 16, presented in an AGU poster by Saul and Bormann (2007), has been derived from global velocity broadband IVmB\_BB amplitude measurements. It may just serve as an illustration of what one can expect, namely a much smoother distance dependence of QPZ in general and major revisions, most likely regionally variable, for distances below 20°. That is, why the current global standards do not yet propose to calculate mb and mB\_BB at distances below 20°. We encourage, however, to measure and report to international data centers related amplitude values as Iamb and IV\_mB\_BB also if they have been measured at distances less than 20° in order to collect masses of standardized data from all seismic regions for investigating such regional variations in amplitude-distance relationships.



**Figure 16** Comparison between the original Gutenberg-Richter (1956a) Q values for vertical component P waves from earthquakes at depth < 70 km with preliminary data based on broadband velocity P-wave amplitude measurements by Saul and Bormann (2007). The Saul and Bormann curve is associated with curves representing +/- the standard error, which is also represented in the chart at the top of the figure.

Veith and Clawson (1972) developed an alternative vertical-component short-period body-wave scale  $P(\Delta, h)$ , which was scaled to the level of the Gutenberg-Richter  $Q(\Delta)_{PZ}$  for surface focus.  $P(\Delta, h)$  is based on WWSSN-SP records from strong explosions. The calibration curves for deeper sources (see Figure 2 of DS 3.1) were calculated using an attenuation model developed by Veith and Clawson (1972). There are large differences between the Gutenberg-Richter and the Veith-Clawson calibration function for deep earthquakes, up to 0.6 m.u. The Veith and Clawson (1972) body-wave scale has been adopted by the CTBTO/IDC, and differences between the Veith and Clawson  $P(\Delta, h)$  values and the Gutenberg and Richter  $Q(\Delta, h)$  values are, for intermediate-depth and deep-focus earthquakes, a significant component of the discrepancy between the  $mb(\text{NEIC})$  and  $mb(\text{IDC})$  discussed in section 1 (Murphy and Barker, 2003).

In general, there is a need to develop improved global and region-specific calibration functions for both  $mb$  and  $mB$ , or to confirm some of the various calibration functions that have already been proposed. In addition to the previously discussed global calibration functions of Gutenberg and Richter (1956a) and Veith and Clawson (1972), Murphy and Barker (2003) have tabulated proposed global calibration functions from  $0^\circ$  to  $180^\circ$ , and Christokov et al. (1983), Lilwall (1987), and Rezapour (2003) have tabulated proposed global calibration functions for distances beyond  $20^\circ$ . The effect of frequency dependence on the attenuation of P-waves must also be considered. The ability to confidently use  $mB$  calculations from the near regional distance range would allow seismologists to exploit still better the superior potential of  $mB$  for yielding less saturating magnitude estimates of strong earthquakes in real time for tsunami early warning and emergency assistance purposes. Greater confidence in the short-period magnitudes assigned to deep earthquakes might shed light on the source physics of these earthquakes. The future work will strongly benefit from world-wide adherence to the newly recommended IASPEI measurement standards, because they will significantly reduce data scatter due to non-standard procedures, thus bringing out more clearly biases in currently assumed empirical and theoretical Earth models.

### 5.3.3 $Ms_{BB}$ and $Ms_{20}$ calibration function

The empirical stability and global representativeness of  $Ms_{BB}$  measurements (see section 4.5) notwithstanding, there are a number of factors that would be expected to produce biased station  $Ms_{BB}$  in some seismotectonic settings. “Biased results”, in the sense used here, would typically be  $Ms_{BB}$  values at regional distances that are significantly different than the values at teleseismic distances. Marshall and Basham (1972) summarize a number of such factors. For example, Rayleigh wave amplitudes of periods near 10s, from which  $Ms_{BB}$  might be measured at  $\Delta \sim 15^\circ$ , are expected to be sensitive to differences in velocity structure and attenuation properties of propagation paths, particularly to the difference between a continental path and an oceanic path, but also to the differences between some types of continental path. Other potential sources of bias are the difference between the source spectrum at long and short periods, which will itself be different for earthquakes of different  $M_w$ , and the sensitivity of the Rayleigh-wave spectrum to focal depth and focal-mechanism (Tsai and Aki, 1970). A mass collection of standardized  $Ms_{BB}$  and related amplitude and period values in the years to come will provide a good basis for investigations on possible  $Ms_{BB}$  biases associated with particular seismotectonic environments.

More is already known about the systematic distance-dependent biases of  $Ms_{20}$  when scaled with the IASPEI recommended standard  $Ms$  formula (10) according to Vaněk et al. (1962) (von Seggern, 1977; Herak and Herak, 1993; Rezapour and Pearce, 1998). As noted

previously, the IDC of the CTBTO calculates already its  $M_s$  with formula (18) of Rezapour and Pearce (1998) (see DS 3.1 and Chapter 17 of this Manual) and is successfully calculating  $M_s$  at near-regional and regional distances, in addition to teleseismic distances. A significant component of the distance-dependent bias of formula (10), when used for scaling amplitude measured around 20 s only, is likely to arise from the fact that constants in the adopted  $M_{s\_20}$  attenuation relation, equation (3)/equation (10), were determined with data over a broader range of periods with average  $T$  increasing significantly as a function of epicentral distance (see Fig. 8 in Bormann et al., 2009).

Bormann et al. (2009) showed that the difference of relation (10) to relations given by Herak and Herak (1993) or Rezapour and Pearce (1998) reduces to less than 0.1 m.u. between  $50^\circ$  and  $170^\circ$  when using instead of the calibration term in the IASPEI  $M_s$  formula the respective tabulated calibration values in Table 4 of DS 3.1 of this Manual. Using tabulated calibration values instead of an approximated formula was proposed already earlier by Vaněk (1995) with reference to the publication about the homogeneous magnitude system (HMS) studies for Eurasia (Christoskov et al., 1991). Yet between  $1^\circ$  and  $130^\circ$  tabulated and formula-derived Prague-Moscow calibration values agree within 0.02 m.u. The global introduction of a revised  $M_{s\_20}$  specific calibration scale may be reconsidered again after several years of standardized  $M_{s\_20}$  measurements are available.

### 5.3.4 How amplitude, period and measurement time should be measured

The first sentence in section 3.2, which outlines the IASPEI recommended measurement standards, reads: “The amplitudes used in the magnitude formulas below are to be measured as one-half the maximum peak-to-adjacent-trough (sometimes called “peak-to-peak”) deflection of the seismogram trace.” This recommendation is based on common practice with analog records. Analog records had a very limited dynamic range of about 30-40 db. Most trace amplitudes were measured with low signal-to-noise ratio in the millimeter range, sometimes only 2- to 3-times the thickness of the record trace and with the zero line difficult to fix. By measuring one-half peak-to-trough, relative reading errors could be reduced. This may be the reason why Richter, who initially (Richter, 1935) measured zero-to-peak amplitudes changed later (Gutenberg and Richter, 1956b) to one-half peak-to-trough measurement. This practice also reduces measurement errors when short-period signals are riding over long-period noise or the surface-wave train of a previous earthquake, making it difficult to fix the zero reference line. Moreover, the majority of record signals, especially in narrowband records of the WWSSN-type, and dispersive surface-wave trains anyway, appeared rather symmetric. Empirically, both one-half peak-to-trough and zero-to-peak reading practices yield in many circumstances the same magnitude values to within 0.1 m.u. (Rezapour and Rezaei, 2011). Thus it became the dominant practice in older times and continued to be also the easiest and rather unambiguous way of amplitude measurement in modern interactive seismogram analysis programs, because an experienced analyst will always be able to recognize most easily and pick correctly the largest peak and adjacent trough. Accordingly, almost all reading examples presented in the 1979 Willmore Manual of Observatory Practice as well as in the first edition of the NMSOP favor the (P-to-T)/2 approach of amplitude.

However, the problem is more complex. In the first edition of the NMSOP there is also an example given for amplitude measurement on a strongly asymmetric P-wave onset, as they often appear on broadband displacement records. The more broadband a displacement record is, the more asymmetric are the records of body-wave onsets (see the two lowermost record



traces in Fig. 4.13 and Fig. 4.15 of Chapter 4). Many classical medium- to long-period analog seismographs, widely used for measuring mB and Ms in the past, such as those of type Wiechert, Mainka, Galitzin and Kirnos, were more or less broadband displacement recorders. NMSOP then recommends to measure the largest half-swing amplitude from the zero line.

In this context one should be aware that an ideal seismic rupture does not produce an harmonic oscillating ground displacement. Rather, when the rupture can be approximated in a time-displacement diagram by a ramp function in the *near-field* range then this is observed in the *far-field* as a more or less bell-shaped and single-sided displacement pulse (see Fig. 2.4 in Chapter 2), which, if directivity is present, may have asymmetric rise and decay times (e.g., the Brune model displacement pulse in Fig. 4.17 of Chapter 4) Only its time-derivative, i.e., the far-field ground-motion velocity, is a two-sided oscillation with comparable amplitudes. Also narrow-band records in general as well as acceleration and surface-wave records show more symmetric oscillations. The differences in the symmetry of oscillations are very obvious, when comparing in Fig. 4.18 of Chapter 4 the unfiltered velocity broadband and the narrow-band short-period WWSSN-SP and long-period SRO\_LP records with the Kirnos and WWSSN-LP displacement records.

Therefore, many algorithms for automatic record analysis prefer zero-to-peak (or trough) measurement, e.g., the current version of SeisComp3 (Hanka et al., 2010) and used for the mB study by Bormann and Saul (2008). Zero-to-peak (trough) measurement can be more easily and unambiguously implemented, even a moving zero-line can most easily be determined by a moving long-term average and de-trending algorithm. Measuring the maximum amplitude between two zero crossings avoids the risk of getting trapped in a secondary minimum or maximum when looking for the largest adjacent trough (or peak).

Clearly, although the *Magnitude WG* has specified a procedure for measuring amplitudes, there is not consensus in the seismological community on the question of optimum procedure for amplitude measurement. Comparative measurements of amplitudes for standard mb, mB\_BB, Ms\_20 and Ms\_BB using both the peak-to-trough/2 and the zero-to-peak (or trough) approach are currently still under way and will be reported soon in a related section of the revised Chapter 3 of this Manual.

The new IASPEI recommended procedure for determining periods differs from the procedure that was commonly used with analog instruments and that is recommended in the 1979 Willmore Manual of Observatory Practice as well as in the first edition of the NMSOP, namely, that period should be read by measuring the time difference between the two neighboring maximum peaks or troughs. The newly recommended procedure -- "...periods are to be measured as twice the time-intervals separating the peak and adjacent-trough from which the amplitudes are measured" -- would have been impracticable using large amplitude analog records with fixed low time resolution and the difficulty to project precisely the time position of the peak down to the level where the time difference to the trough had to be measured. But most modern interactive analysis programs can now perform this easily and give the time of the zero-crossing between the peak and adjacent-trough as the amplitude-phase arrival-times. A modern zero-to-peak amplitude measuring algorithm would use neither the procedure recommended in the earlier manuals nor the current IASPEI recommended procedure: a zero-to-peak algorithm would measure the double time difference between the zero crossings relating to the measured maximum amplitude as period and give the time of the peak position itself as the measurement time. The fully automatic procedure of amplitude and period measurement at the IDC of the CTBTO determines even the period from the three half



periods between which the maximum amplitude has been measured (see Fig. 17.9 in Chapter 17 of this Manual).

For all the IASPEI magnitudes except *mb*, the value of the magnitude does not actually depend on the measured period. In fact, also *mb* could be measured directly via *V<sub>max</sub>* with likely comparable or even better precision than the standard procedure (currently under investigation) after convolving the broadband velocity record directly with the WWSSN-SP response. The cataloging of *T* is, however, important because *T* provides useful supplemental information that can help assess the reliability of a station magnitude measurement and because *T* for some magnitude-types (*mB\_BB* or *Ms\_BB*) contains information about the seismic source that is independent of the information provided by amplitude. Alternative strategies for measuring *T* (e.g., Boore, 1986; Kanamori, 2005) may provide estimates of *T* that are more informative for some purposes (e.g. engineering seismological) than those provided by the IASPEI procedure.

## 6 Summary

The information sheet outlines both the aim and the rationale of the IASPEI (2005 and 2011) recommended measurement procedures for widely used magnitude scales, and emphasizes the need for a unambiguous unique standardized nomenclature for reported and published amplitudes and calculated magnitudes. The new IASPEI magnitude formulas are presented and the essentials of the procedures and parameter ranges outlined, explained in their historical development, critically analyzed with respect to alternative procedures, their significance, range of tolerance and acceptable modifications. The advantages of the (in western countries still uncommon) broadband body and surface-wave magnitudes *mB* and *Ms\_BB* as compared to the bandlimited magnitudes *mb* and *Ms\_20* are highlighted.

Also stressed is the need for overlapping comparative measurements at seismic stations and analysis centers in conjunction with the implementation of the new standard procedures. Previous and the newly recommended standard procedures should be applied to the broadband waveform data from an IASPEI authorized set of reference events and their results be analyzed statistically. As an example of such a comparison the results of *Ms\_20* determinations on simulated WWSSN-LP and SRO-LP records, respectively, are presented, proving that they yield identical results. Guidelines for carrying out and documenting the results of such a comparative test are presented in IS 3.4.

Finally, future research into improving magnitude calibration functions, especially in the local and regional distance ranges, is strongly encouraged. It is hoped that such investigations will greatly benefit from the expected forthcoming masses of new data based on the agreed IASPEI measurement standards which are hoped to significantly reduce data scatter due to inconsistent or incompatible procedures of magnitude measurements.

## Acknowledgments

We are indebted to Members of the IASPEI Working Group on Magnitude Measurements and others who have essentially contributed to the discussions, consensus and results presented above, or made Figures available for this information sheet, in particular Siegfried Wendt, Joachim Saul and Klaus Stammer. We thank George Choy and Morgan Moschetti for

helpful reviews. Any use of trade, firm, or product names is for descriptive purposes only and does not imply endorsement by the U.S. Government.

## References

- Abe, K. (1981). Magnitudes of large shallow earthquakes from 1904 to 1980. *Phys. Earth Planet. Interiors*, **27**, 72-92.
- Abe, K. (1984). Complements to “Magnitudes of large shallow earthquakes from 1904 to 1980”, *Phys. Earth Planet. Interiors*, **34**, 17-23.
- Abe, K., and Kanamori, H. (1980). Magnitudes of great shallow earthquakes from 1953 to 1977. *Tectonophysics*, **62**, 191-203.
- Aki, K. (1967). Scaling law of seismic spectrum. *J. Geophys. Res.*, **72**, 1217-1231.
- Boore, D. M. (1986). Short-period *P*- and *S*-wave radiation from large earthquakes: implications for spectral scaling relations. *Bull. Seism. Soc. Am.*, **76**, 43-64.
- Booth, D. C., Marshall, P. D., and Young, J. B. (1974). Long and short period *P*-wave amplitudes from earthquakes in the range 0°-114°. *Geophys. Jour. R. Astr. Soc.*, **39**, 523-537.
- Bonner, J. L., Russell, D. R., Harkrider, D. G., Reiter, D. T., and Herrmann R. B. (2006). Development of a time-domain, variable-period surface-wave magnitude measurement procedure for application at regional and teleseismic distances, part II: Application and Ms-mb performance, *Bull. Seism. Soc. Am.* **96**, 2, 678-696, doi: 10.1785/0120050056.
- Bormann, P. (1969): Some features of seismic body-wave onsets and their consideration in improving source location (in German). In *Veröff. d. Inst. für Geodynamik Jena, Reihe A, Heft 14*, Akademie-Verlag, Berlin, S. 21-32.
- Bormann, P. (2002a). Magnitude of seismic events. Section 3.2. In P. Bormann (Editor), *New Manual of Seismological Observatory Practice (NMSOP)*, GeoForschungsZentrum Potsdam, Vol. 1, Chapter 3, 16-49.
- Bormann, P. (2002b). DS 3.1, Magnitude calibration functions and complementary data. In P. Bormann (Editor), *New Manual of Seismological Observatory Practice (NMSOP)*, GeoForschungsZentrum Potsdam, Vol. 2, 8 pp.
- Bormann, P., and Khalturin, V. I. (1975). Relations between different kinds of magnitude determinations and their regional variations. *Proceed. XIVth General Assembly of the European Seismological Commission, Trieste, 16-22 September 1974*. Nationalkomitee Geod. Geophys., AdW der DDR, Berlin, 27-39.
- Bormann, P., and Di Giacomo, D. (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bormann, P., and Saul, J. (2008). The new IASPEI standard broadband magnitude  $m_B$ , *Seismol. Res. Lett.* **79** (5), 699-706.
- Bormann, P., and Saul, J. (2009a). A fast, non-saturating magnitude estimator for great earthquakes, *Seism. Res. Lett.* **80** (5), 808-816.
- Bormann, P., and Wylegalla, K. (1975). Investigation of the correlation relationships between various kinds of magnitude determination at station Moxa depending on the type of

- instrument and on the source area (in German). *Public. Inst. Geophys. Polish Acad. Sci.*, **93**, 160-175.
- Bormann, P., Liu, R., Ren, X., Gutdeutsch, R., Kaiser, D., Castellaro, S. (2007). Chinese national network magnitudes, their relation to NEIC magnitudes, and recommendations for new IASPEI magnitude standards. *Bull. Seism. Soc. Am.*, **97**, 114-127.
- Bormann, P., Liu, R., Xu, Z., Ren, K., Zhang, L., Wendt S. (2009). First application of the new IASPEI teleseismic magnitude standards to data of the China National Seismographic Network, *Bull. Seism. Soc. Am.* **99** (3), 1868-1891; doi: 10.1785/0120080010
- Choy, G. L., and Boatwright J. (1995). Global patterns of radiated seismic energy and apparent stress, *J. Geophys. Res.* **100**, 18,205-18,228.
- Christoskov, L., Kondorskaya, N. V., and Vaněk J. (1991). Homogeneous magnitude system with unified level for usage in seismological practice, *Studia geoph. Et geod.* **35**, 221-233.
- Christoskov, L., Kondorskaya, N. V., and Vanek, J. (1983). Homogeneous magnitude system of the Eurasian continent: S and L waves. *World Data Center A for Solid Earth*, Boulder Report SE-34.
- Ekström, G., and Dziewonski, A. M. (1988). Evidence of bias in estimations of earthquake size. *Nature*, **332**, 319-323.
- Evernden, J. F. (1967). Magnitude determination at regional and near-regional distances in the United States. *Bull. Seism. Soc. Am.*, **57**, 591-639.
- Geller, R.J. (1976). Scaling relations for earthquake source parameters and magnitudes. *Bull. Seism. Soc. Am.*, **66**, 1501-1523.
- Granville, J. P., Kim, W.-Y., and Richards, P. G. (2002). An assessment of seismic body-wave magnitudes published by the Prototype International Data Centre. *Seism. Res. Lett.*, **73**(6), 893-906.
- Granville, J. P., Richards, P. G., Kim, W.-Y., and Sykes, L. R. (2005). Understanding the differences between three teleseismic  $m_b$  scales. *Bull. Seism. Soc. Am.*, **95**(5), 1809-1824.
- Gutenberg, B. (1945a). Amplitudes of surface waves and magnitudes of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 3-12.
- Gutenberg, B. (1945b). Amplitudes of P, PP, and S and magnitude of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 57-69.
- Gutenberg, B. (1945c). Magnitude determination of deep-focus earthquakes. *Bull. Seism. Soc. Am.*, **35**, 117-130.
- Gutenberg, B., and Richter, C. F. (1954). *Seismicity of the Earth and associated Phenomena*. 2<sup>nd</sup> edition, Princeton University Press, 310 pp.
- Gutenberg, B., and Richter, C. F. (1956a). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Gutenberg, B. and Richter, C.F. (1956b). Earthquake magnitude, intensity, energy, and acceleration (second paper), *Bull. Seism. Soc. Am.*, **46**, p. 105-145.
- Hanka, W., Saul, J., Weber, B., Becker, J., Harjadi, P., Fauzi, and GITEWS Seismology Group (2010). Real-time earthquake monitoring for tsunami warning in the Indian Ocean and beyond. *Nat. Hazards Earth. Syst. Sci.*, **10**, 2611-2622; doi: 105194/nhess-10-2611-2010.
- Hara, T. (2007a). Measurement of the duration of high-frequency energy radiation and its application to determination of the magnitudes of large shallow earthquakes. *Earth Planets Space* **59**, 227-231.

- Hara, T. (2007b). Magnitude determination using duration of high frequency radiation and displacement amplitude: application to tsunami earthquakes. *Earth. Planet Space* **59**, 561-565.
- Herak, M., and Herak, D. (1993). Distance dependence of  $M_S$  and calibrating function for 20 s Rayleigh waves. *Bull. Seism. Soc. Am.*, **83**, 6, 1881-1892.
- Herak, M., Panza, G., and Costa, G. (2001). Theoretical and observed depth corrections for  $M_s$ . *Pure Appl. Geophys.*, **158**, 1517-1530.
- Houston, H., and Kanamori, H. (1986). Source spectra of great earthquakes: teleseismic constraints on rupture process and strong motion. *Bull. Seism. Soc. Am.*, **76**, 19-42.
- Hunter, R. N. (1972). Use of LPZ for magnitude. In: *NOAA Technical Report ERL 236-ESL21*, J. Taggart, Editor, U.S. Dept. Commerce, Boulder, Colorado.
- Hutton, L. K., and Boore, D. M. (1987). The  $M_L$  scale in southern California. *Bull. Seism. Soc. Am.*, **77**, 2074-2094.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf).
- IASPEI (2011) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_of\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_of_WG_recommendations_20130327.pdf)
- IASPEI (2013). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)
- Kanamori, H. (1983). Magnitude scale and quantification of earthquakes, *Tectonophysics* **93**, 185-199.
- Kanamori, H. (2005). Real-time seismology and earthquake damage mitigation. *Annu. Rev. Planet. Sci.*, **33**, 195-214.
- Kanamori, H. (2006). The radiated energy of the 2004 Sumatra-Andaman earthquake, in: Abercrombie R, McGarr A, Kanamori H (eds): Radiated energy and the physics of earthquake faulting, *AGU Geophys Monogr Ser* **170**, 59-68.
- Kanamori, H., and Anderson, D. L. (1975). Theoretical basis of some empirical relations in seismology. *Bull. Seism. Soc. Am.*, **65**, 1073-1095.
- Karnik, V., N. V. Kondorskaya, Ju. Riznitchenko, E. F. Savarensky, S. L. Sobolev, N. V. Shebalin, J. Vanek, and A. Zatopek (1962). Standardization of the earthquake magnitude scale. *Studia geoph. et geod.* **6**, 41-48.
- Kim, W.-Y. (1998). The  $M_L$  scale in Eastern North America. *Bull. Seism. Soc. Am.*, **88**, 4, 935-951.
- Kondorskaya, N. V., Z. I. Aranovich, O. N. Solov'eva, and N. V. Shebalin (Eds.) (1981). Instrukciya o poryadke proizvodstva i obrabotki nabludenii na seismicheskikh stanciyach ESSN *Inst. Fiziki Zemli Akad. Nauk SSSR*, Moscow, 272 pp.
- Langston, Ch. A., and Helmberger, D. V. (1975). A procedure for modeling shallow dislocation sources. *Geophys. J. R. astr. Soc.*, **42**, 117-130.
- Lienkaemper, J. J. (1984). Comparison of two surface-wave magnitude scales:  $M$  of Gutenberg and Richter (1954) and  $M_s$  of "Preliminary Determination of Epicenters". *Bull. Seism. Soc. Am.*, **74**(6), 2357-2378.

- Lillwall, R. C. (1987). Empirical amplitude-distance/depth curves for short-period P waves in the distance range 20° to 180°. *AWRE Report No. O 30/86*, HMSO, London.
- Lomax, A., and Michelini, A. (2009).  $M_{wpd}$ : A duration-amplitude procedure for rapid determination of earthquake magnitude and tsunamigenic potential from P waveforms. *Geophys. J. Int.*, **176**, 200-214; doi: 10.1111/j.1365-246X.2008.03974.x .
- Lomax, A., Michelini, A., and Piatanesi A. (2007). An energy-duration procedure for rapid and accurate determination of earthquake magnitude and tsunamigenic potential. *Geophys. J. Int.* **170**, 1195-1209; DOI: 10.1111/j.1365-246X.2007.03469.x.
- Marshall, P. D., and Basham, P. W. (1972). Discrimination between earthquakes and underground explosions employing an improved  $M_S$  scale. *Geophys. J. R. Astr. Soc.*, **28**, 431-458.
- Murphy, J.R., and Barker, B.W. (2003). Revised distance and depth corrections for use in estimation of short-period P-wave magnitudes. *Bull. Seism. Soc. Am.*, **93**, 1746-1764
- Nuttli, O.W. (1973). Seismic wave attenuation and magnitude relations for eastern North America. *J. Geophys. Res.*, **78**, 876-885.
- Nuttli, O.W. (1986). Yield estimates of Nevada Test Site explosions obtained from seismic Lg waves. *J. Geophys. Res.*, **91**, 2137-2151.
- Patton, H. J. (2001). Regional magnitude scaling, transportability, and  $M_S:m_b$  discrimination at small magnitudes, *Pure and Applied Geophysics*, **158**, 1951-2015.
- Rezapour, M. (2003). Empirical global depth-distance correction terms for mb determination based on seismic moment. *Bull. Seism. Soc. Am.*, **93**, 172-189.
- Rezapour, M., and Pearce, R. G. (1998). Bias in surface-wave magnitude  $M_S$  due to inadequate distance corrections. *Bull. Seism. Soc. Am.*, **88**, 1, 43-61.
- Rezapour, M., and Rezaei, R. (2011). Empirical distance attenuation and the local magnitude scale for northwest Iran, *Bull. Seism. Soc. Am.*, **101**, 3020-3031.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bull. Seism. Soc. Am.*, **25**, 1-32.
- Richter, C. F. (1958). Elementary seismology. *W. H. Freeman and Company*, San Francisco and London, viii + 768 pp.
- Saul, J., and Bormann, P. (2007). Rapid estimation of earthquake size using the broadband P-wave magnitude mB. *Eos, Transactions, American Geophysical Union, AGU 2007 fall meeting*, **88**, no. 52, Suppl., Abstract S53A-1035.
- Soloviev, S. L. (1955). Classification of earthquakes in order of energy (in Russian). *Trudy Geofiz. Inst. AN SSSR*, **30**, 157, 3-31.
- Stevens, J.L., and McLaughlin, K.L. (2001). Optimization of Surface Wave Identification and Measurement, *Pure appl. Geophys.*, **158**, 1547-1582.
- Tsai, Y.-B., and Aki, K. (1970). Precise focal depth determination from amplitude spectra of surface waves. *J. Geophys. Res.*, **75**, 5729-5743.
- Uhrhammer, R. A., and Collins, E. R. (1990). Synthesis of Wood-Anderson seismograms from broadband digital records. *Bull. Seism. Soc. Am.*, **80**, 702-716.
- Uhrhammer, R. A., Hellweg, M., Hutton, K., Lombard, P., Walters, A. W., Hauksson, E., and Oppenheimer, D. (2011). California Integrated Seismic Network (CISN) local magnitude determination in California and vicinity. *Bull. Seism. Soc. Am.*, **101**, 2685-2693.

- Utsu, T. (2002). Relationships between magnitude scales. in Lee, W.H.K., Kanamori, H., Jennings, P.C., and Kisslinger, C., (eds.). International Handbook of Earthquake and Engineering Seismology, v. 81A, Chapter 44, p. 733 – 746.
- Vaněk, J. (1995). Comment on “distance dependence of  $M_s$  and calibrating function for 20 second Rayleigh waves, by M. Herak and D. Herak, *Bull. Seism. Soc. Am.*, **85**, 961.
- Vaněk, J., Zapotek, A., Karnik, V., Kondorskaya, N.V., Riznichenko, Yu.V., Savarensky, E.F., Solov'yov, S.L., and Shebalin, N.V. (1962). Standardization of magnitude scales. *Izvestiya Akad. SSSR., Ser. Geofiz.*, **2**, 153-158 (108-111 in the English translation).
- Veith, K. F., and Clawson, G. E. (1972). Magnitude from short-period P-wave data. *Bull. Seism. Soc. Am.*, **62**, 435-452.
- von Seggern, D. (1977). Amplitude-distance relation for 20-second Rayleigh waves. *Bull. Seism. Soc. Am.*, **67**, 405-411.
- Willmore, P.L. (1979). Manual of Seismological Observatory Practice. World Data Center A for Solid Earth Geophysics, Report SE-20, September 1979, Boulder, Colorado, 165 pp.

| Topic   | Preliminary guidelines for using the IASPEI standard magnitude reference data set                                                                                                                                                                                                                                                                                                            |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, Dept. 2, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> <sup>1)</sup><br><b>Siegfried Wendt</b> , Geophysical Observatory Collm, University of Leipzig, D-04779 Wermsdorf, Germany, E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a> <sup>2)</sup> |
| Version | November 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_3.4                                                                                                                                                                                                                                                                                                                                               |

## 1. Introduction

1.) The authors are Co-chairman (P. B.) and member (S. W.) of the IASPEI Working Group on Magnitude Measurement and entrusted with promoting the widest possible implementation and comparative testing of recent IASPEI/CoSOI summary recommendations of standard procedures for widely used magnitude measurements (see **IS 3.3** and [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf))

2.) The current IASPEI standard magnitude reference data set allows to train yourself in the determination of the four **teleseismic magnitude standards** mb, mB\_BB, Ms\_20 and Ms\_BB. Read these related recommendations thoroughly before embarking on the exercise outlined below. The reference data set and guidelines will in a later version of this IS be amended by data and procedures also for the determination of the two **local-regional magnitude standards** ML and mb\_Lg.

3.) Comparing the results of the exercise with those derived by applying your current procedures of magnitude measurements to the same reference waveform data set will give you and your institution a clear understanding of the extent to which the new measurement standards yield results that agree or differ from your own results and to understand the reasons for possible differences.

4.) In this context there is an urgent need that stations and agencies which contribute parameter data to the International Seismological Centre (ISC) and the US National Earthquake Information Center (NEIC) provide an authoritative up-to-date **Documentation of station/agency magnitude procedures**. For your attention and easy response the related questions have been added as Annex 1 to this Information Sheet.

5.) **Note:** Should your agency not intend to introduce the recommended international standards in the future but to continue reporting your agency-specific *different* amplitude-magnitude data to the international data centres, then they need to be “flagged” by an agency-specific nomenclature. This permits data users to consult your documentation of station/agency magnitude procedures, to understand and account for related differences to the international standards and to decide about the suitability for your amplitude, period and magnitude data for their specific task.

6.) In the case that your institution uses some automatic procedures you should document how they work, e.g.:

- what type of original record data you have, their bandwidth and frequency range, whether displacement-proportional or velocity-proportional;
- which component and wave types are used for magnitude measurement;
- what type of filter (or none) is applied before amplitude, period and magnitude measurements are made;
- definition of time-window within which measurement is made, how the record amplitudes and periods are measured, which magnitude calibration functions are used, etc.).

7.) The current IS 3.4 provides reference data only for determining regional ( $M_s$ \_BB) and teleseismic magnitudes ( $m_b$ ,  $mB$ \_BB,  $M_s$ \_20 and  $M_s$ :BB) and related amplitudes, periods and measurement times. IS 3.4 will be amended in the course of 2012 by some more reference data and instructions for determining standard ML and  $m_b$ \_Lg. Also recommendations for using analysis software other than SEISAN may be added in future, provided that these analysis tools permit data filtering and presentation of measured amplitudes, periods, measurement times and magnitudes in a way that is compatible with the standard recommendations.

## 2. Data and software to be used and to be reported

In order to get an objective and reproducible comparison of IASPEI standard magnitudes with your institutional magnitudes both have to be measured on identical digital records.

With the exception of ML, all other magnitude standards are based on vertical-component recordings only. Therefore, we provide you in this first version of IS 3.4 only with single Z-component unfiltered STS-2 broadband velocity records, sampled at 20 Hz. These data will be complemented in 2012 by some horizontal component waveform files for ML determination. The records cover the whole wave train from the first P onset to the later part of the surface-wave group (respectively Sg-Lg group when ML and  $m_b$ \_Lg have to be determined).

The vertical component reference records have been made in a wide distance range from  $2^\circ$  to  $167^\circ$  from earthquakes with magnitudes between 4.7 to 8.5 (**Figure 1**). This allows you to check, whether differences between results of IASPEI standard procedures and your institutional procedures depend on distance and/or earthquake size.

**Annex 1** lists all reference event records, sorted according to epicentral distance, giving besides the record file name also the source region, hypocentral coordinates, epicentral distance of the recording station and the magnitude calibration value Q according to this distance and the source depth for the considered event.

Calibration values Q are given in the reference event list of **Annex 1** only for those magnitudes which can be determined for the given source depth and record distance. According to the recommended standards, no Q values are given for  $m_b$ ,  $mB$ \_BB and  $M_s$ \_20 at record distances less than  $20^\circ$  (then only for  $M_s$ \_BB), for  $m_b$  and  $mB$ \_BB at distances  $> 100^\circ$  and for  $M_s$ \_20 and  $M_s$ \_BB for earthquakes deeper than 60 km.

Further, the list contains columns for entering your measurements of the amplitude and periods used for calculation of the magnitude as well as of the time at which amplitude and period have been measured.



3

with the SEISAN manual, an introduction to exercises and to data from either the ftp site of the Institute of Geosciences of the University of Bergen, Norway (<http://www.uib.no/rg/geodyn/artikler/2010/02/software>) or directly via hyperlink **seisan** from the summary list of *Download Programs & Files* (see **Overview** of contents on the NMSOP-2 front page). Additional information can also be found in Havskov and Ottemöller (2010), and Havskov et al. (2011).

Different from the IASPEI Standard Format (ISF), all magnitude names in SEISAN - and in most other analysis software - still comprise only two letters. Yet, with the exception of Mw and duration/coda magnitude Mc, SEISAN calculates only IASPEI standard magnitudes. Yet, the calculation of mb\_Lg is not yet implemented although related amplitude measurements can be made. SEISAN mb is identical with the mb standard, provided that you measure the maximum amplitude in the whole P-wave group; the same applies to mB, which stands for standard mB\_BB. Further, SEISAN Ms means Ms\_20 and MS means Ms\_BB.

The measured amplitudes will be stored in the SEISAN event analysis files according to the unique IASPEI standard amplitude-phase names (see above and Table 1 below).

For measuring amplitudes and periods proceed as follow:

- 1.) SEISAN needs subdirectories for all years and months for which reference events are available. These subdirectories are stored in directories REA (S files) and WAV (waveform data), which are in turn stored in the basic directory (in general TEST\_). Type in the WOR (work) directory the command **makerea** and answer the questions. Answer GO AHEAD with Y (yes). New subdirectories are then created without changing already existing subdirectories.
- 2.) Load the magnitude reference event files into SEISAN with the commands **mulplt**, and **Regis**. Then you will be asked whether the event was local, regional or distant. You answer accordingly with either L, R or D and give your Operator Code (up to 4 letters).

The required vertical component broadband waveform files (and additional horizontal component waveform file for ML determination; to be specified in the later update of IS 3.4) can be downloaded via hyperlink from the folder **reference\_events** in the list of *Download Programs & Files* (see **Overview** on the NMSOP-2 front page). Therein you find the following tar-subfolders:

**ref\_evt\_00\_05.tar.gz** contains the waveform data files for the years 2000-2005;

**ref\_evt\_06\_09.tar.gz** those for 2006-2009 and

**ref\_evt\_94\_99.tar.gz** those for 1994-1999.

The reference events were recorded at station CLL in Germany. Note that two different digitizers were used, with the change occurring on 2 February 2007. The required response data files for SEISAN you find also in the folder **reference\_events**:

CLL\_BH\_Z.1993-04-13-000\_SEI and

CLL\_BH\_Z.2007-02-07-000\_SEI, respectively.

These files have to be loaded into the CAL directory.

- 3.) After this is done, load a specific event record file with **eev** yyyymmddhh. SEISAN then creates a line with running number, day, month, year, beginning of the time

window of the event record file and a question mark. After ? you may write the following **commands**:

- **po** for plot the seismogram (without further questions);
- **e** for read/edit the s file;
- **update** for saving the s file;
- **q** for finishing the evv.

In earlier versions SEISAN writes after the **po** command “aux station code not empty, use a station code ?”. Answer with any letter, but not with blank. In the new SEISAN version you can type ENTER. Then the requested complete seismogram will appear. Click on Menu and then - on the appearing command line - the “boxes” (buttons) for either WA (for Wood Anderson), mb, mB, Ms, or MS for which you wish to measure the amplitudes, periods and the amplitude-phase times. For ML, mb or Ms determination the respectively filtered record will then appear. For mB or MS you work on the instrument corrected unfiltered velocity broadband record.

- 4.) First choose the time resolution of the record so that the whole event record is present on the screen, including both the earlier body waves and the later surface waves with their usually (in the case of crustal earthquakes) much larger amplitudes and the longer and dispersive periods.
- 5.) For measuring mb or mB, increase the time resolutions by setting markers to cut the whole P-wave group out of the record. P usually has its largest amplitude between the first onsets of P and PP within the first minute (maybe including depth phases pP and/or sP). However, in the case of very great earthquakes ( $M_w > 8$ ) the rupture duration may last even for several (up to about 10) minutes and you may find the P-wave maximum only more than one min. after the first P-wave onset. In the case of ML and mb\_Lg measurement (which will be added in a later version of this exercise) increase the time resolution by cutting out the Sg-Lg group.
- 6.) Mark with the cursor the highest peak and deepest adjacent trough position and press at each marker position the letter **a**. Then the program will automatically measure the respective displacement or velocity amplitude, as well as the period and measurement time and write these parameter values into the **s(single)-file**. As measurement time the program takes the position of a vertical marker line which appears at the picked deepest trough position. This differs in time by about  $\pm$  a quarter of a period from the amplitude-phase arrival-times according to the IASPEI standard, which is measured as the time of the zero-crossing between the peak and adjacent-trough from which the amplitudes are measured. But this is close enough.
- 7.) This interactive measurement procedure is preferred for our exercise and reference event task. It will train both your own assessment of the different record appearance as well as of the specifics of the SEISAN program in relation to the recommended IASPEI standards for magnitude measurements. However, SEISAN allows also to measure automatically Amax (or Vmax) and T within the whole time window visible on the screen when you set the cursor somewhere near to the maximum of the whole local event train or of the teleseismic P-wave time train and press the capital letter **A**. This works fine, however, only in the case of good signal-to-noise ratio.
- 8.) For Ms (Ms\_20) and MS (Ms\_BB) proceed in the same way after selecting the Rayleigh (LR) surface-wave part of the record and then, with a higher time resolution, the time window around the LR maximum.

In Table 1 we show an example of the S file (with the original name 14-0926-00D.S200111) into which SEISAN has written the results of the amplitude (A) and period (T) measurements for the event 2001 1114 926 0.0D. Note that SEISAN stores the measured amplitudes with the correct IASPEI standard amplitude nomenclature in accordance with the IASPEI Standard Format (ISF), which is used at the international seismological data centers such as the ISC and NEIC. This amplitude nomenclature allows to associate the measured amplitudes unambiguously with the respective standard magnitudes. Therefore, standard amplitude measurements have in future to be reported to the international data centers in this format.

Table 2 will be added later, giving a similar data example for IAML, IAmb\_Lg, and IVMs\_BB (measured down to epicentral distance of 2°).

**Table 1** Plot of the amplitude-period data measured with SEISAN for 4 different telesismic IASPEI standard magnitudes from one reference event record.

```

2001 1114 926 0.0 D                                TES                                1
ACTION:UP 11-10-20 20:49 OP:sw STATUS: ID:20011114092600 I
011114_0926.bhz                                     6
STAT SP IPHASW D HRMM SECON CODA AMPLIT PERI AZIMU VELO AIN AR TRES W DIS CAZ7
CLL BZ IP 935 46.37
CLL BZ IVmB_BB 936 5.04 9581.4 4.69
CLL BZ IAmb 937 24.18 736.8 1.42
CLL BZ IVMs_BB 10 1 40.06 25.2e4 18.1
CLL BZ IAMS_20 10 1 45.48 70.7e4 18.6

```

- 9.) Insert for the analyzed events the measured displacement amplitudes (and periods, were required) as well as the values in **Annex 1** for QP and QL into the following formulas:

$$mb = \log_{10}(A/T) + QP - 3.0, \text{ respectively}$$

$$Ms_{20} = \log_{10}(A/T) + QL - 3.0,$$

and the related measured velocity amplitudes into the formulas:

$$mB_{BB} = \log_{10}(V_{\max}/2\pi) + QP - 3.0, \text{ respectively}$$

$$Ms_{BB} = \log_{10}(V_{\max}/2\pi) + QL - 3.0$$

and calculate the respective magnitudes with a spread sheet or pocket calculator. The -3.0 is required, since SEISAN and most other currently used analysis programs measure amplitudes in nm, whereas the classical calibration values and formulas for body and surface-waves were based on amplitude measurements in  $\mu\text{m}$ .

- 10.) Finally, insert the measured amplitudes, periods, measurement times and the calculated magnitudes into the respective empty (hyphen-less) column-lines of the reference event list (**Annex 1**). **txt**, **xls** and **doc** file versions of the tables can be

downloaded via the folder **reference\_events** from the listing of *Download Programs & Files* (see **Overview** on the NMSOP-2 front page)

**For clarification:** When you pick the onset times on records of several seismic stations, either local, regional or globally distributed ones, and locate first the seismic event yourself with SEISAN then the respective source parameters and epicentral distances are automatically calculated and stored in the respective SEISAN event file, thus enabling direct calculation of the magnitudes from the measured amplitudes and periods..

In our case, however, we have provided you only with one single component record. From this SEISAN can not calculate the source parameters and required Q values, unless you type yourself the event coordinates and station distance into the respective SEISAN event file. This, however, might be a potential source for typos and thus wrong results. Therefore, we request you to calculate for this comparative exercise the magnitudes yourself.

This detour, however, you will not encounter in the routine use of SEISAN for event location and magnitude determination using local, regional or global network records.

#### 4. How to measure and report your institutional magnitudes?

If you use at your station/network either an interactive or an automatic program for seismogram analysis and parameter determination please let us know the name and version (year) of the program.

If you determine at your station/network centre magnitudes with an interactive analysis program then please load the reference event files into your program, and the 4 required response parameters for the events files. Then proceed by measuring the amplitude and period as you usually do, calculate the related body-wave and/or surface-wave magnitude according to your respective formulas and report it/them to us in the last but one column of the list in **Annex 3**.

If you avail only of a fully automatic program for magnitude determination then you should check whether you can load the reference event files, together with the response parameters and pre-determined epicentral distance and source depth data into the program and get it run off-line for determining the magnitude value(s) according to the program's procedure. Report to us then these values with your common magnitude symbols in the last but one column (second from the right).

If your center operates a hybrid automatic program which also allows working in an interactive mode for a more detailed expert-controlled final data analysis (like e.g. SeisCompP) then load the reference event waveforms and the 4 required response parameters for the events files into the program and use the interactive program mode for measuring amplitudes and periods and determine the magnitudes according to your common station/agency procedure of data analysis. Report these results in the last but one column.

If your agency had determined for any of the reference events already earlier magnitudes - interactively or automatically - using your own station or network data and procedures, then please report these values and the magnitude type to us in the last column. Indicate which parameter data were determined interactively (int) or automatically (aut). Data provided in the last column will permit us to assess how the magnitudes based on your own data relate to

both the NEIC magnitudes and the single station standard reference magnitudes determined by Dr. S. Wendt for CLL station.

## 5. To whom you may address related questions and submit your station/agency analysis results and questionnaire answers?

In the case of any questions related to the SEISAN analysis software, the use of the reference data set, results or encountered problems you may address them to

Dr. Siegfried Wendt at the Observatory Collm, University of Leipzig  
E-mail: [obssw@hpwork30.collm.uni-leipzig.de](mailto:obssw@hpwork30.collm.uni-leipzig.de) or [wendt@rz.uni-leipzig.de](mailto:wendt@rz.uni-leipzig.de).

Your analysis results (filled-in list in **Annex 1**) and answers to the requested documentation of your station/agency magnitude procedures (**Annex 2**) should be sent to both Dr. Wendt and Prof. Dr. Peter Bormann, E-mail: [pb65@gmx.net](mailto:pb65@gmx.net).

We will analyze your response, feedback with you and pass the final outcome for further consideration to the IASPEI WG on Magnitude measurement, to the ISC and the NEIC. Good feedback and interesting results may justify to envisage a joint journal publication. Thanks in advance for your valuable co-operation.

## Acknowledgment

We thank Dr. James W. Dewey, USGS, and Prof. Lars Ottemöller, University of Bergen, for reviewing this contribution and their valuable hints, which will also be considered in future upgrading of the IS.

### Annex 1: Reference event list

### Annex 2: Documentation of station/agency magnitude procedures

## References

- Havskov, J. (Ed.) (1996). The SEISAN earthquake analysis software for the IBM PC and SUN, Version 5.2. *Institute of Solid Earth Physics, Univ. of Bergen*, August 1996.
- Havskov, J. and L. Ottemöller, Routine Processing in Earthquake Seismology, Springer, 347 pp, 2010.
- Havskov, J., L. Ottemöller and P. Voss (2011). Introduction to SEISAN and computer exercises in processing earthquake data. Printed material by the University of Bergen, Norway.
- IASPEI (2011) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf)
- Ottemöller L, P. Voss and J. Havskov (2011). SEISAN earthquake analysis software, User's manual. Printed material by the University of Bergen, Norway.

## Annex 1

## Reference event list

Below find a reference event list for the analysis of IASPEI standard magnitudes mb, mB\_BB, Ms\_20 and Ms\_BB in comparison with your station/agency procedures and types of magnitude. Note that no measurements and data entries are required in column-lines with – . The hyphen indicates that at the given event distance and/or source depth the respective magnitude is not defined. No calibration values QP or QL are given in this case. If, however, your station/center has determined a magnitude for this event, because your station/network recorded this event at a suitable distance for magnitude determination, then please give the respective value and type of magnitude in the last column.

The reference list has been printed below as a table in a compressed form with rather small letters. Magnify it for reading, plotting, filling in your answers and sending them back to the authors. Alternatively you may use the respective ASCII files: **ref\_evt\_1.text** and **ref\_evt\_2.txt** or the Exel file **ref\_evt.xls** in the directory [reference\\_events](#) (see in **Overview** on the NMSOP-2 front page for listing of *Download programs and files*).

x

## ref\_evt\_2.txt

|                                            |         |       | Time<br>hh:mm:ss.s | Per.<br>(s) | Ampl.<br>(nm,nm/s) | Magn.<br>IASPEI | Magn.<br>(your<br>method) | Magn.<br>(own netw.<br>or stat.) |
|--------------------------------------------|---------|-------|--------------------|-------------|--------------------|-----------------|---------------------------|----------------------------------|
| 2009/07/09 11:19:16.8                      | D= 2.4  | mb    | -                  | -           | -                  | -               | -                         | -                                |
| 25.636N 101.081E 10                        |         | mB_BB | -                  | -           | -                  | -               | -                         | -                                |
| mb=5.5( ) Ms= ( ) Mw(GCM)=5.7 Me=          |         | Ms_20 | -                  | -           | -                  | -               | -                         | -                                |
| YUNNAN, CHINA                              | QL=3.93 | Ms_BB | -                  | -           | -                  | -               | -                         | -                                |
| GSE2-File: 200907091119142.YN.TNC.GSE      |         |       |                    |             |                    |                 |                           |                                  |
| 2004/07/12 13:04:07.2                      | D= 5.0  | mb    | -                  | -           | -                  | -               | -                         | -                                |
| 46.296N 13.641E 8                          |         | mB_BB | -                  | -           | -                  | -               | -                         | -                                |
| mb=5.0( 82) Ms=4.9( 36) Mw(HRV)=5.2 Me=    |         | Ms_20 | -                  | -           | -                  | -               | -                         | -                                |
| AUSTRIA                                    | QL=4.46 | Ms_BB | -                  | -           | -                  | -               | -                         | -                                |
| GSE2-File: 040712_1304.bhz                 |         |       |                    |             |                    |                 |                           |                                  |
| 2009/04/05 20:20:53.0                      | D= 7.1  | mb    | -                  | -           | -                  | -               | -                         | -                                |
| 44.236N 11.999E 28                         |         | mB_BB | -                  | -           | -                  | -               | -                         | -                                |
| mb=4.6( 36) Ms= ( ) Mw(GCM)= Me=           |         | Ms_20 | -                  | -           | -                  | -               | -                         | -                                |
| NORTHERN ITALY                             | QL=4.71 | Ms_BB | -                  | -           | -                  | -               | -                         | -                                |
| GSE2-File: 090405_2020.bhz                 |         |       |                    |             |                    |                 |                           |                                  |
| 1997/09/26 09:40:26.3                      | D= 8.2  | mb    | -                  | -           | -                  | -               | -                         | -                                |
| 43.084N 12.812E 10                         |         | mB_BB | -                  | -           | -                  | -               | -                         | -                                |
| mb=5.7( 57) Ms=6.0( 50) Mw(HRV)=6.0 Me=6.1 |         | Ms_20 | -                  | -           | -                  | -               | -                         | -                                |
| CENTRAL ITALY                              | QL=4.82 | Ms_BB | -                  | -           | -                  | -               | -                         | -                                |
| GSE2-File: 970926_0940.bhz                 |         |       |                    |             |                    |                 |                           |                                  |
| 2009/04/05 09:36:26.2                      | D= 12.4 | mb    | -                  | -           | -                  | -               | -                         | -                                |
| 32.007N 131.417E 26                        |         | mB_BB | -                  | -           | -                  | -               | -                         | -                                |
| mb=5.8(172) Ms=5.4(117) Mw(GCM)=5.8 Me=5.3 |         | Ms_20 | -                  | -           | -                  | -               | -                         | -                                |
| KYUSHU, JAPAN                              | QL=5.12 | Ms_BB | -                  | -           | -                  | -               | -                         | -                                |
| GSE2-File: 200904050936204.JL.CN2          |         |       |                    |             |                    |                 |                           |                                  |
| 1999/08/17 00:01:39.1                      | D= 15.7 | mb    | -                  | -           | -                  | -               | -                         | -                                |

10



11

|                                                      |               |   |            |      |           |        |               |                      |  |
|------------------------------------------------------|---------------|---|------------|------|-----------|--------|---------------|----------------------|--|
| 2006/04/29 16:58: 6.3                                | D= 66.7 mb    |   |            |      |           |        |               |                      |  |
| 60.491N 167.516E 11                                  | QP=6.96 mb_BB |   |            |      |           |        |               |                      |  |
| mb=6.4(242) Ms=6.6( 14) Mw(HRV)=6.6 Me=6.5           | Ms_20         |   |            |      |           |        |               |                      |  |
| EASTERN SIBERIA                                      | QL=6.33 Ms_BB |   |            |      |           |        |               |                      |  |
| GSE2-File: 060429_1658.bhz                           |               |   |            |      |           |        |               |                      |  |
| 2006/07/10 07:21:37.9                                | D= 66.7 mb    |   |            |      |           |        |               |                      |  |
| 11.627S 13.432W 10                                   | QP=6.96 mb_BB |   |            |      |           |        |               |                      |  |
| mb=5.3(103) Ms=5.3(146) Mw(HRV)=5.5 Me=              | Ms_20         |   |            |      |           |        |               |                      |  |
| ASCENSION ISLAND REGION                              | QL=6.33 Ms_BB |   |            |      |           |        |               |                      |  |
| GSE2-File: 060710_0721.bhz P-SNR only 1.8            |               |   |            |      |           |        |               |                      |  |
| =====                                                |               |   |            |      |           |        |               |                      |  |
|                                                      |               |   | Time       | Per. | Ampl.     | Magn.  | Magn.         | Magn.                |  |
|                                                      |               |   | hh:mm:ss.s | (s)  | (nm,nm/s) | IASPEI | (your method) | (own netw. or stat.) |  |
| 2008/11/24 09:02:58.8                                | D= 70.0 mb    |   |            |      |           |        |               |                      |  |
| 54.203N 154.322E 492                                 | QP=6.39 mb_BB |   |            |      |           |        |               |                      |  |
| mb=6.5(267) Ms= Mw(GCM)=7.3 Me=7.0                   | Ms_20         | - | -          | -    | -         |        |               |                      |  |
| SEA OF OKHOTSK                                       | Ms_BB         | - | -          | -    | -         |        |               |                      |  |
| GSE2-File: 081124_0902.bhz                           |               |   |            |      |           |        |               |                      |  |
| 2002/06/28 17:19:30.3                                | D= 71.0 mb    |   |            |      |           |        |               |                      |  |
| 43.752N 130.666E 566                                 | QP=6.37 mb_BB |   |            |      |           |        |               |                      |  |
| mb=6.7(205) Ms= Mw(HRV)=7.3 Me=7.1                   | Ms_20         | - | -          | -    | -         |        |               |                      |  |
| E. RUSSIA-N.E. CHINA BORDER REG.                     | Ms_BB         | - | -          | -    | -         |        |               |                      |  |
| GSE2-File: 020628_1719.bhz                           |               |   |            |      |           |        |               |                      |  |
| 2003/07/15 20:27:50.5                                | D= 71.3 mb    |   |            |      |           |        |               |                      |  |
| 2.598S 68.382E 10                                    | QP=6.89 mb_BB |   |            |      |           |        |               |                      |  |
| mb=6.1(141) Ms=7.6( 69) Mw(HRV)=7.6 Me=7.6           | Ms_20         |   |            |      |           |        |               |                      |  |
| CARLSBERG RIDGE                                      | QL=6.38 Ms_BB |   |            |      |           |        |               |                      |  |
| GSE2-Files: 030715_2027.bhz + lh and 030715_2038.bhz |               |   |            |      |           |        |               |                      |  |
| 2008/06/29 20:53:04.9                                | D= 72.6 mb    |   |            |      |           |        |               |                      |  |
| 45.156N 137.446E 326                                 | QP=6.50 mb_BB |   |            |      |           |        |               |                      |  |
| mb=5.6(293) Ms= Mw(GCM)=6.0 Me=6.0                   | Ms_20         | - | -          | -    | -         |        |               |                      |  |
| NEAR SOUTHEAST COAST OF RUSSIA                       | Ms_BB         | - | -          | -    | -         |        |               |                      |  |
| GSE2-File: 080629_2053.bhz                           |               |   |            |      |           |        |               |                      |  |
| 2006/07/02 03:53:56.5                                | D= 76.5 mb    |   |            |      |           |        |               |                      |  |
| 51.713N 176.930E 9                                   | QP=6.87 mb_BB |   |            |      |           |        |               |                      |  |
| mb=5.5(218) Ms=5.3(151) Mw(HRV)=5.7 Me=              | Ms_20         |   |            |      |           |        |               |                      |  |
| RAT ISLANDS, ALEUTIAN ISLANDS                        | QL=6.43 Ms_BB |   |            |      |           |        |               |                      |  |
| GSE2-Files: 060702_0353.bhz + lh and 060702_0405.bhz | P-SNR = 2.3   |   |            |      |           |        |               |                      |  |
| 2007/02/17 00:02:56.8                                | D= 77.8 mb    |   |            |      |           |        |               |                      |  |
| 41.794N 143.553E 31                                  | QP=6.90 mb_BB |   |            |      |           |        |               |                      |  |
| mb=5.9(209) Ms=5.9(179) Mw(GCM)=6.0 Me=5.8           | Ms_20         |   |            |      |           |        |               |                      |  |
| HOKKAIDO, JAPAN REGION                               | QL=6.44 Ms_BB |   |            |      |           |        |               |                      |  |
| GSE2-Files: 070217_0002.bhz + lh and 070217_0014.bhz |               |   |            |      |           |        |               |                      |  |
| 2003/09/25 19:50: 6.4                                | D= 77.9 mb    |   |            |      |           |        |               |                      |  |
| 41.815N 143.910E 27                                  | QP=6.80 mb_BB |   |            |      |           |        |               |                      |  |
| mb=6.9(220) Ms=8.1(114) Mw(HRV)=8.3 Me=8.0           | Ms_20         |   |            |      |           |        |               |                      |  |
| HOKKAIDO, JAPAN REGION                               | QL=6.44 Ms_BB |   |            |      |           |        |               |                      |  |
| GSE2-Files: 030925_1950.bhz + lh and 030925_2001.bhz |               |   |            |      |           |        |               |                      |  |
| =====                                                |               |   |            |      |           |        |               |                      |  |
|                                                      |               |   | Time       | Per. | Ampl.     | Magn.  | Magn.         | Magn.                |  |
|                                                      |               |   | hh:mm:ss.s | (s)  | (nm,nm/s) | IASPEI | (your method) | (own netw. or stat.) |  |
| 2005/08/16 02:46:28.4                                | D= 80.3 mb    |   |            |      |           |        |               |                      |  |
| 38.276N 142.039E 36                                  | QP=6.76 mb_BB |   |            |      |           |        |               |                      |  |
| mb=6.5(220) Ms=6.8(136) Mw(HRV)=7.2 Me=6.7           | Ms_20         |   |            |      |           |        |               |                      |  |

13

14

|                                              |         |       |            |      |           |        |               |                      |   |
|----------------------------------------------|---------|-------|------------|------|-----------|--------|---------------|----------------------|---|
| GSE2-Files: 001117_2101.bhz + lhz            |         |       |            |      |           |        |               |                      |   |
| 2000/11/16 07:42:16.9                        | D=123.4 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 5.233S 153.102E 30                           |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=6.2( 40) Ms=7.8( 83) Mw(HRV)=7.8 Me=7.2   |         | Ms_20 |            |      |           |        |               |                      |   |
| NEW IRELAND REGION                           | QL=6.77 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 001116_0742.bhz + lhz            |         |       |            |      |           |        |               |                      |   |
| 1995/08/16 10:27:28.6                        | D=124.4 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 5.799S 154.178E 30                           |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=6.5(106) Ms=7.8( 71) Mw(HRV)=7.7 Me=7.3   |         | Ms_20 |            |      |           |        |               |                      |   |
| SOLOMON ISLANDS                              | QL=6.78 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 950816_1027.bhz + lhz            |         |       |            |      |           |        |               |                      |   |
| =====                                        |         |       |            |      |           |        |               |                      |   |
|                                              |         |       | Time       | Per. | Ampl.     | Magn.  | Magn.         | Magn.                |   |
|                                              |         |       | hh:mm:ss.s | (s)  | (nm,nm/s) | IASPEI | (your method) | (own netw. or stat.) |   |
| 1996/08/02 12:55:29.3                        | D=132.1 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 10.769S 161.445E 33                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=6.2( 68) Ms=7.1( 49) Mw(HRV)=6.9 Me=6.8   |         | Ms_20 |            |      |           |        |               |                      |   |
| SOLOMON ISLANDS                              | QL=6.82 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 960802_1255.bhz + lhz            |         |       |            |      |           |        |               |                      |   |
| 1997/08/17 20:11:10.8                        | D=137.1 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 13.592S 167.391E 26                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=5.4( 47) Ms=6.1( 57) Mw(HRV)=6.0 Me=      |         | Ms_20 |            |      |           |        |               |                      |   |
| VANUATU ISLANDS                              | QL=6.85 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 970817_2011.bhz + lhz            |         |       |            |      |           |        |               |                      |   |
| 2008/11/07 07:19:35.7                        | D=138.4 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 14.829S 168.032E 13                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=6.0(192) Ms=6.3(184) Mw(GCM)=6.4 Me=6.3   |         | Ms_20 |            |      |           |        |               |                      |   |
| VANUATU                                      | QL=6.85 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 081107_0719.bhz + lhz            |         |       |            |      |           |        |               |                      |   |
| =====                                        |         |       |            |      |           |        |               |                      |   |
|                                              |         |       | Time       | Per. | Ampl.     | Magn.  | Magn.         | Magn.                |   |
|                                              |         |       | hh:mm:ss.s | (s)  | (nm,nm/s) | IASPEI | (your method) | (own netw. or stat.) |   |
| 1997/08/29 06:54: 0.2                        | D=143.4 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 15.235S 175.576W 33                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=5.6( 60) Ms=6.4( 54) Mw(HRV)=6.4 Me=6.7   |         | Ms_20 |            |      |           |        |               |                      |   |
| TONGA ISLANDS                                | QL=6.88 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 970829_0654.bhz + lhz            |         |       |            |      |           |        |               |                      |   |
| 1995/04/07 22:06:56.9                        | D=143.6 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 15.199S 173.529W 21                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=6.8( 90) Ms=8.0( 46) Mw(HRV)=7.4 Me=      |         | Ms_20 |            |      |           |        |               |                      |   |
| TONGA ISLANDS                                | QL=6.88 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-File: 950407_2206.lhz                   |         |       |            |      |           |        |               |                      |   |
| 2007/03/25 00:40: 1.6                        | D=144.2 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 20.617S 169.357E 34                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=6.5( 77) Ms=7.0(151) Mw(GCM)=7.1 Me=6.8   |         | Ms_20 |            |      |           |        |               |                      |   |
| VANUATU ISLANDS                              | QL=6.88 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 070325_0040.bhz + lhz superposed |         |       |            |      |           |        |               |                      |   |
| 2007/03/25 01:08:19.1                        | D=144.3 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 20.754S 169.354E 35                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |
| mb=5.9( 50) Ms=7.1( 6) Mw(GCM)=6.9 Me=       |         | Ms_20 |            |      |           |        |               |                      |   |
| VANUATU ISLANDS                              | QL=6.88 | Ms_BB |            |      |           |        |               |                      |   |
| GSE2-Files: 070325_0040.bhz + lhz superposed |         |       |            |      |           |        |               |                      |   |
| 1995/05/16 20:12:44.2                        | D=146.6 | mb    | -          | -    | -         | -      | -             | -                    | - |
| 23.008S 169.900E 20                          |         | mB_BB | -          | -    | -         | -      | -             | -                    | - |

16

## Annex 2

**DOCUMENTATION OF STATION/AGENCY MAGNITUDE PROCEDURES**

(Copied from the SUMMARY OF IASPEI MAGNITUDE WORKING GROUP RECOMMENDATIONS ON DETERMINING EARTHQUAKE MAGNITUDES FROM DIGITAL DATA, updated version 2011; see [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf))

Documentation of procedures for amplitude/period based magnitudes by seismographic stations/analysis centers should generally address the following points:

1. Phase-type from which the amplitude measurement is made.
2. Units of the reported amplitude. Specify if amplitudes are reported in units of trace-amplitude motion instead of ground motion.
3. Time-window in which the amplitude measurement is made.
  - a. For example, a flexible time-interval between the P onset and the PP onset or a fixed time window after the first P onset (e.g. 5 s, 10s or other).
  - b. For example, the time-interval spanned by waves having group-velocities between 3.2 and 3.8 km/s.
4. Amplitude-response, filter characteristics, or transfer-function of the seismograph or simulated seismograph through which the amplitude measurement is made.
5. Orientation of seismograph (horizontal or vertical) from which the measurement is made.
6. Details of measuring amplitude
  - a. For example, does the amplitude correspond to  $0.5 \times (\text{peak-to-trough amplitude})$ , where “peak-to-trough amplitude” corresponds to difference between a maximum positive excursion and a maximum negative excursion of the trace, or is the amplitude instead measured as the maximum absolute excursion from the “zero” position of the seismograph trace?
  - b. For example, if the amplitude corresponds to  $0.5 \times (\text{peak-to-trough amplitude})$ , are the “peak” and “trough” respectively the absolute maximum and absolute minimum values of the entire wave-group, or are they the adjacent peak and trough corresponding to the maximum trace excursion that is associated with a single zero-crossing?
  - c. For example, are displacement amplitude(A) and period(T) measured at the time of maximum A or at the time of the maximum of the quotient (A/T)?
  - d. For example, any way in which station/agency procedure differs from the corresponding IASPEI standard procedure.

7. Details of measuring period, for example, is it the time between the neighboring peaks, respectively troughs or twice the time span measured between the largest peak and adjacent trough at which the double amplitude has been measured.
8. To what part of a phase the amplitude-measurement time refers. For example, is the amplitude-measurement time the time of the zero-crossing associated with a peak-to-adjacent trough measurement or is it the time of an absolute maximum or absolute minimum?
9. The equations that are used for calculating particular types of magnitudes
  - a. Specify if distance is measured as epicentral distance or hypocentral distance.
  - b. Specify the distance range for which the equation is applied.
  - c. Specify restrictions on hypocentral focal-depth, if any.
10. Other restrictions on the calculation of the magnitude
  - a. For example, must signal-amplitude satisfy a signal-to-noise ratio criterion?
  - b. For example, is the magnitude measured only for earthquakes of a certain size, as defined by an independent measure of earthquake size?
11. Any way in which the magnitude procedure departs from an IASPEI standard procedure if the procedure is intended to produce the IASPEI standard magnitude.
  - a. For example, if measurements are made over a somewhat different distance or depth range than specified in the IASPEI procedure.
  - b. For example, if non-standard filtering is applied in some situations in which the non-standard filtering **is not expected to affect the magnitude** (e.g., when measuring magnitudes for very small events for which standard-filtering would produce records of too small signal-to-noise ratio). Please specify.
12. How data are averaged for a network average.
  - a. For example, is the network average a simple arithmetic mean, a trimmed mean, the median value, or some other average?
  - b. If, as for example might be the case for ML, data from each of two horizontal components at a single station are used, are data from each component treated as a separate observation in the network average, or are the two components first averaged into a station magnitude, which is then treated as a single observation in the network average?

For some magnitude types (e.g., coda-based magnitudes) the procedures to be documented will have to be substantially altered from the preceding to better reflect the measurements that are critical to those magnitude-types, but the level of detail should be analogous to what is proposed above for amplitude/period based magnitudes.



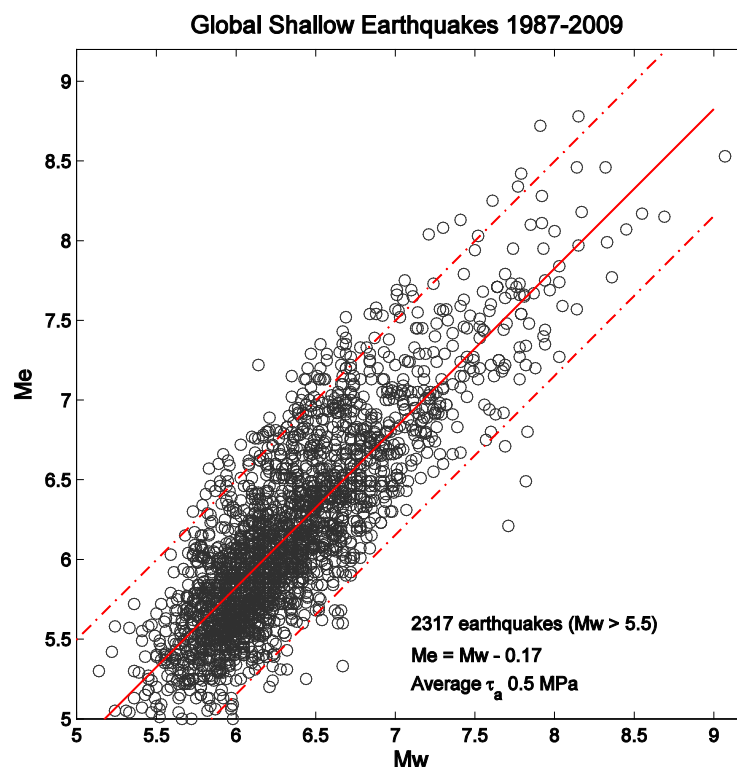
Please insert your answers to the above questions in color for easier recognition and send them back to

Dr. Wendt ([wendt@rz.uni-leipzig.de](mailto:wendt@rz.uni-leipzig.de)) and  
Prof. Bormann ([pb65@gmx.net](mailto:pb65@gmx.net)).

|                |                                                                                                         |
|----------------|---------------------------------------------------------------------------------------------------------|
| <b>Title</b>   | <b>Stress conditions inferable from modern magnitudes:<br/>development of a model of fault maturity</b> |
| <b>Author</b>  | <b>George L. Choy</b> , U. S. Geological Survey, Denver, CO 80215,<br>E-mail: choy@usgs.gov             |
| <b>Version</b> | August 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_3.5                                                            |

## 1 Introduction

Although  $M_e$  and  $M_w$  are magnitudes that describe the size of an earthquake, they are not equivalent.  $M_e$ , being derived from velocity power spectra, is a measure of the radiated energy in the form of seismic waves and, thus, of the seismic potential for damage to anthropogenic structures.  $M_w$ , being derived from the low-frequency asymptote of displacement spectra, is physically related to the final static displacement of an earthquake. As seen in the  $M_e$ - $M_w$  plot of global earthquakes from 1987-2009 (Figure 1), for any given  $M_w$ , the corresponding  $M_e$  may vary by as much as an entire magnitude unit and vice versa.



**Figure 1**  $M_e$ - $M_w$  for global shallow (depth < 70 km) earthquakes with magnitude greater than about 5.5.  $M_e$  are taken from the USGS/NEIC source parameters catalog which has included direct computation of radiated energy  $E_s$  since 1987.  $M_w$  are taken from the global Centroid Moment Tensor (gCMT) catalog. The least-squares linear regression with an assumed slope of one (solid line) yields a global average apparent stress for earthquakes of 0.5 MPa. But the 95% spread about the mean (indicated by the dashed lines) shows that for any given  $M_w$ ,  $M_e$  scatters about one-half a magnitude unit above and below the mean value, with some outliers up to one magnitude unit. No single empirical formula can adequately represent such a spread.

Events with anomalously high  $M_e$  relative to  $M_w$  should have greater potential for seismic damage than those earthquakes with relatively low  $M_e$ . Indeed, Choy et al. (2001) [see also Table 3.2 in Chapter 3 of the first edition of the NMSOP (Bormann et al., 2002)] first demonstrated this dramatically by comparing the significantly different macroseismic effects of two earthquakes in the same epicentral region that had nearly the same  $M_w$  but had  $M_e$ 's differing by 1.0 unit. The objective of this Information Sheet is to demonstrate that the apparent scatter in  $M_e$ - $M_w$  is non-random. Subsets of earthquakes chosen on the basis of a specific tectonic setting, seismic region and focal-mechanism type occupy specific sectors of the  $M_e$ - $M_w$  plot. Moreover, the stress conditions unique to specific sectors can be incorporated into a model of fault maturity. Rather than be daunted by the large spread in  $M_e$ - $M_w$ , methods of estimating tsunami and seismic hazards can be enhanced by exploiting the patterns of stress conditions available from  $M_e$ - $M_w$  data.

## 2 Ways to measure the amount of energy released per unit moment

There are several equivalent ways to represent the radiated energy to moment ratio of an earthquake:

- Apparent stress,  $\tau_a = \mu E_S / M_0$
- Scaled energy,  $E_S / M_0$
- Slowness parameter,  $\Theta = \log( E_S / M_0 )$
- Differential magnitude,  $\Delta M = M_e - M_w$

The advantage of using apparent stress  $\tau_a$  (where  $\mu$  is rigidity) is that it can be related to other stresses associated with rupture (Wyss and Brune, 1968). Although the relationship between apparent stress and stress drop is model dependent, larger apparent stress generally implies larger stress drop. The dimensionless ratio  $E_S/M_0$ , often called scaled energy (Kanamori and Heaton, 2000), is independent of  $\mu$ , but is slightly cumbersome to enunciate as it typically ranges from  $10^{-4}$ - $10^{-6}$  for teleseismically analyzable earthquakes. Taking the log of the scaled energy yields a more manageable number which Newman and Okal (1998) call the slowness parameter  $\Theta$ . Another equivalent method presented here is differential magnitude  $\Delta M$ , defined as the difference between the energy and moment magnitudes,  $M_e$  and  $M_w$ ,

$$\Delta M = M_e - M_w \quad (1)$$

From the equations for  $M_e$  and  $M_w$ , we can derive useful relationships between the various representations. We use  $M_w = (2/3)(\log M_0 - 9.1)$  and  $M_e = (2/3)(\log E_S - 4.4)$ . These forms of  $M_w$  and  $M_e$ , first proposed in Chapter 3 of the NMSOP (Bormann et al., 2002) and accepted by IASPEI (2005) as standard formulas, avoid occasional rounding errors up to 0.1 m.u. which can occur in the formulas originally published by Hanks and Kanamori (1979) for  $M_w$  and by Choy and Boatwright (1995) for  $M_e$ .

Between differential magnitude and scaled energy we have

$$\Delta M = (2/3) [\log ( E_S/M_0 ) + 4.7] \quad (2)$$

Between differential magnitude and  $\Theta$  :

$$\Delta M = (2/3) [\Theta + 4.7]. \quad (3)$$

An important relationship between  $M_e$ ,  $M_0$ , and  $\tau_a$  was recognized [eq. 11 of Choy and Boatwright (1995)] as  $M_e = (2/3) [\log M_0 + \log(\tau_a/\mu)] - 3.2$ , which can also be written as

$$M_e = (2/3) [\log M_0 + \log(\tau_a/\mu) - 4.8], \quad (4)$$

The constant 4.8 corresponds to the constant in the classical Gutenberg-Richter relationship  $\log E_s = 4.8 + 1.5 M_s$ . However, the U. S. Geological Survey's National Earthquake Information Center accepted in its Monthly Listings of the Preliminary Determination of Epicenters July 1995 the new Choy and Boatwright (1995) constant 4.4, derived from the best fitting slope of 1.5 through directly measured  $\log E_s$  data vs.  $M_s$  [see Figure 4 and eq. 6 in Choy and Boatwright (1995)]. Replacing in eq. (4) the constant 4.8 by 4.4 yields

$$M_e = (2/3) [\log M_0 + \log(\tau_a/\mu) - 4.4], \quad (5)$$

and when introducing therein the IASPEI (2005) standard formula  $M_w = (2/3) (\log M_0 - 9.1)$

$$M_e = M_w + (2/3) [\log(\tau_a/\mu) + 4.7]. \quad (6)$$

Equivalent relationships between  $M_e$ ,  $M_w$  and apparent stress were published by Bormann and Di Giacomo (2011).

The relationship between differential magnitude and apparent stress is then:

$$\Delta M = (2/3) [\log(\tau_a/\mu) + 4.7]. \quad (7)$$

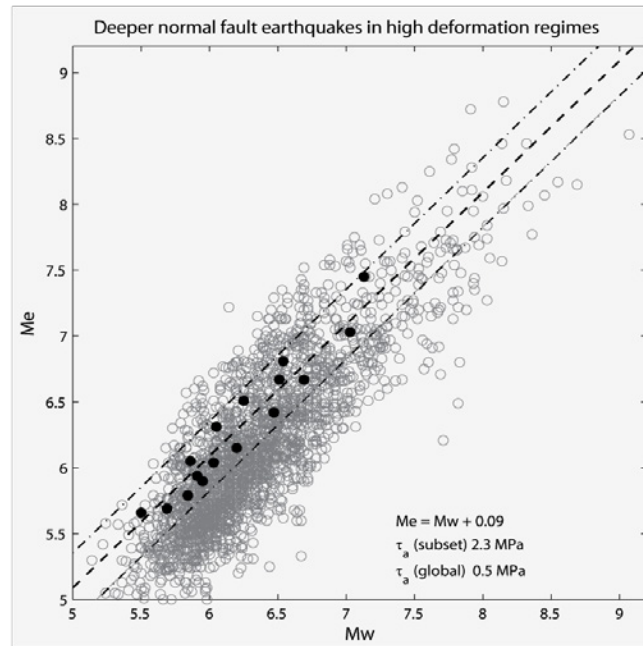
In the  $M_e$ - $M_w$  plots that follow,  $M_e$ - $M_w$  data pairs for earthquakes based on tectonic setting and focal mechanism are fit by a least squares regression to a line with slope 1,

$$M_e = M_w + c.$$

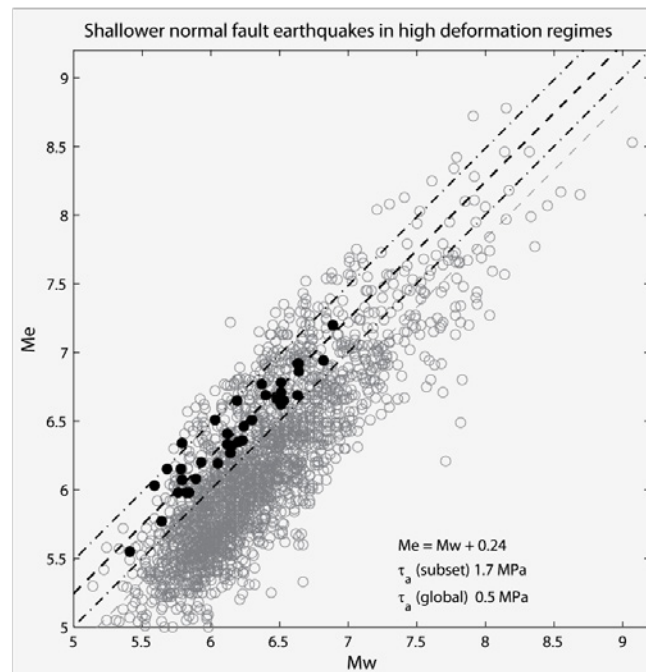
The apparent stress is related to the constant  $c$  through eq. (6). Note that in order to compute average  $\tau_a$  we use  $\mu$  appropriate to the source depth for each event.

### 3 Identifying events that radiate exceptionally high energy

To identify an earthquake as having radiated exceptionally high energy relative to its moment, we adopt the criterion of Choy and Kirby (2004) that  $\tau_a > 1$  MPa or, equivalently,  $\Delta M$  greater than about 0.0 (i.e., whenever  $M_e \geq M_w$ ). This criterion was derived from their global investigation of subduction-zone earthquakes, in which they found that normal-fault earthquakes occurring in high-deformation regimes were always associated with exceptionally higher energy release than other normal-fault earthquakes. Characteristics of high-deformation regimes include sharp changes in slab geometry, colliding slabs and oblique convergence at a subducting plate boundary. Less than 20% of all normal-fault earthquakes are energetic events. Figures 2 and 3 show that these high energy events occupy narrow sectors in the  $M_e$ - $M_w$  plot.



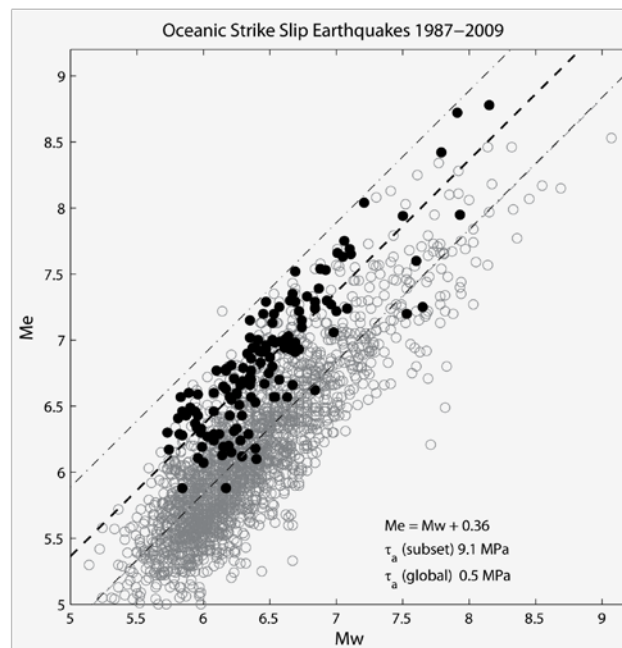
**Figure 2** Me-Mw of deeper (  $35 \text{ km} < h < 70 \text{ km}$  ) normal-fault events (solid black circles) that occur at a sharp change in slab geometry or in the deformation zone of converging but oppositely oriented slabs. These events have an average apparent stress nearly 4 times higher than the global population (gray circles).



**Figure 3** Me-Mw of shallower ( $h < 35 \text{ km}$ ) events (solid black circles) that occur intraslab and landward of the trench axis at subduction boundaries where plate convergence is oblique to the fabric of the seafloor. These events have an average apparent stress more than 3 times higher than that of the global data (gray circles).

The subset of intraoceanic strike-slip events occupies a unique sector of the Me-Mw plot. Although they constitute only 5% of the global shallow earthquakes, they have the highest average  $\tau_a$  of any earthquake subset. Figure 4 plots Me-Mw for the set of strike-slip

earthquakes in the oceans (excluding subduction zones). The average  $\tau_a$  for oceanic strike-slip earthquakes is about 18 times higher than the global average. At first glance this seems contrary to the original simple tenets of plate tectonic tenets, wherein strike-slip earthquakes in the oceans are thought to be dominated by the simple slippage along transform faults of weak material. Choy and McGarr (2002), however, have shown that the actual situation is far different: more oceanic strike-slip earthquakes occur in the vicinity (less than a couple of degrees) of transforms than on the transforms themselves. Far from being a simple plate boundary, ridge-transform systems are apparently the site plate boundary reorganization. The intraplate earthquakes are the consequence of fracture on new faults that result from locally intense deformation. Even the events that do occur on transforms either (1) have nodal planes not coincident with the strike of the transform; (2) occur on short-offset transform segments; or (3) occur on the inside corners of the ridge-transform intersect. These conditions also require the fracture of fresh rock or newly formed faults as the local plate boundary evolves. Moreover, the depths of nucleation of these earthquakes have been shown from broadband seismogram modeling (Choy and McGarr, 2002; Abercrombie and Ekström, 2001) to be between 5-25 km, well within the oceanic mantle. Thus, the higher strength of oceanic vs. continental mantle also contributes to the high energy release.

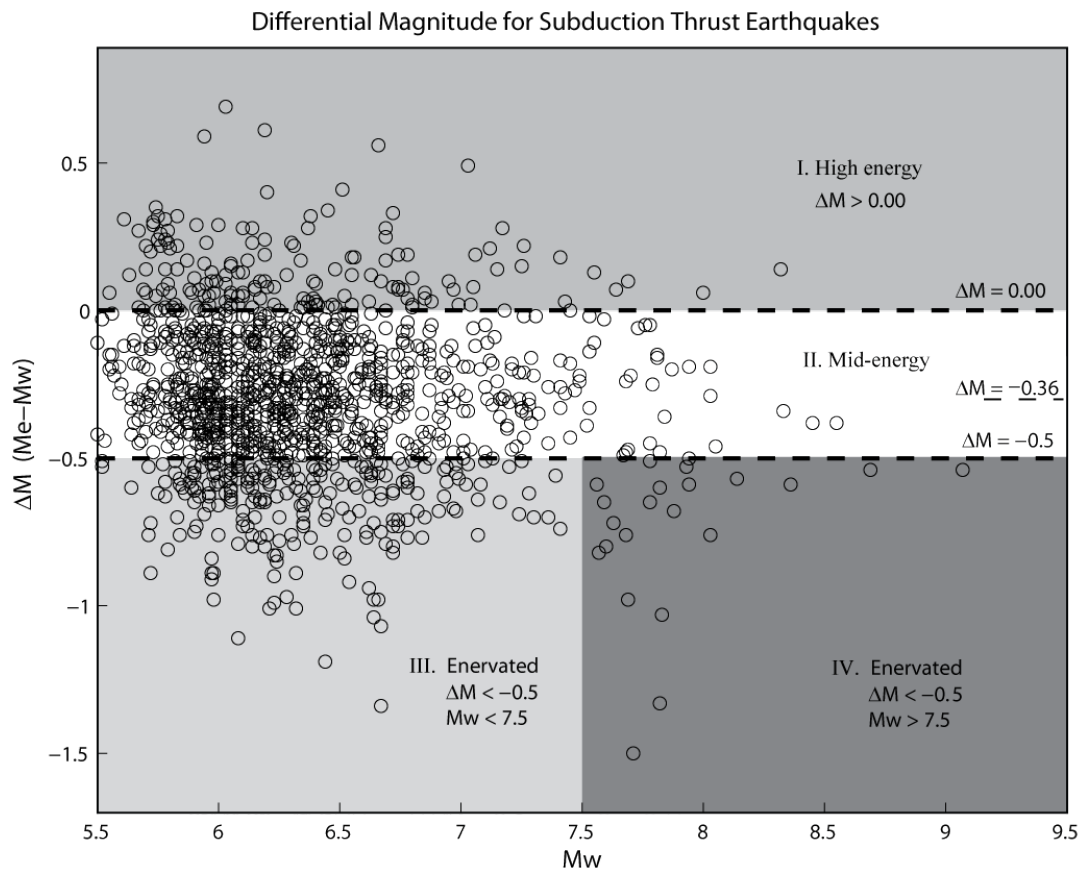


**Figure 4** Me vs. Mw for the subset of 135 intraoceanic strike-slip earthquakes (dark solid circles) occurring on or near transform faults or intraplate compared to that of global earthquakes (gray circles).

#### 4 Identifying events that radiate exceptionally low energy

To characterize an earthquake as being enervated, that is, having radiated exceptionally low energy, we adopt the criterion of Newman and Okal (1998) that its slowness parameter  $\Theta < -5.5$ . This is equivalent to a differential magnitude  $\Delta M$  less than about -0.50. Figure 5 plots  $\Delta M$  vs.  $M_W$  for all thrust earthquakes that occurred in subduction environments. The global average  $\Delta M$  is -0.36. Earthquakes which radiate anomalously low energy fall below the line  $\Delta M = -0.5$  (Quadrants III and IV). Quadrant IV contains the class of enervated earthquakes

with  $M_w > 7.5$ . Earthquakes in this class have been called “slow” earthquakes because of their abnormally long source durations and anomalously low radiated energy relative to seismic moment (Newman and Okal, 1998). There are only 22 such events, but virtually all of them have generated tsunamis. It should be noted that in quadrant III are another 257 earthquakes that also radiated anomalously low energy but with smaller magnitudes,  $M_w$ 's  $< 7.5$ . The earthquakes in quadrants III and IV do not necessarily occur in the same seismic regions. Nor is there any obvious spatial correlation of the smaller enervated events with locations of notable slow tsunami earthquakes. This may be because patterns have not yet developed in the relatively short time for which accurate radiated energies have been available. Whether there are other tectonic, geophysical or geological connections between the large tsunamigenic earthquakes and the smaller enervated earthquakes is not known at this time and requires further research. Finally, we note that the combined population of mid-energy and enervated earthquakes (i.e., quadrants II, III and IV with  $\Delta M < 0.0$ ) are predominantly interface thrust events for which the average  $\tau_a$  a little less than 0.3 MPa. The class of subduction-interface thrust events, thus, has the lowest  $\tau_a$  of any subset of earthquakes based on tectonic setting and focal mechanism.



**Figure 5**  $\Delta M$  vs.  $M_w$  for 1306 thrust earthquakes occurring in subduction regions from 1987-2009. The global average  $\Delta M$  is -0.36 (short dashed line). Earthquakes radiating anomalously low energy fall below the line  $\Delta M = -0.5$  (lower dashed line). Earthquakes radiating anomalously high energy fall above the line  $\Delta M = 0.0$ .

Note that in quadrant I there are 163 subduction-thrust earthquakes with  $\Delta M > 0.0$  which, by the criterion of our previous section, are considered to be high energy events. These

earthquakes comprise only 12% of the subduction-thrust population. They are most often found in subduction regions involving complex plate deformation (Choy and Kirby, 2009). These region types include: marine collision zones involving seamount chains or fracture zones, submerged continent-continent collisions, colliding slabs, regions of multiple plate interactions, and slab distortions. Many of these events may be intraslab based on their greater depths (compared to shallower events that are presumed to be on the subduction interface) as well as their focal mechanisms whose nodal planes are not aligned with the slab interface.

## 5 Development of a fault maturity model

The first three rows of Table 1 below summarize in ascending order the average apparent stresses of earthquake populations we have described, their respective tectonic settings and their dominant focal mechanism type. If we define fault maturity as the amount of offset accumulated by a fault, then we see that fault maturity is inversely related to average apparent stress. At one extreme, the least mature faults would have had little to no previously accumulated displacement and, hence, they should have maximum fault roughness. An example would be strike-slip earthquakes in oceanic lithosphere, which have the highest average  $\tau_a$  of any class of earthquake. The majority of these earthquakes are either intraplate or they occur on short transforms or the inside corners of ridge-transforms (where fresh material is being ruptured). At the other extreme, the most mature faults would have large total displacements. This is typified by subduction-thrust earthquakes occurring on the frictional contact between overriding and subducting plates. Thus, the level of  $\tau_a$  appears to be related to the degree to which lithosphere can sustain strain accumulation before rupturing.

**Table 1** The inverse relationship between average  $\tau_a$  ( $\langle \tau_a \rangle$ ) of a tectonic setting and fault maturity.

| $\langle \tau_a \rangle$ MPa | <div> <div>Low</div> <div>0.3</div> <div>0.4</div> <div>1.7</div> <div>2.1</div> <div>High</div> <div>9.1</div> </div> |                                                     |                                                       |                                                           |              |
|------------------------------|------------------------------------------------------------------------------------------------------------------------|-----------------------------------------------------|-------------------------------------------------------|-----------------------------------------------------------|--------------|
| Subduction zone environment  | Interplate                                                                                                             | Outer-rise/Near-trench reactivated mid-ocean fabric | Outer-rise/Near trench cross-cutting mid-ocean fabric | Intraslab High deformation (slab bends and dueling slabs) | Intraoceanic |
| Mechanism                    | Thrust                                                                                                                 | Normal                                              | Normal                                                | Normal                                                    | Strike-slip  |
| Fault Maturity               | <div> <div>High</div> <div>Low</div> </div>                                                                            |                                                     |                                                       |                                                           |              |

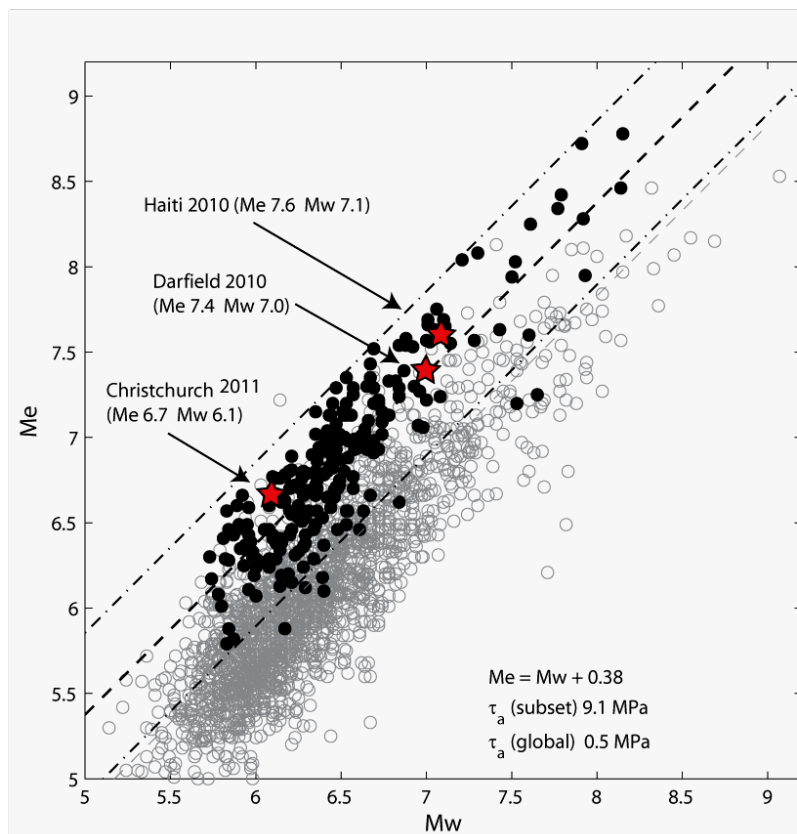
Although the range of apparent stress for teleseismically analyzable earthquakes is a continuum from less than 0.05 MPa up to 250 MPa, the outlier populations of earthquakes that radiate anomalously low or high amounts of energy are easily identifiable. The criteria stated in terms of apparent stress, slowness parameter  $\Theta$  and differential magnitude  $\Delta M$  are summarized in Table 2. These outlier populations have significant implications for tsunami



and seismic hazards evaluation, respectively. Slow earthquakes associated with transoceanic tsunamis have been characterized as having abnormally long source durations and anomalously low radiated seismic energy. On the other hand, the  $M_w$  of highly energetic earthquakes might belie the actual potential for damage from shaking. Figure 6 highlights three recent events (red stars) which inflicted considerable destruction or generated macroseismic effects. The  $\Delta M$ 's of the Haiti 2010, Darfield, New Zealand 2010 and Christchurch, New Zealand 2011 earthquakes range from 0.4 to 0.7. From these  $\Delta M$  values we would rank their faults as immature and capable of higher seismic potential than their  $M_w$  alone would imply.

**Table 2** Approximate relationship between apparent stress  $\tau_a$ , slowness  $\Theta$ , and differential magnitude  $\Delta M$ . The relations are approximate as the value of shear modulus  $\mu$  is a function of depth and earth structure. Thus,  $\tau_a$  is influenced more by focal depth than  $\Theta$  and  $\Delta M$ .

|                        | Slow tsunami earthquake | Energized earthquake | Global average    | High energy            |
|------------------------|-------------------------|----------------------|-------------------|------------------------|
| $\tau_a$               | $< 0.1 \text{ MPa}$     | $< 0.1 \text{ MPa}$  | $0.5 \text{ MPa}$ | $\geq 1.0 \text{ MPa}$ |
| $\Theta$               | $< -5.5, M_w \geq 7.5$  | $< -5.5, M_w < 7.5$  | $-4.9$            | $\geq -4.6$            |
| $\Delta M (M_e - M_w)$ | $< -0.5, M_w \geq 7.5$  | $< -0.5, M_w < 7.5$  | $-0.2$            | $\geq 0.0$             |



**Figure 6** The red stars highlight some recent damaging earthquakes with  $M_e \gg M_w$ : the Haiti 2010, Darfield, New Zealand 2010 and Christchurch, New Zealand 2011 earthquakes. For comparison, global earthquake  $M_e$ - $M_w$  data are plotted as gray open circles and black closed circles show the subset of oceanic strike-slip earthquakes which are categorized as immature.

## 6 Summary

Modern magnitudes  $M_w$  and  $M_e$  are direct measures of radiated energy and seismic moment, respectively. The amount of energy radiated per unit of seismic moment, representable in several ways such as differential magnitude  $\Delta M$  and apparent stress  $\tau_a$ , can be interpreted as an indicator of stress conditions in the lithosphere. Given the observed scatter in  $M_e$ - $M_w$  plots for the global earthquake population, no single empirical formula can accurately predict  $M_w$  from  $M_e$ . However, the  $M_w$  and  $M_e$  for subsets of earthquakes chosen on the basis of a specific tectonic setting, seismic region and focal-mechanism type occupy specific sectors of the  $M_e$ - $M_w$  plot. These patterns can be exploited to identify conditions of elevated seismic hazard, to identify conditions of elevated tsunami potential, and to be related to a model of fault maturity.

## Acknowledgment

The author is deeply indebted to Dr. Peter Bormann for comprehensive comments and detailed suggestions that significantly improved the manuscript. The author also benefited from reviews by Drs. Domenico DiGiacomo, Emily So and Morgan Moschetti.

## References

- Abercrombie, R.E. and Ekström G. (2001). Earthquake slip on oceanic transform faults. *Nature*, **410**, 74-76.
- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bormann, P., Baumbach, M., Bock, M., Grosser, H., Choy, G. L., and Boatwright J. (2002). Seismic sources and source parameters in: Bormann, P. (ed), *IASPEI New Manual Seismological Observatory Practice*, GeoForschungsZentrum Potsdam, Vol. 1, Chapter 3, 94 pp.
- Choy, G. L., and Boatwright, J. L. (1995). Global patterns of radiated seismic energy and apparent stress. *J. Geophys. Res.*, **100**, B9, 18,205-18,228.
- Choy, G. L., Boatwright, J., and Kirby, S. (2001). The radiated seismic energy and apparent stress of interplate and intraslab earthquakes at subduction zone environments: Implications for seismic hazard estimation. *U.S. Geological Survey Open-File Report 01-0005*, 18 pp.
- Choy, G. L., and McGarr, A. (2002). Strike-slip earthquakes in the oceanic lithosphere: observations of exceptionally high apparent stress. *Geophys. J. Int.*, **150**, 506-523.
- Choy, G. L., and Kirby, S. (2004). Apparent stress, fault maturity and seismic hazard for normal-fault earthquakes at subduction zones. *Geophys. J. Int.*, **159**, 991-1012.
- Choy, G. L., and Kirby, S. H. (2009). Patterns in the spatial distribution of enervated and energetic subduction earthquakes based on broadband seismic wave-forms: Scientific Program of the USGS Tsunami Source Working Group, Menlo Park, CA.
- Hanks, T. C., and Kanamori, H. (1979). Moment magnitude. *J. Geophys. Res.*, **84**, 2348-2350.

- Kanamori, H., and Heaton, T. H. (2000). Microscopic and macroscopic physics of earthquakes. In: *GeoComplexity and the Physics of Earthquakes*, edited by J. B. Rundle, D. L. Turcotte, and W. Klein, Washington, D. C., AGU, 147–163.
- Newman, A.V., and Okal, E.A. (1998). Teleseismic estimates of radiated seismic energy: The E/Mo discriminant for tsunami earthquakes. *J. Geophys. Res.*, **103**, 26885-26898.
- Wyss, M., and Brune, J.N. (1968). Seismic moment, stress, and source dimensions for earthquakes in the California-Nevada regions. *J. Geophys. Res.*, **73**, 4681-4694.

| Title   | Radiated seismic energy and energy magnitude                                                                                                  |
|---------|-----------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | George L. Choy and John L. Boatwright,<br>U. S. Geological Survey, Denver, CO 80215; E-mail: <a href="mailto:choy@usgs.gov">choy@usgs.gov</a> |
| Version | August 2002; editorially adapted and amended July 2012;<br>DOI: 10.2312/GFZ.NMSOP-2_IS_3.6                                                    |

|                                                                            | Page |
|----------------------------------------------------------------------------|------|
| <b>1 Introduction</b>                                                      | 1    |
| <b>2 How is radiated seismic energy measured?</b>                          | 2    |
| 2.1 Method                                                                 | 2    |
| 2.2 Data                                                                   | 4    |
| <b>3 Development of an energy magnitude, <math>M_e</math></b>              | 5    |
| <b>4 The relationship of radiated energy to moment and apparent stress</b> | 6    |
| <b>5 The relationship of <math>M_e</math> to <math>M_w</math></b>          | 7    |
| <b>6 Regional estimates of radiated seismic energy</b>                     | 8    |
| <b>7 Conclusions</b>                                                       | 8    |
| <b>Acknowledgment</b>                                                      | 8    |
| <b>Recommended overview readings</b>                                       | 9    |
| <b>References</b>                                                          | 9    |

## 1 Introduction

One of the most fundamental parameters for describing an earthquake is radiated seismic energy. In theory, its computation simply requires an integration of radiated energy flux in velocity-squared seismograms. In practice, energy has historically almost always been estimated with empirical formulas. The empirical approach dominated for two major reasons. First, until the 1980's most seismic data were analog, a format which was not amenable to spectral processing on a routine basis. Second, an accurate estimate of radiated energy requires the analysis of spectral information both above and below the corner frequency of an earthquake, about which energy density is most strongly peaked.

Prior to the worldwide deployment of broadband seismometers, which started in the 1970's, most seismograms were recorded by conventional seismographs with narrowly peaked instrument responses. The difficulties in processing analog data were thus compounded by the limitations in retrieving reliable spectral information over a broad bandwidth. Fortunately, theoretical and technological impediments to the direct computation of radiated energy have been removed. The requisite spectral bandwidth is now recorded digitally by a number of seismograph networks and arrays with broadband capability, and frequency-dependent corrections for source mechanism and wave propagation are better understood now than at the time empirical formulas were first developed.

## 2 How is radiated seismic energy measured?

### 2.1 Method

The method described below for estimating the radiated seismic energy of teleseismic earthquakes is based on Boatwright and Choy (1986). Velocity-squared spectra of body waves are corrected for effects arising from source mechanism, depth phases, and propagation through the Earth.

For shallow earthquakes where the source functions of direct and surface-reflected body-wave arrivals may overlap in time, the radiated energy of a  $P$ -wave group (consisting of  $P$ ,  $pP$  and  $sP$ ) is related to the energy flux by

$$E_S^P = 4\pi \langle F^P \rangle^2 \left( \frac{R^P}{F^{gP}} \right)^2 \mathcal{E}_{gP}^* \quad (1)$$

where the  $P$ -wave energy flux,  $\mathcal{E}_{gP}^*$ , is the integral of the square of the ground velocity, taken over the duration of the body-wave arrival,

$$\mathcal{E}_{gP}^* = \rho\alpha \int_0^\infty \dot{u}(t)^2 dt \quad (2)$$

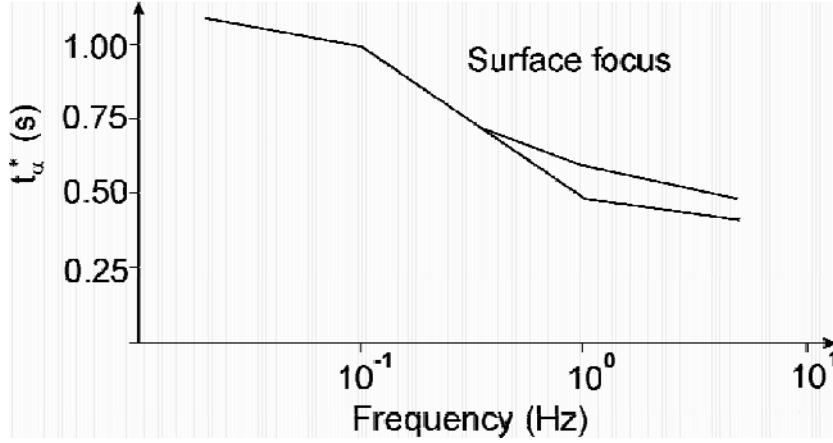
Here,  $\dot{u}$  is velocity, which must be corrected for frequency-dependent attenuation;  $\rho$  and  $\alpha$  are density and velocity at the receiver, respectively;  $\langle F^P \rangle^2$  is the mean-square radiation-pattern coefficient for  $P$  waves;  $R^P$  is the  $P$ -wave geometrical spreading factor;  $F^{gP}$  is the generalized radiation pattern coefficient for the  $P$ -wave group defined as

$$(F^{gP})^2 = (F^P)^2 + (\hat{P}P F^{pP})^2 + \frac{2\alpha q}{3\beta} (\hat{S}P F^{sP})^2 \quad (3)$$

where  $F^i$  are the radiation-pattern coefficients for  $i = P$ ,  $pP$ , and  $sP$ ;  $\hat{P}P$  and  $\hat{S}P$  are plane-wave reflection coefficients of  $pP$  and  $sP$  at the free surface, respectively, corrected for free-surface amplification; and  $q$  is 15.6, the ratio of  $S$ -wave energy to  $P$ -wave energy (Boatwright and Fletcher, 1984). The correction factors explicitly take into account our knowledge that the earthquake is a double-couple, that measurements of the waveforms are affected by interference from depth phases, and that energy is partitioned between  $P$  and  $S$  waves. For teleseismically recorded earthquakes, energy is radiated predominantly in the bandwidth 0.01 to about 5.0 Hz. The wide bandwidth requires a frequency-dependent attenuation correction (Cormier, 1982). The correction is easily realized in the frequency domain by using Parseval's theorem to transform Eq. (2),

$$\mathcal{E}_{gP}^* = \frac{\rho\alpha}{\pi} \int_0^\infty \dot{u}(\omega)^2 e^{\omega t_\alpha^*} d\omega \quad (4)$$

where  $t_\alpha^*$  is proportional to the integral over ray path of the imaginary part of complex slowness in an anelastic Earth. An appropriate operator, valid over the requisite broad bandwidth, is described by Choy and Cormier (1986) and shown in Figure 1. The  $t_\alpha^*$  of the  $P$ -wave operator ranges from 1.0 s at 0.1 Hz to 0.5 s at 2.0 Hz.



**Figure 1** Teleseismic  $t_{\alpha}^*$  derived by Choy and Cormier (1986) plotted as a function of frequency for a surface-focus source and a surface receiver at a distance of  $60^\circ$ . The split in the curve at frequencies higher than 0.3 Hz indicates the variation in regional  $t_{\alpha}^*$  expected for different receiver sites. In practice the mean of the two curves is used for the attenuation correction.

The numerical integration of Eq. (4) is limited to either the frequency at which signal falls below the noise level (typically at frequencies greater than 2.0-3.0 Hz) or to the Nyquist frequency. If this limiting or cutoff frequency,  $\omega_c$ , is greater than the corner frequency, the remainder of the velocity spectrum is approximated by a curve that falls off by  $\omega^{-1}$ . In practice, therefore, Eq. (4) consists of a numerical integral,  $N$ , truncated at  $\omega_c$ , and a residual integral,  $R$ , which approximates the remainder of the integral out to infinite frequency,

$$\mathcal{E}_{gp}^* = \rho \alpha N + \rho \alpha R \quad (5)$$

where, as shown in Boatwright and Choy (1986),

$$R = \frac{\omega_c}{\pi} \left[ \dot{u}_c(\omega_c) \right]^2 \quad (6)$$

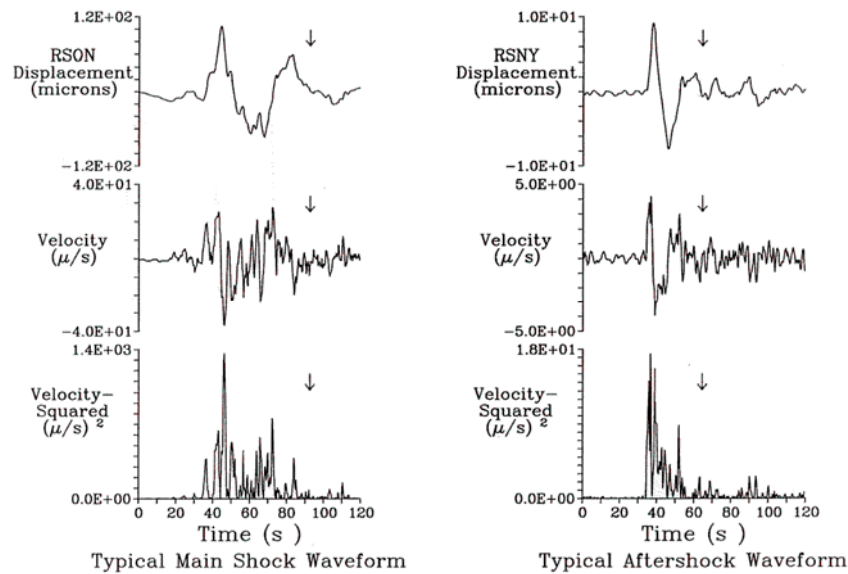
in which  $\dot{u}_c$  is the attenuation-corrected value of velocity at  $\omega_c$ .

Although teleseismic  $SH$ - and  $SV$ -wave groups from shallow earthquakes can be analyzed through a straightforward extension of Eq. (1) as described in Boatwright and Choy (1986), shear waves suffer substantially more attenuation in propagation through the Earth than the  $P$  waves. The loss of seismic signal due to shear attenuation usually precludes retrieving useful spectral information for frequencies higher than about 0.2-0.3 Hz for all but the largest earthquakes. Thus, for the routine estimation of energy, it is more practical and more accurate to use only the  $P$ -wave group (Eqs. (1) and (4)). The formula for computing the total radiated energy when using the  $P$ -wave group alone is

$$E_s = (1 + q) E_s^P. \quad (7)$$

## 2.2 Data

Data used in the direct measurement of energy must satisfy three requirements. First, the implementation of Eq. (4) requires that the velocity data contain spectral information about, above and below the corner frequency of an earthquake. Because the corner frequency can vary from earthquake to earthquake depending on source size and rupture complexity, the bandwidth of the data must be sufficiently wide so that it will always cover the requisite range of frequencies above and below the corner frequency. For body waves from teleseismically recorded earthquakes, a spectral response that is flat to ground velocity between 0.01 Hz through 5.0 Hz is usually sufficient. The second requirement is that the duration of the time window extracted from a seismogram should correspond to the time interval over which the fault is dynamically rupturing. As shown by the examples in Figure 2, when broadband data are used, delimiting the time window is generally unequivocal regardless of the complexity of rupture or the size of the earthquake. The initial arrival of energy is obviously identified with the onset of the direct *P* wave. The radiation of energy becomes negligible when the amplitude of the velocity-squared signal decays to the level of the coda noise. The final requirement is that we use waveforms that are not complicated by triplications, diffractions or significant secondary phase arrivals. This restricts the usable distance range to stations within approximately 30°-90° of the epicenter. In addition, waveforms should not be used if the source duration of the *P*-wave group overlaps a significant secondary phase arrival. For example, this may occur when a very large earthquake generates a *P*-wave group with a duration of such length that it does not decay before the arrival of the *PP*-wave group.



**Figure 2 (Left)** Broadband displacement, velocity, and velocity-squared records for the large ( $M_s = 7.8$ ,  $M_e = 7.5$ ,  $M_w = 7.7$ ) Chilean earthquake of 3 March 1985. Rupture complexity, in the form of a tiny precursor and a number of sub-events, is typical for large earthquakes. **(Right)** Broadband displacement, velocity and velocity-squared records for an aftershock ( $m_b = 5.9$ ,  $M_e = 6.2$ ,  $M_w = 6.6$ ) to the Chilean earthquake that occurred 17 March 1985. The waveforms are less complex than those of the main shock. Despite the differences in rupture complexity, duration and amplitude, the time window over which energy arrives is unequivocal. In each part of the figure the arrows indicate when the velocity-squared amplitude has decreased to the level of the coda noise.

### 3 Development of an energy magnitude, $M_e$

In the Gutenberg-Richter formulation, an energy is constrained once magnitude is known through  $\log E_s = a + b M$  where  $a$  and  $b$  are constants. For surface-wave magnitude,  $M_s$ , the Gutenberg-Richter formula takes the form

$$\log E_s = 4.8 + 1.5 M_s \quad (8)$$

where  $E_s$  is in units of Joules (J). In the normal usage of Eq. (8), an energy is derived after an  $M_s$  is computed. However, it is now recognized that for very large earthquakes or very deep earthquakes, the single frequency used to compute  $M_s$  is not necessarily representative of the dimensions of the earthquake and, therefore, might not be representative of the radiated energy. Since radiated energy can now be computed directly, it is an independent parameter from which a unique magnitude can be defined. In Figure 3, the radiated energies for a set of 378 global shallow earthquakes from Choy and Boatwright (1995) are plotted against their magnitudes,  $M_s$ . The Gutenberg-Richter relationship is plotted as a dashed line in Figure 3. Assuming a  $b$ -value of 1.5, the least-squares regression fit between the actual energies and magnitude is

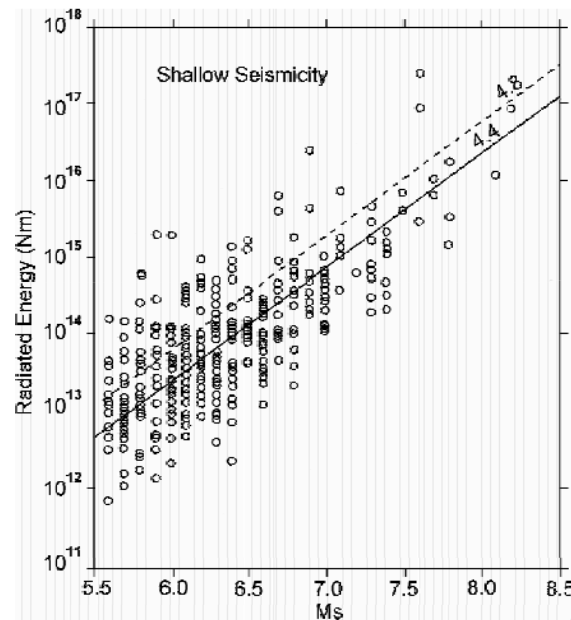
$$\log E_s = 4.4 + 1.5 M_s \quad (9)$$

which is plotted as the solid line in Figure 3. The  $a$ -value of 4.4 indicates that on average the original Gutenberg-Richter formula overestimates the radiated energy by a factor of two. To define energy magnitude,  $M_e$ , we replace  $M_s$  with  $M_e$  in Eq. (9)

$$\log E_s = 4.4 + 1.5 M_e \quad (10)$$

or

$$M_e = (\log E_s - 4.4)/1.5. \quad (11)$$



**Figure 3** Radiated energy ( $E_s$ ) of global data as a function of surface-wave magnitude ( $M_s$ ). The energy predicted by the Gutenberg-Richter formula,  $\log E_s = 4.8 + 1.5 M_s$  (in units of Newton-meters), is shown by the dashed line. From a least-squares regression, the best-fitting line with the slope of 1.5 is  $\log E_s = 4.4 + 1.5 M_s$  (according to Choy and Boatwright, 1995).



The usage of Eq. (11) is conceptually antithetical to that of Eq. (8). In Eq. (11) magnitude is derived explicitly from energy, whereas in Eq. (8) energy is dependent on the value of magnitude.

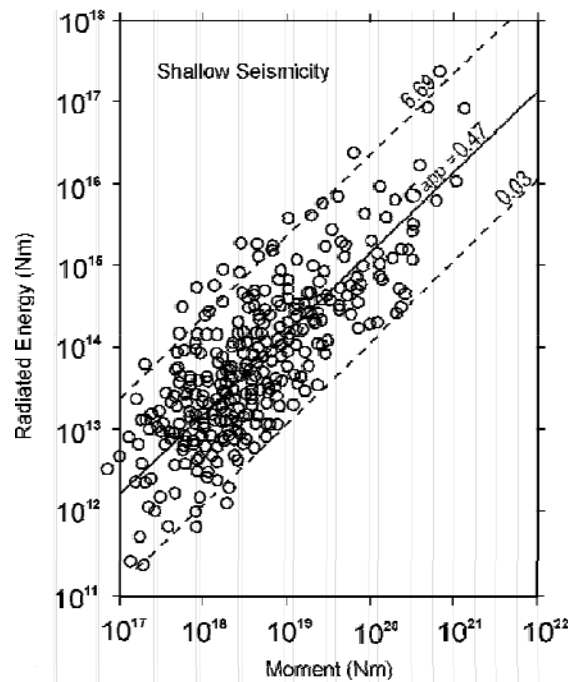
#### 4 The relationship of radiated energy to moment and apparent stress

The energy and moment for a particular earthquake are related by apparent stress,  $\sigma_{app}$  (see Equation (59) in IS 3.1),

$$\sigma_{app} = \mu E_s / M_0 \quad (12)$$

where  $\mu$  is the average rigidity at the source. When radiated energy,  $E_s$ , is plotted against seismic moment,  $M_0$ , for global shallow earthquakes (Figure 4), the best fit by least-squares regression of  $E_s$  on  $M_0$  (solid line) yields

$$E_s = 1.6 \cdot 10^{-5} M_0. \quad (13)$$



**Figure 4** Radiated energy,  $E_s$ , of 394 shallow-focus earthquakes as a function of seismic moment,  $M_0$ . The slope of the least-squares log-normal regression (solid line) yields a global average apparent stress  $\bar{\sigma}_{app}$  of about 0.5 MPa assuming a source rigidity of  $0.3 \cdot 10^5$  MPa. The 95% spread (or width of distribution) about the regression line is indicated by the dashed lines (according to Choy and Boatwright, 1995).

Assuming an average rigidity for shallow earthquakes of  $0.3 \cdot 10^5$  MPa, the slope of the regression line yields a worldwide average apparent stress,  $\bar{\sigma}_{app}$  of about 0.47 MPa. The spread about the regression line is very large. In terms of apparent stress it is between 0.03 to 6.69 MPa. Empirical formulas, like those employing  $M_0$  or  $M_s$ , ignore the spread and, thus,

would be poor predictors of energy. Viewing the spread of  $E_s$ - $M_0$  values about the regression line in terms of apparent stress, rather than random scatter, may provide significant insight into the physics of earthquake occurrence. For example, the release of energy and apparent stress could vary systematically as a function of faulting type, lithospheric strength and tectonic region (Choy and Boatwright, 1995). As more statistics on the release of energy are accumulated, spatial and temporal variations in energy release and apparent stress might also be identified.

## 5 The relationship of $M_e$ to $M_w$

Although  $M_e$  and  $M_w$  are magnitudes that describe the size of an earthquake, they are not equivalent.  $M_e$ , being derived from velocity power spectra, is a measure of the radiated energy in form of seismic waves and thus of the seismic potential for damage to anthropogenic structures.  $M_w$ , being derived from the low-frequency asymptote of displacement spectra, is physically related to the final static displacement of an earthquake. Because they measure different physical properties of an earthquake, there is no *a priori* reason that they should be numerically equal for any given seismic event. The usual definition of  $M_w$  is:

$$M_w = 2/3 \log M_0 - 6.0 \quad (\text{with } M_0 \text{ in Nm}). \quad (14)$$

The condition under which  $M_e$  is equal to  $M_w$ , found by equating Eq. (11) and Eq. (14), is  $E_s/M_0 \sim 2.2 \cdot 10^{-5}$ . From Eq. (12) this ratio is equivalent to  $\sigma_{app} \sim 2.2 \cdot 10^{-5} \mu$ . For shallow earthquakes, where  $\mu \sim 0.3\text{--}0.6 \times 10^5$  MPa, this condition implies that  $M_e$  and  $M_w$  will be coincident only for earthquakes with apparent stresses in the range 0.66–1.32 MPa. As seen in Figure 4, this range is but a tiny fraction of the spread of apparent stresses found for earthquakes. Therefore, the energy magnitude,  $M_e$ , is an essential complement to moment magnitude,  $M_w$ , for describing the size of an earthquake. How different these two magnitudes may be is illustrated in Table 1. Two earthquakes occurred in Chile within months of each other and their epicenters were less than  $1^\circ$  apart. Although their  $M_w$ 's and  $M_s$ 's were similar, their  $m_b$ 's and  $M_e$ 's differed by 1 to 1.4 magnitude units! Table 1 describes the macroseismic effects from the two earthquakes. The event with larger  $M_e$  caused significantly greater damage!

**Table 1** (Reprinted from Choy et al., 2001.)

| Date               | LAT<br>(°) | LON<br>(E) | Depth<br>(km) | $M_e$ | $M_w$ | $m_b$ | $M_s$ | $\sigma_{app}$<br>(bars) | Faulting Type     |
|--------------------|------------|------------|---------------|-------|-------|-------|-------|--------------------------|-------------------|
| 6 JUL<br>1997 (1)  | -30.06     | -71.87     | 23.0          | 6.1   | 6.9   | 5.8   | 6.5   | 1                        | interplate-thrust |
| 15 OCT<br>1997 (2) | -30.93     | -71.22     | 58.0          | 7.5   | 7.1   | 6.8   | 6.8   | 44                       | intraslab-normal  |

(1) Felt (III) at Coquimbo, La Serena, Ovalle and Vicuna.  
 (2) Five people killed at Pueblo Nuevo, one person killed at Coquimbo, one person killed at La Chimba and another died of a heart attack at Punitaqui. More than 300 people injured, 5,000 houses destroyed, 5,700 houses severely damaged, another 10,000 houses slightly damaged, numerous power and telephone outages, landslides and rockslides in the epicentral region. Some damage (VII) at La Serena and (VI) at Ovalle. Felt (VI) at Alto del Carmen and Illapel; (V) at Copiapo, Huasco, San Antonio, Santiago and Vallenar; (IV) at Caldera, Chanaral, Rancagua and Tierra Amarilla; (III) at Talca; (II) at Concepcion and Taltal. Felt as far south as Valdivia. Felt (V) in Mendoza and San Juan Provinces, Argentina. Felt in Buenos Aires, Catamarca, Cordoba, Distrito Federal and La Rioja Provinces, Argentina. Also felt in parts of Bolivia and Peru.

## 6 Regional estimates of radiated seismic energy

Radiated energy from local and regional records can be computed in a fashion analogous to the teleseismic approach if suitable attenuation corrections, local site and receiver effects, and hypocentral information are available or can be derived. Boatwright and Fletcher (1984) demonstrated that integrated ground velocity from S waves could be used to estimate radiated energy in either the time or frequency domain by,

$$E_s = 4\pi C^2 r^2 \rho_r \beta_r \int_0^\infty \dot{u}_c(t)^2 dt \quad (15)$$

$$= 4\pi C^2 r^2 \rho_r \beta_r \int_0^\infty \dot{u}_c(\omega)^2 d\omega \quad (16)$$

where the ground velocity has been corrected for anelastic attenuation,  $C$  is a correction for radiation pattern coefficient and free-surface amplification,  $r$  is the source-receiver distance, and  $\rho_r$  and  $\beta_r$  are density and S-wave velocity at the receiver. The attenuation correction is usually of the type  $\exp(\omega r/\beta Q)$ , where  $Q$  is the whole-path attenuation. Similarly, Kanamori et al. (1993) use a time-domain method to estimate the S-wave energy radiated by large earthquakes in southern California,

$$E_\beta = 4\pi r^2 C_f^{-2} [r_0 q(r_0)/r q(r)]^2 \rho_0 \beta_0 \int_0^\infty \dot{u}^2(t) dt \quad (3.66)$$

where  $\rho_0$  and  $\beta_0$  are hypocentral density and S-wave velocity,  $C_f$  is the free-surface amplification factor,  $r$  is the source-receiver distance estimated from the epicentral distance  $\Delta$  and a reference depth  $h$  of 8 km (such that  $r^2 = \Delta^2 + h^2$ ). Attenuation is described by  $q(r) = cr^{-n} \exp(-kr)$ , which is the Richter (1935) attenuation curve as corrected by Jennings and Kanamori (1983). For southern California earthquakes,  $c=0.49710$ ,  $n=1.0322$ , and  $k=0.0035 \text{ km}^{-1}$ .

## 7 Conclusions

Energy gives a physically different measure of earthquake size than moment. Energy is derived from the velocity power spectra, while moment is derived from the low-frequency asymptote of the displacement spectra. Thus, energy is a better measure of the severity of shaking and thus of the seismic potential for damage, while moment, being related to the final static displacement, is more related to the long-term tectonic effects of the earthquake process. Systematic variations in the release of energy and apparent stress as a function of faulting type and tectonic setting can now be identified that were previously undetectable because of the lack of reliable energy estimates. An energy magnitude,  $M_e$ , derived from an explicit computation of energy, can complement  $M_w$  and  $M_s$  in evaluating seismic and tsunamigenic potential.

## Recommended overview readings

Bormann and Di Giacomo (2011)  
 Di Giacomo and Bormann (2011)  
 Choy and Boatwright (1995)  
 Choy et al. (2006)  
 Choy (2012)

## References

- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Boatwright, J., and Fletcher, J. B. (1984) The partition of radiated energy between P and S waves, *Bull. Seism. Soc. Am.*, **74**, 361-376.
- Boatwright, J., and Choy, G. (1986). Teleseismic estimates of the energy radiated by shallow earthquakes. *J. Geophys. Res.*, **91**, B2, 2095-2112.
- Choy, G. L. (2012). IS 3.5: Stress conditions inferable from modern magnitudes: development of a model of fault maturity. In: Bormann, P. (Ed.) (2012). *New Manual of Seismological Observatory Practice (NMSOP-2)*, IASPEI, GFZ German Research Centre for Geosciences, Potsdam, 10 pp.; DOI: 10.2312/GFZ.NMSOP-2\_IS\_3.5.
- Choy, G. L., and Cormier, V. F. (1986). Direct measurement of the mantle attenuation operator from broadband P and S Waveforms. *J. Geophys. Res.*, **91**, 7326-7342
- Choy, G. L., and Boatwright, J. L. (1995). Global patterns of radiated seismic energy and apparent stress. *J. Geophys. Res.*, **100**, B9, 18,205-18,228.
- Choy, G. L., and Kirby, S. H. (2004). Apparent stress, fault maturity and seismic hazard for normal-fault earthquakes at subduction zones, *Geophys. J. Int.*, **159**, 991-1012.
- Choy, G. L., McGarr, A., Kirby, S. H., and Boatwright, J. (2006). An overview of the global variability in radiated energy and apparent stress; in: Abercrombie R. , McGarr, A., and Kanamori, H. (eds): Radiated energy and the physics of earthquake faulting, *AGU Geophys. Monogr. Ser.* **170**, 43-57.
- Cormier, V. F. (1982). The effect of attenuation on seismic body waves. *Bull. Seism. Soc. Am.*, **72**, 1, 169-200.
- Di Giacomo, D., and Bormann, P. (2011). Earthquake energy. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 233-236; doi: 10.1007/978-90-481-8702-7.
- Jennings, P. C., and Kanamori, H. (1983). Effect of distance on local magnitudes found from strong motion records. *Bull. Seism. Soc. Am.*, **73**, 265-280.
- Kanamori, H., Mori, J., Hauksson, E., Heaton, Th. H., Hutton, L. K., and Jones, L. M. (1993). Determination of earthquake energy release and  $M_L$  using TERRASCOPE. *Bull. Seism. Soc. Am.*, **83**, 2, 330-346.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bull. Seism. Soc. Am.*, **25**, 1-32.

|                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|----------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Topic</b>   | <b>The Russian K-class system, its relationships to magnitudes and its potential for future development and application</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>Authors</b> | <p><b>Peter Bormann</b>, formerly GFZ German Research Centre for Geosciences, D-14473 Potsdam, Germany; now E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> <p><b>Kazuya Fujita</b> and <b>Kevin G. Mackey</b>, Department of Geological Sciences, Michigan State University, East Lansing, MI 48824-1115, USA; E-mail: <a href="mailto:fujita@msu.edu">fujita@msu.edu</a> and <a href="mailto:mackeyke@msu.edu">mackeyke@msu.edu</a></p> <p><b>Alexander Gusev</b>, Institute of Volcanology and Seismology, Piip Blvd. 9, Petropavlovsk-Kamchatsky, 683006, Russia; E-mail: <a href="mailto:gusev@emsd.ru">gusev@emsd.ru</a></p> |
| <b>Version</b> | July 2012; DOI: 10.2312/GFZ.NMSOP-2_IS_3.7                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |

|                                                                                       | <b>page</b> |
|---------------------------------------------------------------------------------------|-------------|
| <b>1 Abstract</b>                                                                     | <b>1</b>    |
| <b>2 The development of the K-class system and methodology</b>                        | <b>2</b>    |
| <b>3 Regional variations of the K-class method</b>                                    | <b>7</b>    |
| <b>4 Relationship of K class to magnitude scales</b>                                  | <b>11</b>   |
| <b>5 Future development and use of K for energy related earthquake classification</b> | <b>19</b>   |
| <b>Acknowledgments</b>                                                                | <b>24</b>   |
| <b>References</b>                                                                     | <b>24</b>   |

## 1 Abstract

The paper aims at reviewing the development and use of the energy class (K-class) system since the late 1950s in the former Soviet Union (FSU) for quantifying the size of local and regional earthquakes, at comparing its results with current full-fledged estimations of released seismic energy and energy magnitude via the integration of teleseismic broadband velocity P-waveforms and at assessing the potential of modernized and standardized K-class determinations as a valuable complement to current teleseismic procedures. K-class in its original preliminary version was proposed by Victor Bune. His version was radically improved by Tatiana Rautian to constitute a rapid and simple means of estimating the radiated seismic energy  $E_s$  from an earthquake. K-class was defined as  $K = \log_{10} E_s$  with  $E_s$  in units of Joules. In practice, a graphical method was used based on the maximum horizontal (for S-wave) and vertical (for P-wave) amplitudes which was calibrated to independently determined energy estimates in Soviet Central Asia, and was used directly or with modifications as a standard throughout the FSU. The empirical relationships between K and classical magnitude scales vary somewhat between areas, but are, on the whole, consistent with magnitude-energy relationships proposed in the past fifty years, as well as with energy magnitude  $M_e$ . We show that the concept of the K-class can continue to be applied, with modifications, through the use of modern recording, filtering and data analysis tools and applied to regions outside the FSU as well as to the extensive historical data set from the FSU.

## 2 The development of the *K*-class system and methodology

After the devastating magnitude 7.4 Khait earthquake of 1949, the Complex (Interdisciplinary) Seismological Expedition was deployed by the former Geophysical Institute of the Soviet Union (USSR) in the Garm region of Tadjikistan to study its aftershocks and related phenomena. No specifications for seismic stations, procedures for measurements, standards for data processing, or necessary documentation had been established at that time. Thus, members of the expedition developed their own procedures. Due to the large number of events and limited computational and data analysis tools available, relatively simple, primarily graphical, methods were required.

In its original preliminary version, "Energy class K" was proposed by Bune (1955) with the intention to create a tool for absolute source energy calibration. It was based on integration of squared trace and theoretically would be a big advance with respect to the determination of magnitude *M*. Yet, in practice, it was unworkable with its hand-made integration on band-limited photographic records; it also used an imperfect way to reduce recorded wave energy to *E<sub>s</sub>*. To make the *K* scale workable in mass processing, Rautian (1958, 1960) made a critical step by returning to the measurement of peak-amplitude instead of integrating squared velocity amplitudes, thus developing a method to estimate earthquake energy that could be implemented by technical staff with only limited training, as summarized below. To calibrate her scale, Rautian performed hand integration of trace and *A/T* measurements in parallel for a large training data set. She also radically improved the source energy estimation procedure. A more detailed description of its development and calibration is presented in Rautian et al. (2007).

If seismic energy (*E<sub>s</sub>*) radiates uniformly, then at very short distances in a homogeneous Earth,  $E_s = 4\pi r^2 k\varepsilon$ ; where  $\varepsilon$  is the total energy density that crosses normal to a surface of a circular wave front with radius, *r*, per unit area, and *k* is a coefficient which accounts for a diversity of effects such as the Earth's surface topography and the ratio of the measured component to the full displacement vector. Since the short-period seismographs deployed in early years in and around Garm recorded displacement both amplitude and frequency would need to be measured to estimate  $\varepsilon$ ; however, visual "spectral analysis" of a seismogram by technicians was impractical and resulted in considerable scatter.

Both the terminology and the analysis of waves used in developing the *K*-class system have some ambiguity as they were selected more for operational purposes. *A* stands for the displacement amplitude of a specific oscillation, in recent usage generally the first arrival, and *A<sub>max</sub>* for the largest amplitude in the whole wave group belonging to a specific type of seismic phase. *T* refers to the period measured for a specific pulse with amplitude *A* or *A<sub>max</sub>*, whichever is being referred to, within the duration,  $\tau$ , of an arrival as measured from approximately the start to the end of oscillations with  $A > A_{max}/2$ . In practice, *A<sub>max</sub>* was, and is still, used in calculating and calibrating *K*.

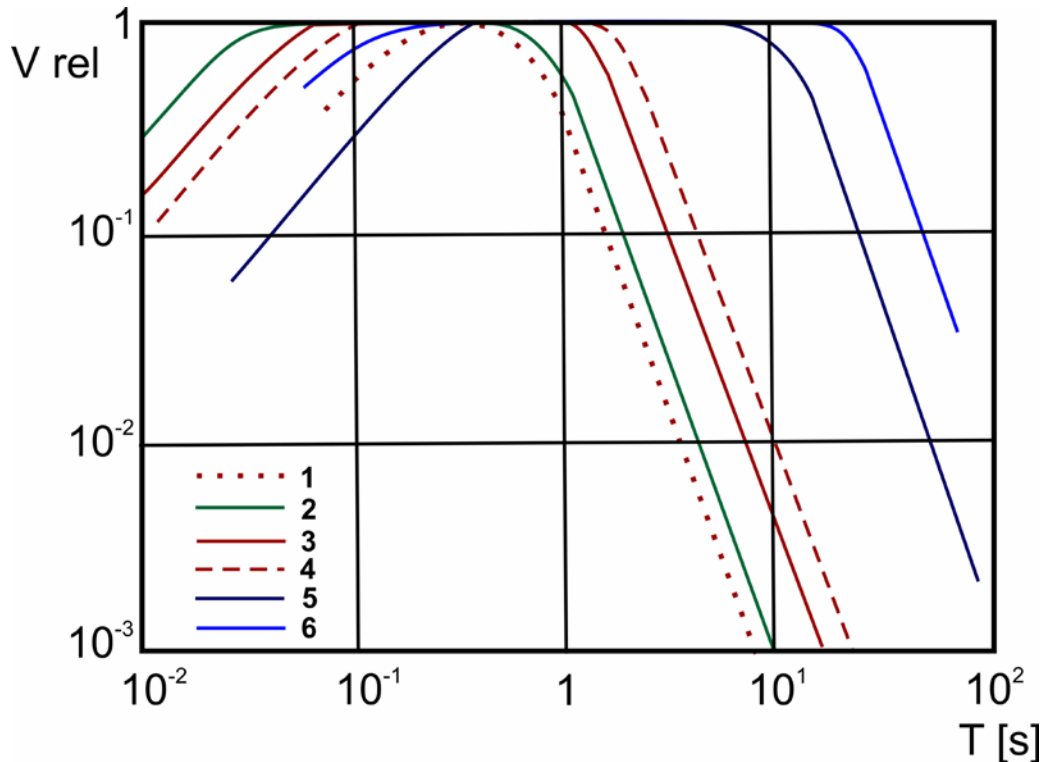
Generally,  $\varepsilon$  depends on the amplitudes, frequency content, and signal duration of the analyzed seismic phases; however, observations at Garm showed that signal duration was controlled primarily by epicentral distance (due to the multiple scattering of waves in the structurally complex Earth medium in this mountainous area), whereas the frequency content in a wave group was largely controlled by the energy released by the earthquake and by its hypocentral distance (Rautian, 1960), due to the energy and stress-drop dependent shift of the corner frequency of the radiated source spectrum on the one hand, and frequency-dependent

attenuation on the other hand (see Sections 3). Thus, by measuring amplitude, which was easy to measure, alone, it was found that energy could be estimated to within a factor of 2 to 3, with proper regional calibration (i.e., by accounting in a rough manner for the combined impact of diverse parameters that influence the measured values). Moreover, to reduce the scatter and smooth the effect of the source radiation pattern, the sum of the **maximum** amplitude of the P wave on the vertical component ( $A_P$ ) and of the greater S-wave on one of the horizontal components ( $A_S$ ) was chosen, with P being Pg and S being Sg at local and shorter regional distances (usually < 400 km), whereas at larger regional distances, the measured arrivals could be Pg and Lg. This yielded a pragmatic solution. Operationally, especially in the continental regions of eastern Russia, Pg is the commonly measured P phase with very few Pn amplitudes ever picked. At distances of greater than 700-800 km (or less in central Asia and Baikal), Sg (Lg) is often the only amplitude measured.

To calibrate the amplitude measurements to energy, Rautian (1960) estimated the energy of a large number of earthquakes, predominantly from the northern Garm region, by visually measuring frequencies and amplitudes on particularly clear records, and using many other simplifying assumptions. In brief, the energy density,  $\varepsilon$ , was determined by summing all visually separable  $(Af)^2$ , which is proportional to the ground motion velocity squared, over the measured pulse duration  $\tau$ , where  $f$  is the measured frequency of a given cycle with displacement amplitude  $A$ . A correction,  $k$ , to account for frequency-dependent magnification of the seismograph, the use of only one record component instead of the complete vector, surface effects, etc., was estimated and applied. The dependence of  $k\varepsilon$  on distance,  $r$ , was observationally determined and then used to normalize the estimate of  $\varepsilon$  to  $r = 10$  km. Then,  $\log E_S(10 \text{ km}) = \log 4\pi k\varepsilon$ . Further, the amplitude measurements,  $[A_P(r) + A_S(r)]$ , were also normalized to  $r = 10$  km using an amplitude-distance relationship.

The measurements were made on short-period instruments, initially of type VEGIK, with a seismometer eigenperiod  $T_s = 0.8$  s, in common use in the 1950s and 1960s, later amended or replaced by Kirnos seismographs of type SKM and SK-III with slightly longer eigenperiod ( $T_s = 1.5$  s and 2 s, respectively), followed by the more medium to long-period Kirnos seismographs of type SK ( $T_s = 10$  s) and SKD ( $T_s = 20$ s). The approximate average displacement response curves of all these seismographs, compared with the response of the short-period Benioff seismographs in common use in the United States in these years, all normalized to a maximum magnification of 1, are depicted in Figure 1.

Amplitude measurements were typically made at periods < 3 s, i.e. for crustal S waves with a wavelength typically less than 10 km. As stated above, the K scale was calibrated to the energy flow through a focal sphere of radius  $r = 10$  km only. Such a sphere will, on the average, only enclose earthquake ruptures up to a magnitude of about 6. Therefore, the K-classification is principally suitable only for scaling local and regional earthquakes up to magnitudes of about 6 to 6.5 or  $K \leq 14$  to 15 (see sections 4 and 5 on magnitude-K relationships). For larger earthquakes which extend far beyond the size of the reference sphere and which release their maximum of seismic energy at periods  $\gg 3$  s the amount of released seismic energy will necessarily be underestimated (see Chapter 3, average seismic source spectrum and “saturation effect”).



**Figure 1** Normalized average displacement response curves of typical classical Soviet seismographs in comparison with the short-period US Benioff seismograph: **1** – Benioff; **2** – VEGIK with  $T_s = 0.8$  s; **3** – SKM with  $T_s = 1.5$  s; **4** – SK-III with  $T_s = 2$  s; **5** – SK with  $T = 10$  s and **6** – SKD with  $T_s = 20$  s.

For amplitude measurements on VEGIK records, the following empirical calibration formula was derived:

$$\log E_s \text{ (at 10 km in Joules)} = 1.8 \log [A_P(10 \text{ km}) + A_S(10 \text{ km})] + 6.4. \quad (1)$$

In terms of trivial physics, formula (1) looks a bit strange, because in the case of wave propagation in a homogeneous medium one would expect  $\log E_s \approx 2 \log A_{rms} + \log(\text{duration}) + \text{const.}$  But, according to formula (1), the waveform duration would then be  $\propto A^{-0.2}$ , i.e., it would decrease with amplitude, contrary to average observation. Yet, empirically, in real Earth, deviation from an ideal coefficient 2 may be correct and have also been found in other empirical magnitude-energy relationships, e.g., the relationship published by Gutenberg and Richter (1956) between  $\log E_s$  and body-wave magnitude  $m$  (with coefficient 2.2), and the one given by Richter (1958) between  $\log E_s$  and surface-wave magnitude  $M_s$  (with 1.5).

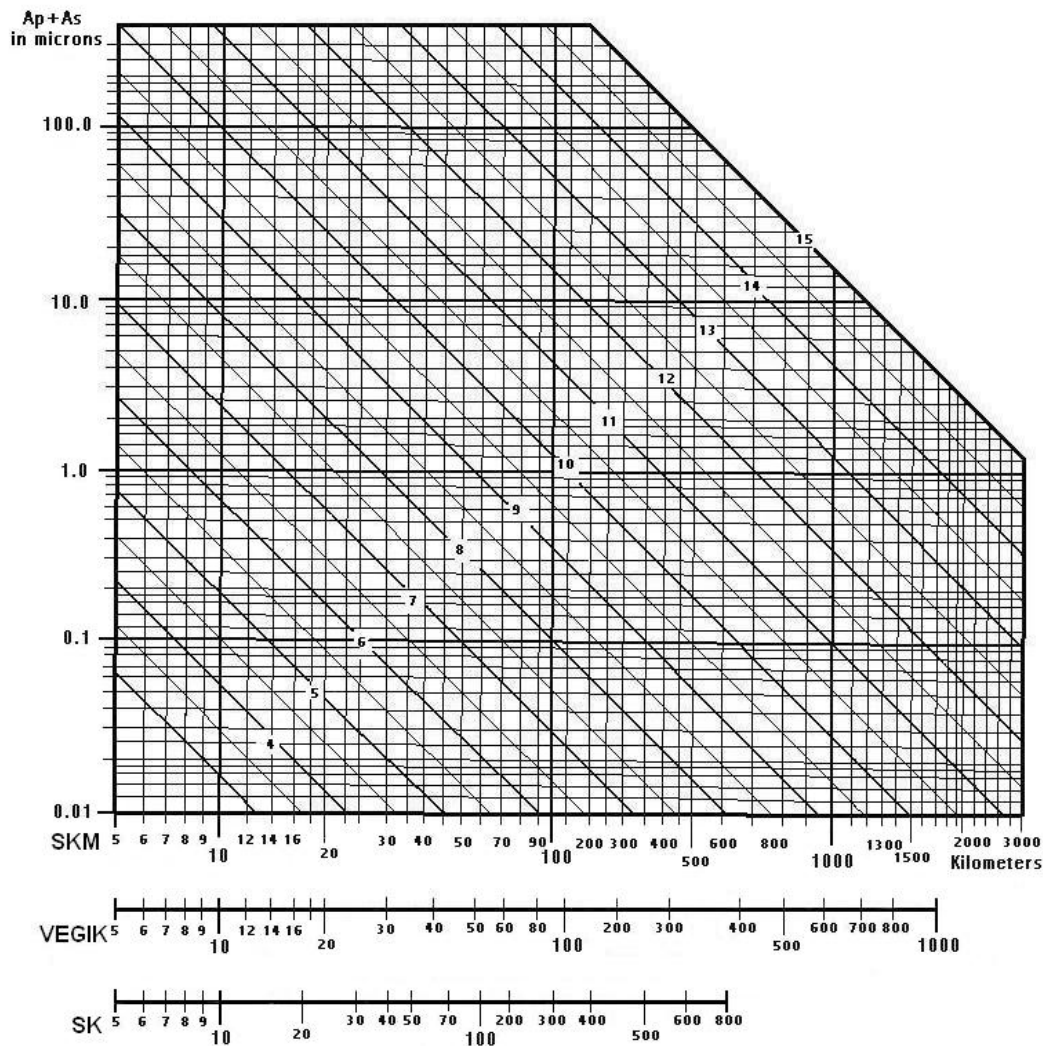
Moreover, also jumping back from considering in the original design of the scale  $\varepsilon$ , i.e., the total energy density crossing the circular wave front with radius  $r$  per unit area, to using the peak amplitude recorded at seismic stations instead, is significant and conceptually problematic. But in practical terms this step was unavoidable with hand measurements and, therefore, does not deserve any criticism.

Formula (1) yields an amplitude sum,  $A_P(10 \text{ km}) + A_S(10 \text{ km})$ , of 100 microns for  $\log E_s$  of 10. The shape of this calibration curve with distance was similar to that of Richter's (1935) local magnitude  $M_L$  but included deviations that were later explained as possibly being due to



refractions from the Moho or other arrivals, as well as to changing spectral content of waves with distance.

For simplicity in practical use, the log-distance scale was adjusted to make the amplitude-distance curves straight lines, as in Figure 2. Thus while the distance axis appears to be logarithmic, close examination of it shows that the horizontal scale was adjusted to incorporate the deviations.



**Figure 2** The Rautian (1964) K-class nomogram calibrated for SKM, VEGIK, and SK seismometers. Note that the horizontal distance scales are not truly logarithmic; see text for explanation. Vertical axis: sum of P and S-wave amplitudes in microns (modified from Figure 2 in Rautian et al., 2007, *Seism. Res. Lett.*, **78**(6), p. 582, © Seismological Society of America).

Since the calibration to  $E_s$  depends on the response of the seismograph, the calibration curves should have been adjusted as new instruments with different responses were deployed. However, the derivation of new calibration formulas for K for newly introduced instrumentation would have required a tremendous amount of empirical and theoretical work.

Moreover, the most commonly measured frequencies of local earthquakes (2-5 Hz) were such that the differences in instrument response did not appear to significantly change the results. Therefore, the same nomogram was used and only the distance scales were adjusted for different instruments (Rautian, 1964). Figure 2 shows the respective nomograms for calibrating the sum of P and S amplitude readings on short-period Kirnos SKM and VEGIK as well as on medium-period SK records, which continue to be used until now. The mismatch between the K-scales for different instruments reflects the fact that their passbands probe different parts of the source spectrum.

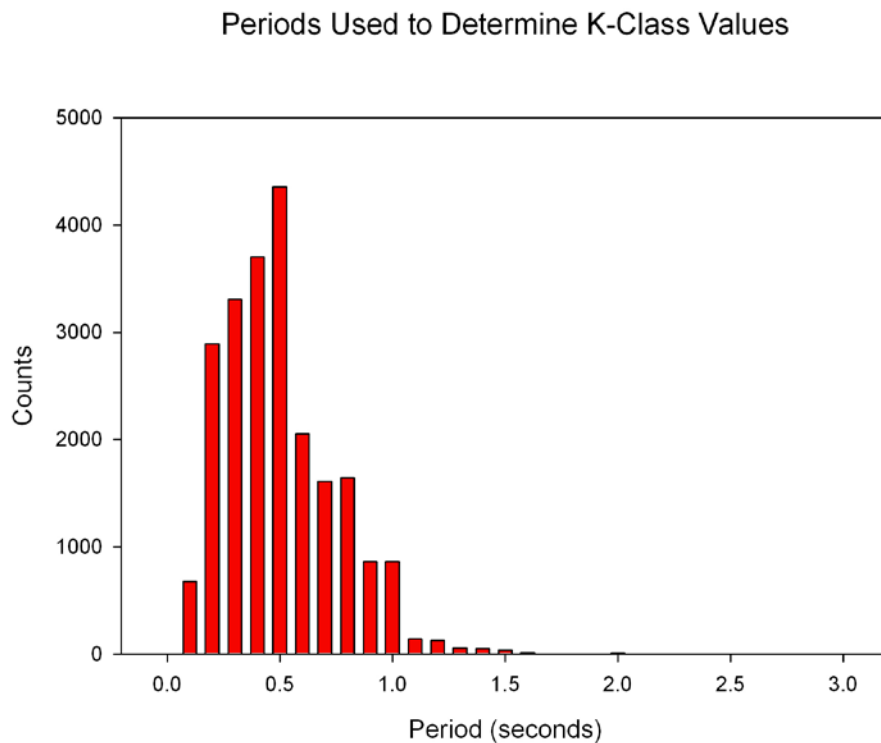
SKM seismographs became the main instrument for K-determination. They had typically a seismometer eigenperiod,  $T_s$ , around 1 to 2 s, a galvanometer period,  $T_g$ , around 0.3 s, and a damping of the seismometer and galvanometer chosen so as to produce a displacement-proportional response in the frequency range between some 0.5 and 20 Hz (see Figure 1).

Today, digital broadband seismograms could easily be accommodated by filtering such records to simulate the records with instrument responses of either the VEGIK, SKM or SK instruments. On Kamchatka, this is now routine procedure, i.e., converting BB digital velocity traces to 1.2s-instrument galvanometric traces of pre-digital era, and calculation of K values from S-waves. We are unaware, however, whether digital recordings in the FSU are nowadays routinely filtered everywhere before measuring parameters such as K – to emulate one of the historic instruments so as to remain compatible with the original definitions and baselines.

Because of the many simplifying assumptions made and uncertainties involved, the scale was called “energy class” instead of “energy” (see also below). Using the Russian spelling класс, the scale was defined as  $K = \log E_s$ , with  $E_s$  in Joules. The first outline of this scale was presented in Rautian (1958) and updated in Rautian (1960). The scale had its problems, but was easy to apply with the facilities and personnel of the time. K-class was adopted to quantify the size of earthquakes throughout the former USSR by 1961 and continues to be used today. Outside the USSR, K-class was used extensively only in Mongolia and in Cuba..

The standard deviation in K-class estimates between stations, due to site effects, station distribution with respect to the radiation pattern, heterogeneities in the crust, using only one horizontal component, etc., was empirically determined to be about 0.35 log units. No site corrections were originally used; however, they were later added by some networks.

In current practice, networks use only the maximum horizontal displacement amplitudes for K determination, usually at hypocentral distances up to about 600 km (see Figures. 2 and 4 to 6), since the amplitude of  $S_g$  (or  $L_g$ ) dominates that of  $P_g$ .  $S_g$  and  $L_g$  are not differentiated and simply referred to as  $S_g$ , even at distances up to 1500 km where  $L_g$  is often still observed on the craton. These arrivals typically have dominant periods below 2.5 s. Operationally, the typical range of periods measured for determining K-class values today is 0.2 to 1.1 s. (see Figure 3). P-arrivals are seldom used in continental eastern Russia; S is used in place of  $S_g$  in the Kuriles and S and P are independently used on Kamchatka.



**Figure 3** Average histogram of periods at which P and/or S-wave amplitudes have been measured by different local networks in the FSU for estimating the K values of mostly local earthquakes between 1978 and 2009.

In summary, K estimates, being based on short-period P- and S-wave amplitudes with periods less than 2 s, and wavelengths typically less than 10 km, are suitable only for scaling local and regional earthquakes with magnitudes up to 5.5 or 6 at most, corresponding to  $K \approx 14$  to 15. For larger earthquakes, the amount of released seismic energy will necessarily be underestimated (“saturation”).

### 3 Regional variations of the K-class method

K-class (or more correct, K-magnitude) was adjusted regionally in the FSU to significant degrees. This is indispensable, as for other short-period local/regional magnitude scales applicable in different seismo-tectonic environments (e.g., ML and mb\_Lg; see IS 3.3 and Table 2 in DS 3.1).

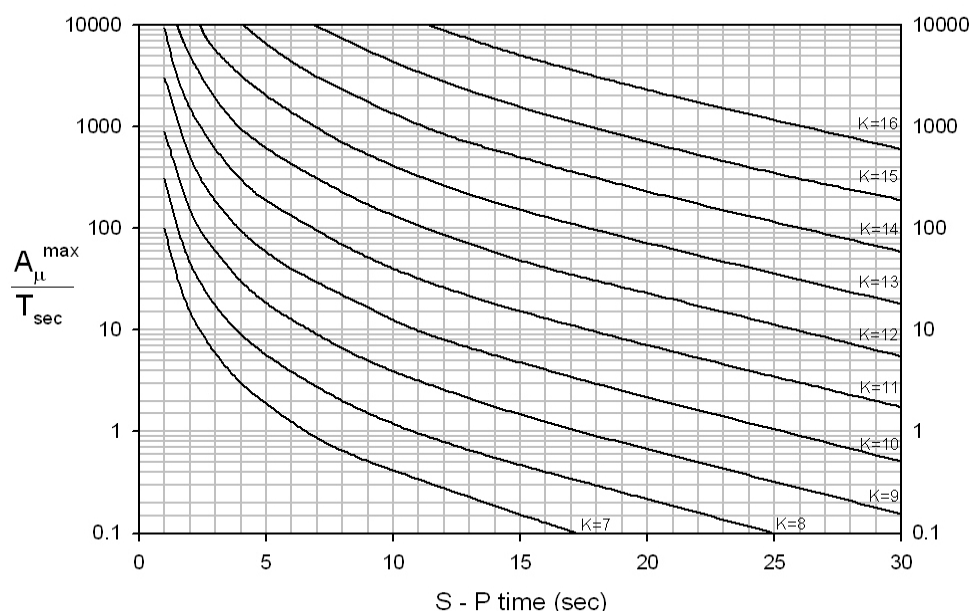
As noted, the routine K-class calculations ignored the specific spectral content of the individual displacement records, which was observed to differ randomly between events in the same area, and systematically between events in different tectonics settings. Thus, when calculated from displacement only, K is more closely related to the logarithm of the seismic moment,  $M_0$ , than to  $\log E_S$  as calculated by using velocity spectra.

The ratio between these two parameters may vary regionally, depending on the seismotectonic environments such as continental crust, subduction zones, deep seismic zones, oceanic ridges, etc. and thus also on the predominant type of source mechanism (e.g., Choy and Boatwright, 1995; Choy et al., 2006; Kanamori and Brodski, 2004; Bormann and Di

Giacomo, 2011). On a global scale the variability reaches about 3 orders of magnitude. It is mainly related to the differences in stress drop, or “apparent stress” (Wyss and Brune, 1968), but also to variable rigidity in the yielding earthquake source volume (e.g., Houston, 1999; Polet and Kanamori, 2000) and thus to “fault maturity” (see IS 3.5). In this context one should note that both stress drop and rigidity are considered to be constant in  $M_0$  calculations.

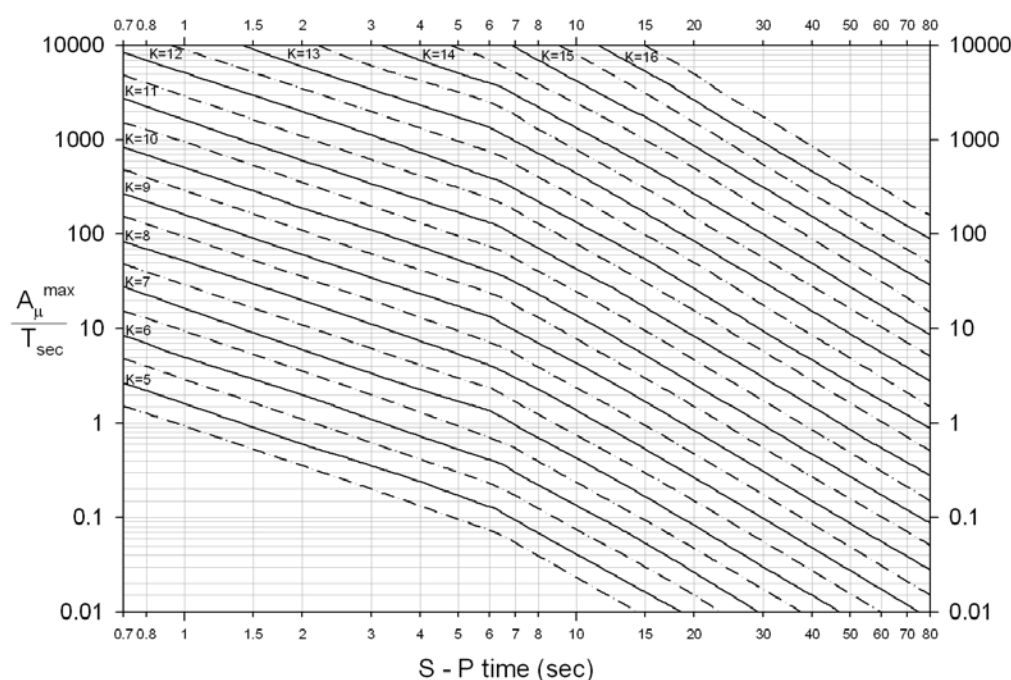
Another useful related parameter is “earthquake hardness  $S_M$ ”. This term was originally introduced by N.V. Shebalin to describe the deviation of macroseismic intensity (see Chapter 12) from its average value at a given  $M_s$ , which are closely related to differences in the high-frequency content of the seismic wave field in the shaken areas. Thus,  $S_M$  is quite relevant for describing variations of the  $E_s/M_0$  ratio too. The correspondence between  $K$  and  $M_0$  was found on FSU territory to be best in crystalline rocks and regions of thrust tectonics ( $S_M \approx 0$ ) and worst (significantly negative  $S_M$ ) for strike-slip events within strongly fractured rocks with low-frequency source spectra.  $S_M$  was found to be significantly positive for intermediate-depth events.

With respect to the Kurile Islands, Fedotov used in 1963 S-wave amplitudes only and replaced the proper distance scale (as used, e.g., in Rautian, 1964) by S-P time (as in Figure 4). His scale was usable for earthquakes in a wide depth range of the Kurile Islands environment. The K-Fedotov63 was improved in 1968, and additionally an analog P-wave K-scale was proposed. Both Fedotov K-scales are still in use on Kamchatka, but their calibration curves and also the way of their absolute calibration are radically different from the Rautian K scale (for details see Fedotov, 1972). This applies, more or less, to other regional K-class formulas as well. Note that on Kamchatka it is routine practice to use besides the two Fedotov K scales also a coda-class Kc. The latter is very efficient when S-wave amplitudes are off scale on photographic records, or out of the dynamic range on a digital channel (see Lemzikov and Gusev, 1991).



**Figure 4** The 1967 Kurile nomogram for calibrating  $A/T$  with  $A$  measured in micron ( $\mu$ ) and  $T$  in seconds on SK records. The horizontal axis is linear in differential S-P travel time, which is proportional to hypocentral distance, in this diagram up to about 250 km (copied from Figure 5 in Rautian et al., 2007, *Seism. Res. Lett.*, **78**(6), p. 585, © Seismological Society of America).

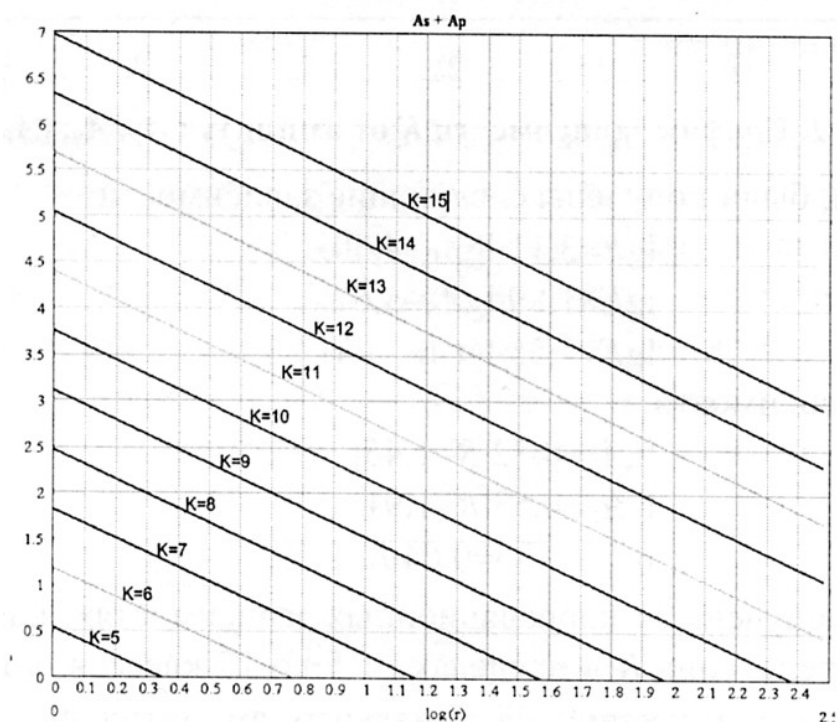
The Kamchatka and Sakhalin regional networks are both located in subduction zone settings with very different velocity structure,  $L_g$  propagation, and a stronger attenuation, than in the continental regions. Thus, they developed their own  $K$  scales (see Rautian et al., 2007) based on using only  $A_{Smax}$ , but normalized by the period,  $T$ , and thus being a measure of the maximum S-wave ground motion velocity. The calibration techniques were similar and different calibration relationships were developed for shallow and for intermediate depth earthquakes. The Sakhalin network nomogram, used in the Kurile Islands for scaling SK records (see Figure 4), and Kamchatka nomograms for scaling SKM records (see Figure 5) were revised several times as both networks were expanded and more data was acquired. In the compilation by Kondorskaya and Shebalin (1977), the 1961 scale (Fedotov et al., 1964;  $K_F$ ) and 1967 scale (Solov'ev and Solov'eva, 1967;  $K_S$ ) were converted to Rautian (1960)  $K$ -class values using  $K = K_F + 0.6$  and  $K = K_S + 1.7$ .



**Figure 5** The current Kamchatka nomogram for scaling  $A/T$  measured on SKM records. Dashed lines represent limits of integer determinations. Vertical axis amplitude is in microns ( $\mu$ ), period,  $T$ , in seconds. The horizontal axis is logarithmic in S-P travel time, and corresponds to hypocentral distances up to about 640 km (copied from Figure 6 in Rautian et al., 2007, Seism. Res. Lett., **78**(6), p. 585, © Seismological Society of America).

A  $K$ -class nomogram was also developed for the Crimea (Pustovitenko and Kul'chitsky, 1974), although it is unclear how extensively it was used. Recently, Adilov et al. (2010) published  $K$ -class nomograms for Dagestan utilizing  $A_P$  only,  $A_S$  only, and  $A_P + A_S$  (Figure 6), using data from 1994-1997. As elsewhere,  $A_S$  dominates and therefore a reasonable approximation can be obtained from  $A_S$  data alone, especially at lower  $K$  values. According to Kondorskaya et al. (1981) the definition of  $K$  on the nomograms gives satisfactory results to within  $\pm 0.5 K$ , which corresponds to about 0.3 m.u. (magnitude units).

The Baikal (Solonenko, 1977) and Caucasus (Dzhibladze, 1971) networks also calculated regional attenuation curves and K nomograms for their regions; however, in both cases, the results were fairly close to those of the Rautian (1958) formulation and therefore these regional nomograms were not put into use. K has also been back-estimated from various magnitudes and, specifically in the northern Caucasus, from duration magnitudes for estimating a duration K (Gabsatarova, 2002).



**Figure 6** The Dagestan nomogram for SKM seismometer records. Vertical axis is in  $\log A$  in microns ( $\mu$ ) and the horizontal axis is logarithmic in distance ( $r$ ) in km, here up to 300 km. (copied from Figure 4 in Adilov et al., 2010),

In summary, we realize a need to derive regional calibration relationships for short-period K estimates, aimed at classifying mainly local and regional earthquakes down to very small size. If properly done, they would allow us to compensate for the unavoidable and often very significant variations in local/regional wave propagation conditions, thus making data derived by different networks, for events in different regions, compatible. This is the same situation as for the classical local magnitude scale  $M_L$ , which is applied in the same frequency range as the K scale. Meanwhile numerous regional  $M_L$  calibration relations have been published (see DS 3.1). Yet, one of the disadvantages of the K class is currently the inconsistency in which some of these regional formulas have been derived and scaled (e.g., using only P or S or the sum of P+S amplitudes, only A or A/T, and different types of responses for deriving or applying such relationships). Thus there is an urgent need for standardization of procedures compatible with modern recording and processing techniques. In conjunction with related international efforts for standardization of the procedures for the calculation of the most common magnitude measurements are needed that ensure the global compatibility of values of the same type, and the stability of the relationships between complementary types of parameters describing earthquake size or strength (see, IS 3.3 specifically, and Chapter 3 in general).

## 4 Relationship of K class to magnitude scales

(Note: In the following we use two types of magnitude nomenclature:

a) the NMSOP generic nomenclature which generally writes magnitude symbols not in *Italics* and without subscript and prefers to write the local magnitude in general as *Ml* instead of the original Richter (1935)  $M_L$  or the new IASPEI (2011) standard *ML*; b) the nomenclature for the IASPEI (2005 and 2011) approved standard magnitudes *mb*, *ML*, *Ms\_20* and *Ms\_BB*.

Magnitude, *M*, began to be used throughout the USSR for larger earthquakes starting in 1955. The so-called “Prague-Moscow magnitude” formula for surface-waves (Vanek et al., 1962) became the accepted standard in the annual *Earthquakes in the USSR (Zemletryaseniya v SSSR)* starting with the 1962 compilation and became the international IASPEI standard in 1967. This magnitude, *Ms\_BB* when written in recent IASPEI standard nomenclature (see below), was computed using the ratio  $(A/T)_{max}$  of the whole surface-wave train in a wide range of periods between 3 s and 30 s (more recently extended to 60 s), originally as recorded by medium-period (SK) and the more long-period (SKD) Kirnos type broadband displacement seismographs (see Fig. 1) in the distance range between 2° and 160° (today measured directly from very broadband velocity meters; see Bormann et al., 2009 and IS 3.3).

Since local and regional networks in the FSU primarily used short-period instruments, the calculation of K class for stronger events with longer period energy was not possible. Thus K class was generally calculated for smaller events, while magnitude was calculated for larger events; however, both were potentially calculated for magnitudes between 4 and 5.5 (corresponding to K class 11 to 14).

The primary difference between K class and *M* is that K class is technically calibrated to energy. Empirically, the 20 s *Ms*, now standard *Ms\_20*, has also been scaled to seismic energy by Richter (1958) using the log *Es*-*mB* and *mB*-*Ms* relationships in Gutenberg and Richter (1956) and, more recently, by Choy and Boatwright (1995) based on direct *Es* estimates from velocity broadband records (see discussion below). The “Prague-Moscow” *Ms*, equivalent to current standard *Ms\_BB* (see IS 3.3), is presumed to be even more directly related to seismic energy than *Ms\_20* since it measures  $(A/T)_{max}$  over a much wider period range. As originally developed, by using amplitudes of both P and S waves, K-class utilized a somewhat better estimate of the overall ground motion, thus making it less dependent on the effects of the focal mechanism. On the other hand, longer period surface wave measurements will be less affected by attenuation and scattering in heterogeneous media and by local site effects.

As noted above, since  $A_S$  dominates over  $A_P$ , many K values calculated throughout Russia today are only based on  $A_{Smax}$ , e.g., in Kamchatka and Sakhalin, so that they should be viewed as being in fact an equivalent to *Ml* and are considered as such in the ISC Bulletin.

Since *Es* is proportional to  $A^2$ , assuming constant *T*, the relationship between K and *Ml* should be of the form

$$Ml = c + 0.5 K \quad (2)$$

where *c* is a constant. However, variations in attenuation and radiated energy spectra, source depth, and tectonic styles will cause differences both between, and within, regions. Conceptionally, formula (2) is disputable because signal duration is not taken into account, yet still it may work in practice reasonably well.

According to Rautian (1960), early studies for Tadzhikistan yielded

$$M = (M_L) = -2.22 + 0.56 K = -2.22 + 0.56 (\log E_S) = 5.62 + 0.56(K - 14) \quad (3)$$

in the K class range  $4 \leq K \leq 13$ .

In this context it is worth mentioning that Berckhemer and Lindenfeld (1986) found that when calculating energy from broadband velocity records,  $\log E_S \sim 2 M_L$ , which corresponds precisely to (2), and is close to (3), the original empirical estimate by Gutenberg and Richter (1956) for Southern California ( $\log E_S \sim 1.92 M_L$ ), and to the much more recent relationship derived by Kanamori et al. (1993) based on US TERRAScope data:

$$\log E_S = 1.96 M_L + 2.05 \quad (4a)$$

which could also be written as

$$M_L = 6.1 + 0.51 (K - 14). \quad (4b)$$

Since K class is calculated from short-period instruments, like mb and  $M_L$ , it saturates at around  $K \approx 15$  to 17 or  $M \approx 6$  to 7, unlike  $M_S$ , which saturates around 8-8.5 (Kanamori, 1983), corresponding to  $K \approx 18$ -19.

To determine the variations between regions of the empirical relationship between M and K class, Rautian et al. (2007) tabulated M and K class values reported for 1970-1997 for each of the seismic regions used in *Earthquakes in the USSR*, and its successor publications, and compared them with mb and  $M_S$  values reported in the catalog of the International Seismological Centre (ISC). Because K class and M are both independent variables with their own uncertainties, they used an orthogonal regression which minimized the sum of the squares of the distance to the regression line in both the ordinate and abscissa directions. Because K class tends to be calculated for smaller events, and magnitude for larger events, and both are calculated using different methods (see above, for K, and Chapter 3 and IS 3.3 for  $M_S$ ) there is some inconsistency and scatter in the different magnitude – K class ( $\log E_S$ ) relationships (for more details see Rautian et al., 2007; also Sadovsky et al., 1986, and Riznichenko, 1992). All regressions in Rautian et al. (2007) and in Table 1 were standardized to the general form

$$M = c + s (K - 14). \quad (5)$$

By using  $K - 14$ , the sign on c is always positive and makes comparisons clearer in the range  $9 \leq K \leq 14$  where both K and M are likely to have been calculated.

The mb regressions (see Figure 7a and for details Table 1) are generally similar, close to an average relationship of

$$mb = 5.41 + 0.43 (K - 14) \quad (6)$$

in the K class range of interest ( $9 \leq K \leq 14$ ), except for the Crimea, Sakhalin, Kurile, and Kamchatka. The Crimea, Sakhalin, and Kamchatka regions have higher constants and slope values than the other regions, while Kurile region has a similar slope, but a larger constant. The different constants suggest differences in attenuation conditions, and different slopes point to different amplitude-distance-source-depth conditions and relationships than those on which the magnitude and K class formulas are based. This clearly applies to the three Far Eastern regions (Sakhalin, Kurile, Kamchatka), which have different tectonic settings in



subduction zones with a different and more complex velocity structure and attenuation conditions. This explains why their K formulas differ significantly from the rest of the former USSR. It is interesting that the Kurile regression differs from both Sakhalin (which administers the Kurile network) and Kamchatka (which is tectonically similar).

For Central Asia, where Rautian (1958) developed the K class scale, Rautian (1960) gave the relationship  $m_b = 5.48 + 0.55 (K - 14)$  and Rautian et al. (2007) calculated  $m_b = 5.53 + 0.449 (K - 14)$ ; both are close to the average relationship given in (6).

If one replaces K by  $\log E_s$  in equation (6) and solves for  $\log E_s$  in order to make it comparable with classical  $\log E_s$ -M relationships, one obtains

$$\log E_s = 2.326 m_b - 1.42 \quad (7)$$

which has a slope remarkably similar to the original Gutenberg and Richter (1956) relationship

$$\log E_s = 2.4 m_B - 1.2 \quad (8)$$

although Eq. (8) is meant to be used for the medium-period broadband  $m_B$  measured by Gutenberg at periods between 2 s and 20 s (primarily between 5 s and 10 s) and not for the short-period  $m_b$ . However, for equal values of  $m_b$  and  $m_B$  between 5 and 7, Eq. (8) yields energy values 4 to 5.5 times larger than from Eq. (7). Moreover, for equal events, broadband  $m_B$  is generally larger than  $m_b$  (on average by about 0.3 – 0.5 m.u. between  $m_b = 5$  and 7 and this difference increases further with magnitude. Thus, event estimates of  $E_s$  in this magnitude range are about 8 to 17 times larger when based on  $m_B$  than those based on  $m_b$ .

**Table 1** Regressions between K class and magnitudes\*

| Region        | $m_b$                   | $M_s$                   | $M_e$                 | N( $M_e$ ) | $M_e$<br>$r^{2**}$ |
|---------------|-------------------------|-------------------------|-----------------------|------------|--------------------|
| Carpathians   | $5.54 + 0.397 (K-14)$   | $5.92 + 0.661 (K-14)$   |                       |            |                    |
| Crimea        | $6.20 + 0.699 (K-14)$   | Insufficient Data       |                       |            |                    |
| Caucasus      | $5.60 + 0.391 (K-14)$   | $6.02 + 0.782 (K-14)$   | $6.49 + 0.547 (K-14)$ | 8          | 0.61               |
| N-Caucasus    |                         |                         | $5.23 + 0.635 (K-14)$ | 3          | 1.00               |
| Kopetdag      | $5.53 + 0.467 (K-14)$   | $5.71 + 0.781 (K-14)$   | $5.15 + 0.672 (K-14)$ | 5          | 0.93               |
| Central Asia  | $5.53 + 0.449 (K-14)$   | $5.36 + 0.594 (K-14)$   | $5.73 + 0.218 (K-14)$ | 15         | 0.10               |
| Altai - Sayan | $5.47 + 0.482 (K-14)$   | $5.37 + 0.633 (K-14)$   | $5.79 + 0.517 (K-14)$ | 5          | 0.59               |
| Baikal        | $5.23 + 0.434 (K-14)$   | $5.54 + 0.828 (K-14)$   | $5.13 + 0.474 (K-14)$ | 5          | 0.60               |
| Yakutia       | $5.49 + 0.427 (K-14)$   | $5.55 + 0.539 (K-14)$   | Insufficient data     |            |                    |
| Northeast     | $5.33 + 0.445 (K-14)$   | Insufficient Data       | Insufficient data     |            |                    |
| Amur          | $5.01 + 0.394 (K-14)$   | $5.10 + 0.755 (K-14)$   |                       |            |                    |
| Sakhalin      | $7.25 + 0.669 (K_S-14)$ | $7.57 + 0.773 (K_S-14)$ | Insufficient data     |            |                    |
| Kurile***     | $6.30 + 0.460 (K_S-14)$ | $6.56 + 0.642 (K_S-14)$ | $6.93 + 0.729 (K-14)$ | 18(39)     | 0.85               |
| Kamchatka     | $6.11 + 0.552 (K_F-14)$ | $6.47 + 0.838 (K_F-14)$ | $6.26 + 0.657 (K-14)$ | 52         | 0.66               |

\*  $m_b$  and  $M_s$  correlations from Rautian et al. (2007).

\*\*  $r^2$  is Pearson's correlation coefficient squared.

\*\*\* see Rautian et al. (2007) for calculation methodology; for N( $M_e$ ) the number of bins is given, the number of total data points is in parentheses.

Ms regressions by Rautian et al. (2007) are more variable (Figure 7b, Table 1), reflecting the fact that the methodology and frequencies used for calculating mb and K are much closer than those for determining Ms. Further, one has to distinguish between Ms as generally reported by NEIC, measured in a the narrow range of periods between 18 and 22 s at teleseismic distances between 20° and 160° only, and the broadband Ms measured today at periods between 3 s and 60 s and in a wider distance range between 2° and 160°. Although both these values of Ms are calculated according to the “Prague-Moscow” IASPEI standard formula  $M_s = \log(A/T) + 1.66 (\log \Delta^\circ) + 3.3$ , the former, now termed Ms\_20, is mostly used in western countries and at all stations and centers reporting to the NEIC, whereas the latter, now termed Ms\_BB, is the more correct application of the “Prague-Moscow” formula and has been used ever since its development in the FSU and allied countries, as well as by China. The orthogonal Ms-K relationships tabulated by Rautian et al. (2007) use Ms values published by the ISC, which are based on a mix of both types of data, depending on what has been reported to it.

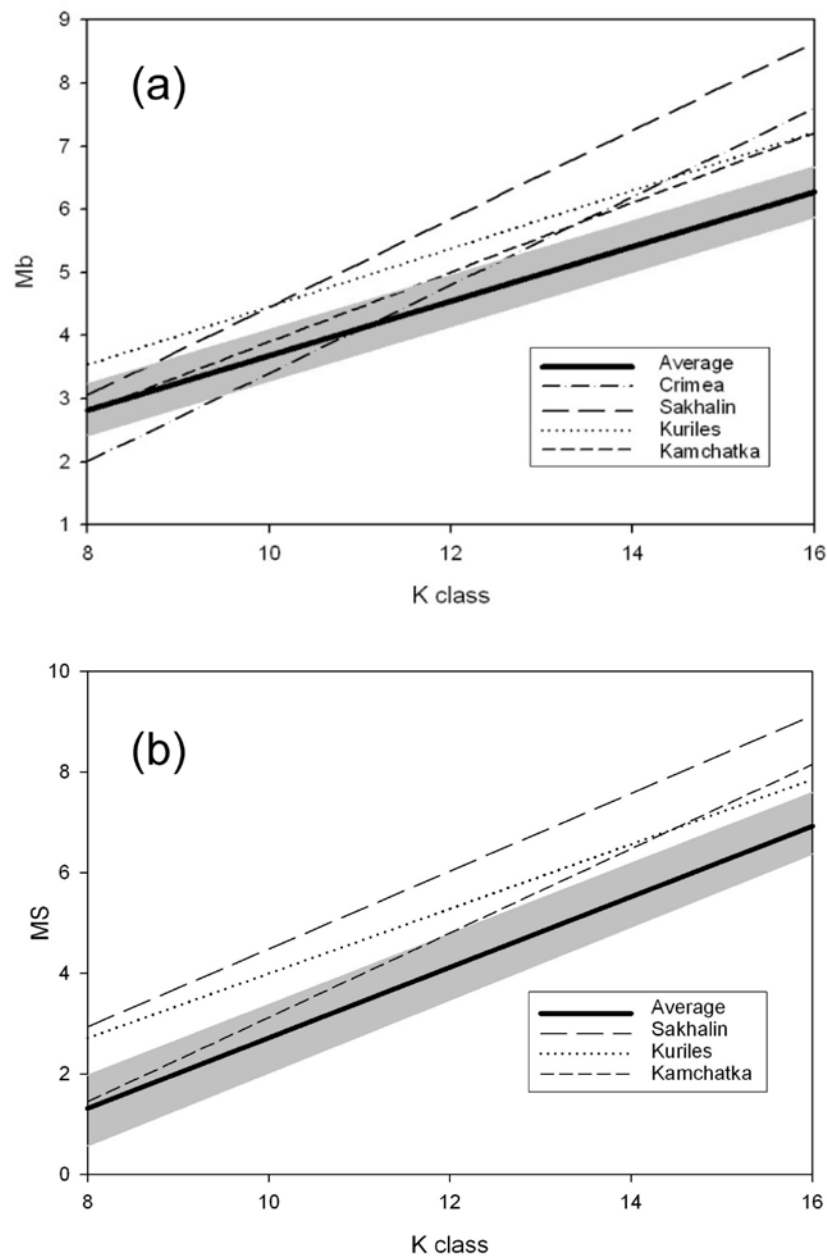
The slopes of Rautian et al. (2007) regressions vary from 0.5 to 0.8 with a mean regression of

$$M_s = 5.52 + 0.702 (K - 14), \quad (9)$$

excluding the Carpathians, which are based on very limited data, and the subduction zone networks of the Russian Far East, which have  $c$  values  $> 6.4$  and used different methodologies for calculating K as noted above. In general, however, most of the curves are fairly close to each other in the range  $10 \leq K \leq 14$  (Figure 7b).

In this context we also draw the attention to the non-linear relationships between K-Fedotov and Mw (see Figure 9) as well as the linear relationships between K-Fedotov and the short-period Obninsk mPVA = mSKM, published by Gusev and Melnikova (1991) and Gusev (1992). Note that mb(NEIC, ISC) is on average smaller by about 0.18 magnitude units than mPVA = mSKM, because mb is mostly measured on records with responses similar to that of the Benioff instrument with a slightly lower seismometer eigenperiod and a narrower bandwidth than SKM.

Russian regional events listed in western seismicity catalogs, including the ISC, are reported with magnitudes calculated from K class using locally derived regressions, although this is not always clearly stated, resulting in uncertainty as to which the primary determination was. Moreover, there have been procedural changes which affect how the magnitudes were calculated from K class which bias reported values. For examples see Rautian et al. (2007).



**Figure 7** Comparisons of linear orthogonal regressions between K class and (a) mb and (b) Ms for various regions of the FSU. Most regressions fall within the shaded region; the mean of those regressions is shown as black line and corresponds to the formulas (6) for mb and (9) for Ms, respectively. Outlier regressions for Sakhalin, Kuriles and Kamchatka are shown and labeled separately (copied from Figure 8 in Rautian et al., 2007, *Seism. Res. Lett.*, **78**(6), p. 586, © Seismological Society of America).

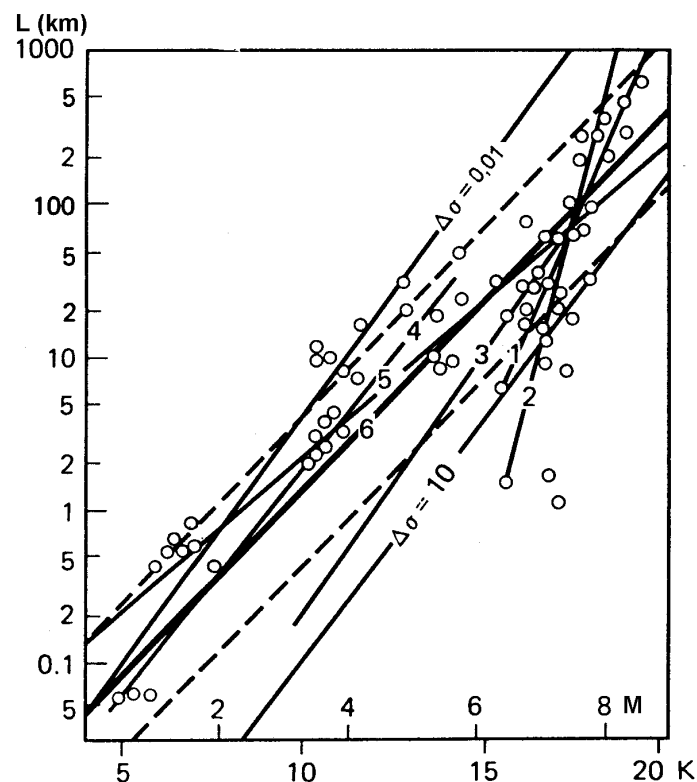
Replacing K by  $\log E_s$  in (9) one obtains

$$\log E_s = 1.424 M_s + 6.14. \quad (10)$$

The slope is similar but with a very different constant relative to the commonly cited relationship  $\log E_s = 1.5 M_s + 4.8$ , derived (see below) from the relationships published by

Gutenberg and Richter (1956). The constant in (10) is even more different than that in the relationship  $\log E_S = 1.5 M_s + 4.4$ , derived by comparing NEIC  $M_{s\_20}$  with  $E_S$  measurements obtained by direct integration and Earth model scaling of squared P-wave velocity amplitudes (Choy and Boatwright, 1995). However, almost all data calculated by Choy and Boatwright (1995) are from  $M_s$  values between about 5.5 and 8.5, i.e., strong earthquakes, at which  $K$  values tend to saturate or are already saturated. This might explain the very large constant in (10) which results in an underestimation of  $M_s$  as derived from  $K$  for weak earthquakes ( $K < 10$ ). This likely could be corrected by correlating  $K$  not with the  $M_{s\_20}$  dominated ISC  $M_s$  but rather with  $M_{s\_BB}$ , which is applicable in the near regional range also for periods much smaller than 20 s. Bormann et al. (2009) showed that, on the average,  $M_{s\_20} = 3.5$  is equivalent to  $M_{s\_BB} = 4$ . Such a revised  $K$ - $M_{s\_BB}$  correlation is strongly recommended in the future when more  $M_{s\_BB}$  data, now a IASPEI standard (see IS 3.3), become available on a global scale, the FSU included. This will allow a direct correlation of data determined by these two independent approaches with different instruments. Regardless, in the range of  $K$  between about 12 and 16,  $M_s$  estimates determined using (9) are reasonable, with errors likely to be smaller than about 0.5 magnitude units.

Finally, we refer to the extremely important compilation of data by Riznichenko (1992), comparing  $K$  and  $M$  with  $\log M_0$  in relation to stress drop variations by more than three orders from 0.01 to more than 10 MPa (Figure 8).



**Figure 8** Correlation of source length  $L$  with magnitude  $M$  and energy class  $K$  according to data from various sources (e.g., curve 1 by Tocher, 1958, curve 2 by Iida, 1959; curve 6 average by Riznichenko, 1992; curves 3 to 5 from other authors quoted in Riznichenko, 1992). Thin straight lines: the associated stress drop,  $\Delta\sigma$ , are given in MPa; broken lines mark the limits of the 68% confidence interval with respect to the average curve 6 (modified from Riznichenko, 1992, Fig. 3; with permission of Springer-Verlag).

The values of  $M$  given in this diagram on the abscissa, along with  $K = \log E_s$ , correspond, depending on the distance and event size, to either  $M_L$  (for values  $< 4$  to about 6),  $mb$  (for  $> 4$  to about 5.5),  $mB$  (between about 5.5 to 7.5) or  $M_s$  ( $> 6$  to about 8.5). The  $M$  and  $K$  values given on the abscissa correspond to the following average “global” relationships between  $K$  and  $M$ :

$$\log E_s \approx K = 1.8 M + 4.1, \quad (11)$$

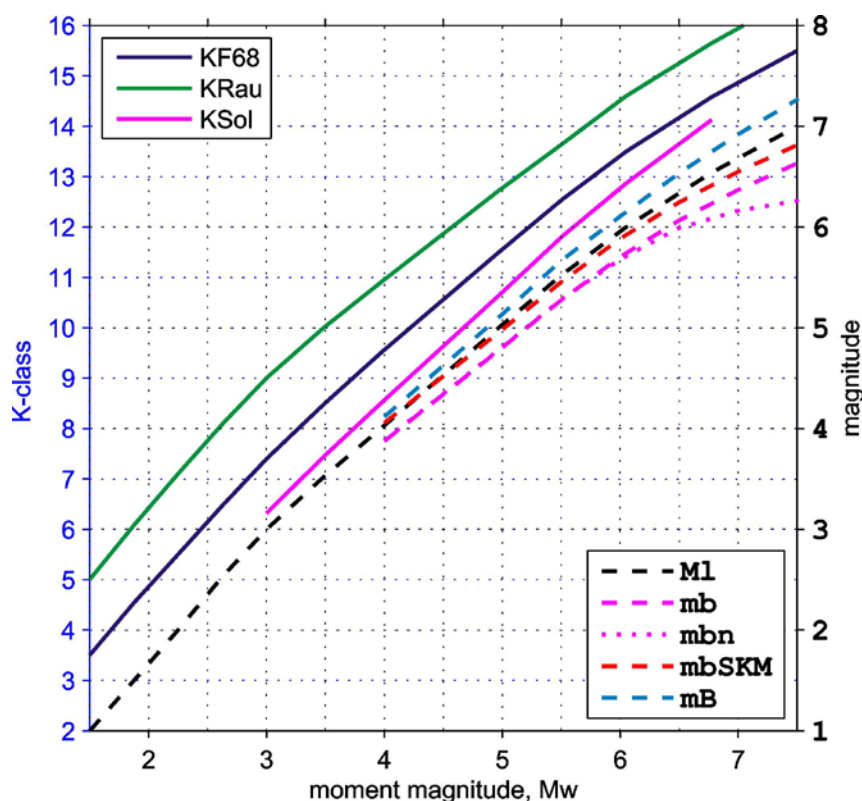
$$M = 0.556 K - 2.3 \quad \text{or} \quad (12)$$

$$M = 5.5 + 0.556 (K - 14) \quad (13)$$

with the understanding as to which approximate type of specific magnitude,  $M$ , this applies to in different magnitude ranges. The range of uncertainty is on the order of about 0.3 magnitude units. If one compares Eq. (13) with the Eq. (4b) for  $M_L$ , (6) for  $mb$  and (9) for  $M_s$ , then one realizes that the constants in all these relations are within a few tenths, that the slope in (13) is very close to that in (4b) for  $M_L$ , and in between, the slopes are 0.43 for the  $mb$ - $K$  and 0.70 for the  $M_s$ - $K$ . The agreement between Eqs. (4b) and (13) could likely be further improved by correlating  $K$  specifically with  $M_s$ \_BB, using values determined in the local and regional distance range, typically measured at periods between 2 s and  $< 15$  s.  $mb$ , however, is generally not as good an estimator for earthquake size and energy release for magnitudes above 5.5, for which  $mB$ \_BB is better suited. Yet, since  $mb$  is measured in the teleseismic range only, few values are available for event magnitudes below 3.5. Therefore, in the local and regional distance range,  $K$  and  $M_L$  are better estimators of the released seismic energy for smaller earthquakes, and  $M_s$ \_BB for stronger earthquakes.

When working with small earthquakes, it is often useful to relate  $K$  values to  $M_w$  which is practically a standard reference magnitude at present. Direct determination of  $M_w$  meets with an obstacle for small earthquakes: their flat low-frequency part of the displacement spectrum is rarely observed with an acceptable signal-to-noise ratio. Thus  $M_0$  or  $M_w$  are impossible to determine en masse on a routine basis, and catalogues of small earthquakes inevitably contain  $K$  (or  $M_L$ ) as a main magnitude value. To convert these catalogue values to  $M_w$  at least approximately, an adequate  $M$  to  $M_w$  relationship is a must. .

Figure 9 gives non-linear relationships between different  $K$  scales,  $M_L$ ,  $mB$ , and variants of  $mb$ , on one side, and  $M_w$ , on the other side. The  $mb$  values are based on records with short-period Benioff or WWSSN-SP instruments, respectively equivalent simulation filter responses (NEIC practice) and  $mb$ SKM is from SKM-3 instrument (Obninsk practice). The  $mb_n(M_w)$  trend reflects  $mb$  values obtained through older NEIC practice with narrow time window of amplitude measurement; see figure caption for details. Figure 9 is partly based on average relationships published by Gusev and Melnikova (1991) and Gusev (1992), extended here to the  $M_w = 2$ -5 range. Various other sources were also used; the most significant ones are Hanks and Boore (1984) and Tan and Helmberger (2007).



**Figure 9** Various K-classes and magnitudes vs. moment magnitude  $M_w$ . The left ordinate scale is K-class. Corresponding relationships (solid curves) are given for Fedotov (1968) S-wave K-class used on Kamchatka, Soloviev (1967) K-class used at Sakhalin for Kurile Islands, and Rautian (1960) K-class used in Central Asia and, in modified form, for many other continental areas of the FSU. The right ordinate scale is magnitude. Corresponding relationships (broken and dotted lines) are given: for classic local magnitude  $M_l$ ; for variants of mPV measured on short-period Benioff or WWSSN-SP type of records within a narrow measurement time window of 5-10s after the P-wave first arrival, denoted  $mbn$ ; its new IASPEI standard variant measured within an “unlimited” time window between the P and PP first arrival (see IS 3.3), denoted  $mb$ ; the similar Russian mPV variant measured on slightly longer-period SKM-3 records, denoted  $mbSKM$  (with  $mbSKM \approx mb + 0.18$  on average); and for medium-period, traditional  $mB$  of Gutenberg.

Note that the relationships  $M_l$  vs.  $M_w$  or  $K$  vs.  $M_w$  vary significantly between regions and even sub-regions. Therefore, the curves in Figure 9 can serve only as a preliminary reference. The “saturation effect” for body-wave magnitudes is clearly seen on Figure 9. Yet, it should be kept in mind that true saturation takes place only for  $mbn$  that scatters around the  $mbn = 6.4$  level at  $M_w \geq 7.5$ . In contrast,  $mb$  values based on amplitude readings within “unlimited” P-wave measurement time windows such as IASPEI standard  $mb$  (Bormann et al., 2009), its forerunner applied by Houston and Kanamori (1986), and the Obninsk  $mbSKM$  (Gusev and Melnikova, 1991) keep growing up to  $M_w$  around 9, attaining then values of about 7.4 to 7.6.

Note that recently, the Kamchatka network began to publish regional  $M_l$  directly calculated from the Fedotov (1968) S-wave K-class via the conversion relationship

$$M_l = 0.5K_F - 0.75. \quad (14)$$

This relationship is in agreement with the related curves of Figure 9. Interestingly, (14) is slope-wise identical with the equivalent formula (4b) derived above for California, although the constants differ by 0.29 m.u.. Accordingly, for equal  $K = \log E_S$ , Eq. (14) would yield about 0.3 m.u. smaller  $M_L$  values than (4b), hinting to regional differences in attenuation, the latter being larger in Kamchatka.

## 5 Future development and use of $K$ for energy-related earthquake classification

The paper by Rautian et al. (2007) reviewed the  $K$  classification of earthquakes mainly to

- explain the methodology of its development and modifications over time in different seismotectonic regions of the FSU,
- provide the most recent relationships between  $K$  and the  $m_b$  and  $M_s$  magnitudes as published by the International Seismological Centre, and
- give recommendations - sometimes cautionary - for the interpretation and use of the extensive  $K$ -class data available in current Russian and past FSU catalogs in terms of common magnitude scales, so as to make seismicity data for the vast territory covered by the  $K$  class system compatible with those of the rest of the world.

However, in this section, we would like to go one step further and look at the  $K$ -class system not only from a historical and data compatibility perspective, but in terms of the merits of this approach in general.

It is interesting to note that Gutenberg and Richter (1956) published the first semi-empirical energy-magnitude relationship, based on rather meager data, relating energy in Joules (J), to body-wave magnitude:

$$\log E_S = 2.4 m_B - 1.2 \quad (15)$$

Combining (15) with the Gutenberg and Richter (1956) body-wave to surface-wave magnitude conversion relationship,

$$m_B = 0.63 M_s + 2.5 \quad (16)$$

the often quoted Gutenberg-Richter magnitude energy relationship

$$\log E_S = 1.5 M_s + 4.8 \quad (17)$$

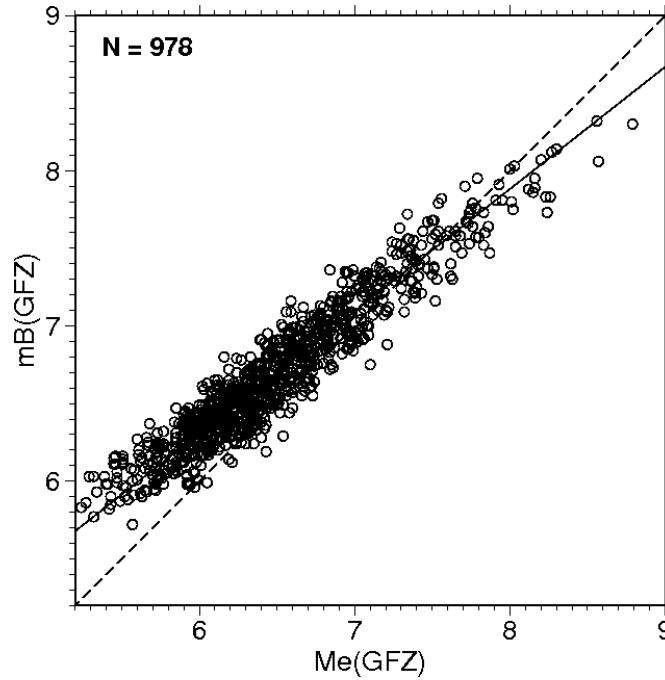
is obtained. Hanks and Kanamori (1979) based their scaling of seismic moment  $M_0$  to  $M_s$ , and thus the definition of their moment magnitude scale,  $M_w$ , on this relationship. Also the Russian authors of the  $K$ -class system calculated  $E_S$  in Joules from this equation, even though it is based on 20 s surface waves, which are not representative for the assessing the high-frequency related damage potential of earthquakes or the total seismic energy release of smaller earthquakes with corner frequencies well above 0.1 Hz. Gutenberg and Richter (1956) published only Eq. (15), relating  $E_S$  to  $m_B$ , which Gutenberg measured in a wide range of periods less than 20 s.

$K$  was supposed to be an estimate of  $\log E_S$ , and therefore should have been determined from the  $(A/T)$  spectrum. However, because of difficulties in conducting proper spectral analyses and integrating analog records in the past, all  $K$  formulations have been based on measuring

only  $A_{max}$ . Gutenberg and Richter (1956) realized that the body-wave scale would yield stable magnitude estimates only by normalizing the  $A_{max}$  value in the P-wave train by its period,  $T$ , which varied in a fairly wide range between some 2 s and 20 s even in records from the limited broadband responses of the medium-period seismographs of that time. Now, modern force-feedback broadband sensors have direct velocity proportional responses at least in the range between about 0.05 and 40 s, or wider, and allow the direct measurement in the entire P- or S-wave train, of the maximum velocity,  $V_{max}$ , which is related to the maximum seismic energy released by the seismic source at the corner frequency of its displacement spectrum (see Bormann, 2011). Thus, we can expect that the single amplitude magnitude

$$mB\_BB = \log (V_{max}/2\pi) + Q(h, \Delta), \quad (18)$$

where  $Q(h, \Delta)$  is the empirically determined Gutenberg and Richter (1956) calibration function, which depends on focal depth,  $h$ , and distance,  $\Delta$ .  $Q(h, \Delta)$  is available in tabulated and diagram form (see DS 3.1) and accounts for the cumulative effects of the diverse, and theoretically not yet sufficiently well known or tractable in detail influences of wave propagation. Thus,  $mB$  is, in fact, a reasonably good estimator for  $E_S$ . This has been confirmed by Di Giacomo, Saul and Bormann (see Bormann, 2011, and Figure 10).



**Figure 10** The relationship between automatically determined values of  $mB$ , equivalent to standard  $mB\_BB$ , and  $Me$ , according to procedures developed at the GFZ German Research Centre for Geosciences by Bormann and Saul (2008) and Di Giacomo et al. (2010). The dashed line is the 1:1 relationship and the solid line the orthogonal regression relation (19). (Figure by courtesy of D. Di Giacomo, 2011).

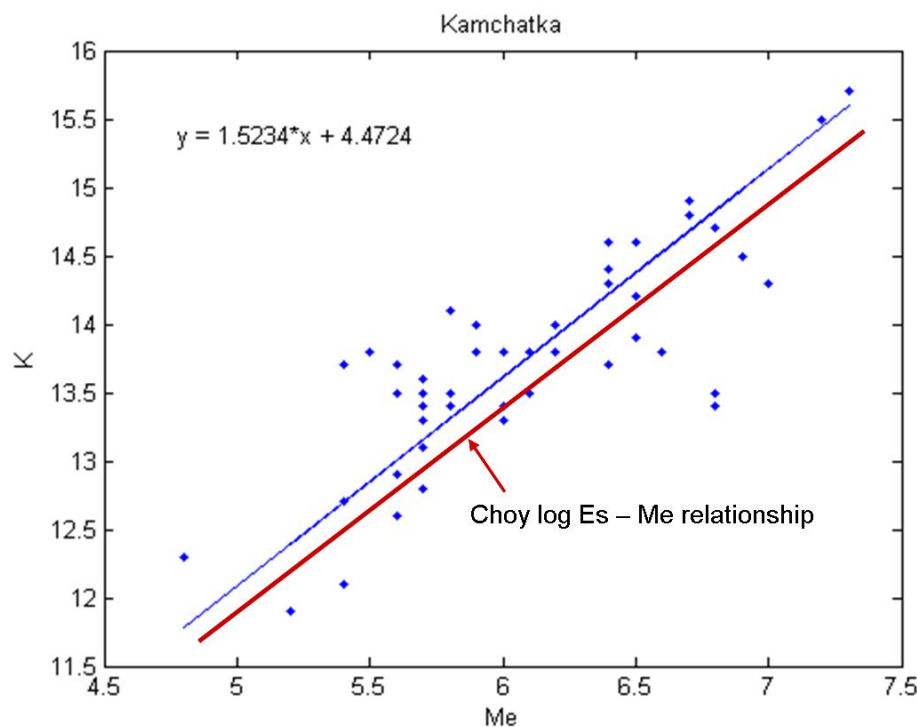
In the range  $5.5 < mB\_BB < 8.5$  the data in Figure 10 satisfy the following orthogonal relationship between  $mB\_BB$  and the energy magnitude  $Me = (\log E_S - 4.4)/1.5$ :

$$mB\_BB = 0.79 Me + 1.58. \quad (19)$$



Eq. (19) permits the estimation of  $M_e$  (and thus  $E_S$ ) by using very simple and easily to automate  $mB\_BB$  measurements (see Bormann and Saul, 2008) with a standard deviation  $\sigma = 0.18$  magnitude units, about the best average of direct  $M_e$  measurements. This “error” corresponds to an approximately 95% ( $2\sigma$ ) uncertainty in  $E_S$  estimates by a factor of less than 4, which is close to the factor of 2-3 given by Kanamori and Brodsky (2004) for  $E_S$  estimates via the integration of very good broadband seismic data.

Choy and Boatwright (1995) and Boatwright et al. (2002) suggested a method of calculating energy magnitude,  $M_e$ , based on broadband digital data, which has been applied by the U. S. Geological Survey. Therefore, we have attempted orthogonal regressions between  $K$  and  $M_e$  from the U. S. Geological Survey database for a limited number of regions in which data were available (see Table 1) using the same regression method as in Rautian et al. (2007). As noted above, in most cases, the number of events is small ( $< 5$ ), because  $M_e$  is calculated only for larger events (generally  $M_e > 5.5$ ) and  $K$  for smaller local and regional events (generally  $K < 14$ ). Only for the region of Kamchatka enough data have been available to plot  $K$  as a function of  $M_e$ (USGS) (see Figure 11). Despite the large data scatter, the agreement between the orthogonal  $K$ - $M_e$  regression relationship and the  $\log E_S$ - $M_e$  relationship by Choy is surprisingly good, for constant  $M_e$  being within 0.15 to 0.25 units of  $K$ , i.e., just a factor of 1.4 to 1.8 larger in  $E_S$  than from the teleseismic estimate, or, for equal  $K$ , about 0.1 to 0.2 m.u. smaller in related  $M_e$  as compared with the Choy  $\log E_S$ - $M_e$  relationship.



**Figure 11** Orthogonal regression relationship between  $K$  measured for Kamchatka earthquakes and the related teleseismic  $E_S$  - $M_e$ (USGS) relationship.

For other regions, with much less data, the relationships between  $K$  and  $M_e$  are so far less clear and more variable, also with respect to level differences. Central Asia data are the most varied and the regression is biased by a series of low  $M_e$ , but relatively high  $K$ , events (see Table 1). This may hint to rather high-frequency waves radiated from high stress drop earthquakes, which are not well covered by teleseismic observations.. As the data are

concentrated in a narrow range around  $M_e = 5.5-6.0$  and  $K = 14-15$ , this regression has a very low correlation coefficient. Other relationships are more stable, e.g., for the Kurile Islands and the Caucasus, yielding for  $K$  values between 12 and 15.5  $M_e$  estimates that differ  $-0.1$  to  $+0.6$  m.u. from the values derived via the Choy relationship  $M_e = (\log E_s - 4.4)/1.5$ . The latter (for derivation see IS 3.5 and Bormann and Di Giacomo, 2011) can be written in terms of  $K$  as.

$$M_e = 0.667 K - 2.933 = 6.40 + 0.667 (K-14). \quad (20)$$

For magnitudes below 5.5, however, global  $M_e$  is rarely measured because of the small signal-to-noise ratio of  $P$  waves in teleseismic broadband records as well as the difficulties in including frequencies higher than 2 Hz into the complex model calculations.

In their conclusions, Rautian et al. (2007) correctly state that currently used methods for estimating seismic energy and energy magnitude,  $M_e$ , that are based on global models, commonly make simplified theoretical assumptions with respect to Earth structure, attenuation, scattering, site amplification, source radiation pattern, and/or other local or regional geologic factors that preclude their use at non-teleseismic distances.. Especially for small earthquakes recorded at local and regional distances, no easy practical and reliable methodology is currently available. On the other hand, although  $K$  is “ideologically” an energy scale, its practical implementation and use so far is more of a magnitude-like nature, with  $K$  being associated with energy only by correlation (Kondorskaya et. al., 1981). More truly energy-related  $K$  scales would require the consistent measurement of maximum ground motion velocity instead of displacement.

Despite the drawback of current  $K$  scales, the agreement in trend and level of the relationship between the (A/T) derived  $K$  values for earthquakes in Kamchatka (see Figure 5) and teleseismic  $M_e$ (USGS) is very promising. It lets us expect that a  $K$ -type procedure could be developed, which allows to estimate  $E_s$  and  $M_e$  for large numbers of local and regional earthquakes sufficiently reliably in a similarly simple, interactive or easy to automate way as via teleseismic  $mB_{BB}$ . Therefore, we recommend to further develop the  $K$  class system for events with magnitudes below 6, down to magnitudes of about 2, which are recordable with good signal-to-noise ratio at regional and local distances with standardized high-frequency responses. They should have a velocity-proportional pass-band of not less than about 5 octaves relative bandwidth, covering the frequency range between about 0.4 to 10-20 Hz). This would allow to measure in this frequency range directly the largest event velocity amplitudes. This eases both the derivation of local energy-related  $K$  calibration functions as well as later routine  $K$  measurements for estimating  $E_s$  and  $M_e$  of local earthquakes.

On the other hand, we realize from Figure 7 that the orthogonal regression relationships between  $K$  and teleseismic  $m_b$ , respectively  $M_s$  for earthquakes in the Russian subduction zone environments of Kamchatka, Kurile Islands and Sakhalin differ systematically, partially depending on  $K$ , from those derived for other intra-continental regions in FSU states. The differences in  $M_s$  (for equal  $K$ ) reach up to one and for  $m_b$  up to two magnitude units. This is not acceptable and hints, as mentioned already above for Kamchatka, to significant differences in the absolute calibration of the various regional  $K$  scales. Thus there is an urgent need to develop standards for both calibration in continental (shallow earthquakes) and subduction zone environments (with also deep and intermediate depth earthquakes) which yield  $K$  data that are really compatible in a wide range of  $K$  between the different seismotectonic environments and for larger events also with teleseismic  $M_e$  data.

Although there is no more need to measure period specifically for deriving the ground motion velocity, local records in a wide range of magnitudes and from different seismotectonic settings are likely to show a great variability of frequencies at which  $V_{max}$  is measured. The higher the measured frequencies the more the heterogeneity of the Earth medium and of the surface topography at local and regional scale will impact the attenuation effects. They may have to be compensated for by frequency-dependent and maybe even source–distance and azimuth-dependent local and regional calibration functions.

To derive and use such calibration functions goes well beyond traditional K formulas and their application. It will require years of careful and complex empirical data collection and analysis. However, digital data recording, filtering and processing in a standardized way eases this task. Yet, for assuring the better compatibility of future K estimates the new calibration functions have to be scaled so that they agree with original definitions and values at a suitable reference distance, in a similar way as regional Ml calibration functions in other countries have to be scaled to the standard Southern California Ml calibration function (see DS 3.1 and IS 3.3) in order to make them mutually compatible. This degree of compatibility has obviously not yet been achieved with respect to the various regional K-scales currently in use. E.g., in Figure 9, most of the difference between KF68 and KSol has no seismological reason. Rather, it is due to an error in calculating the absolute level of the  $(A_{max}/T)(\Delta, h=40\text{km})$  curve of Fedotov (1963). This makes the K-Sol values by 0.6 units lower than they should be (see Fedotov, 1972, p63). Moreover, the slight differences in slope of these two curves are most likely due to the difference of seismometer periods used, which are 0.7 s for KSol vs. 1.2 s for KF68. Minor differences between the regional K-scales may also originate from different assumptions made by the scale authors on the input parameters used in calculating the wave energy density on the reference sphere, e.g., Fedotov (1972) used  $v_s = 3 \text{ km/s}$ , whereas Rautian used  $v_s = 3.5 \text{ km/s}$ . Such details need to be checked and agreed upon in order to assure an unambiguous absolute calibration as well as standard (simulated) instrument responses and measurement procedures in order to assure interregional K-value compatibility. Only then it will be possible to interpret differences in measurement values that are beyond the range of measurement errors in seismological terms only.

Although most of the seismic energy is radiated in the S-wave group, teleseismic procedures for estimating  $E_s$  generally use the first arriving P waves. The reason is that S-waves from strong earthquakes with large rupture durations of several 10 seconds or even minutes are often superimposed by later secondary, phases. This may bias the results of full waveform integrations in a wide range of periods and rupture durations. Moreover, theoretical corrections for propagation and attenuation effects in a wide range of periods are more difficult to account for S waves than for P waves. But these objections against the use of  $V_{Smax}$  instead of  $V_{Pmax}$  are less relevant for the analysis of weak earthquakes in the regional and local distance range. Earthquakes with magnitudes below 6 have average rupture durations of less than 6 s. Recorded at distances typically less than 1000 km, they have clearly developed distinct later Sg/Lg groups that are well separated in time from the earlier and usually much weaker P-wave phases. This eases significantly proper phase identification and parameter measurement both in automatic and interactive modes.

Finally, another approach, considered by Rautian et al. (2007) to be the most appropriate way to get seismic energy along with seismic moment, apparent stress, etc., is to use the later part of the coda for source spectral estimation (Rautian and Khalturin, 1978). This method had been originally developed for analog records with multi-bandpass filtering and could easily be

adapted as well for digital instrumentation and processing techniques now available. A certain step in this direction is the routine use of short-period coda-class Kc on Kamchatka mentioned above.

## Acknowledgments

We thank Tatyana G. Rautian for extensive help in clarifying various points in the development and calibration of the original K-class scale. Anthony Kendall provided the K to Me correlations and Tea Godoladze assisted in obtaining information on the use of K in the FSU. Thanks go also to George Choy for his careful review and encouraging assessment. This research was supported, in part, by Department of Energy contracts DE-FC03-02SF22490, DE-FC52-2004NA25540 and DE-AC52-09NA29323 to Michigan State University.

## References

- Adilov, Z. A., O. A. Asmanov, M. G. Daniyalov, and M. A. Isaev (2010). Nomograms for determining the class of earthquakes in Dagestan, in A. A. Malovichko (editor), *Modern Methods for the Processing and Interpretation of Seismological Data, Materials of the 5<sup>th</sup> International Seismological School: Obninsk, Russian Academy of Sciences, Geophysical Survey*, 6-9.
- Berckhemer, H., and M. Lindenfeld (1986). Determination of source energy from broadband seismograms. *Geologisches Jahrbuch. Reihe E: Geophysik* **35**, 79-83.
- Boatwright, J., G. L. Choy, and L. C. Seekins (2002). Regional estimates of radiated seismic energy. *Bull. Seism. Soc. Am.*, **92**, 1241-1255.
- Bormann, P. (2011). Earthquake magnitude. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 207-218; doi: 10.1007/978-90-481-8702-7.
- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bormann, P., and J. Saul (2008). The new IASPEI standard broadband magnitude  $m_B$ , *Seism. Res. Lett.* **79**, 698-705.
- Bormann, P., R. Liu, Z. Xu, K. Ren, L. Zhang, and S. Wendt (2009). First application of the new IASPEI teleseismic magnitude standards to data of the China national seismographic network, *Bull. Seism. Soc. Am.*, **99**, 1868-1891.
- Bune, V. I. (1955). On classification of earthquakes by their strength on the basis of instrumental observations. *Doklady Acad. Sci. Tajik SSR* (Stalinabad-Dushanbe), issue 1, 48-55 (in Russian).
- Choy, G. L., and J. L. Boatwright (1995). Global patterns of radiated seismic energy and apparent stress, *J. Geophys. Res.*, **100**, 18205-18228.
- Choy, G. L., A. McGarr, S. H. Kirby, and J. Boatwright (2006). An overview of the global variability in radiated energy and apparent stress, in: Abercrombie R., A. McGarr, H. Kanamori (eds): Radiated energy and the physics of earthquake faulting, *AGU Geophys. Monogr. Ser.* **170**, 43-57.

- Di Giacomo, D., S. Paroli, P. Bormann, H. Grosser, J. Saul, R. Wang, and J. Zschau (2010). Suitability of rapid energy magnitude determinations for emergency response purposes: *Geophys. J. Intern.*, **180**, 361-374.
- Dzhibladze, E. A. (1971). Nomograms for the determination of the value of K of Caucasian earthquakes, *Izvestiya Academy of Sciences, USSR, Physics of the Solid Earth* **7**, 807-809.
- Fedotov S. A. (1963). On attenuation of transverse seismic waves in the upper mantle and on energy classification of near earthquakes with intermediate depth. *Izvestiya Acad. Sci. USSR, Sect. geophys.*, No.6.
- Fedotov, S. A. (1972). Energeticheskaya klassifikatsiya kurilo-kamchatskih zemletryasenii i problema magnitud (Energy classification of Kuril-Kamchatka earthquakes and the problem of magnitudes.) Moscow, *Nauka*, 116 pp.
- Fedotov, S. A., I. P. Kuzin, and M. F. Bobkov (1964). Detailed seismological investigations in Kamchatka in 1961-1962, *Bulletin (Izvestiya), Academy of Sciences, USSR, Geophysics Series* **8**, 822-831.
- Gabsatarova, I. P., chief compiler, 2002. Northern Caucasus (including Dagestan), in O. E. Starovit, editor-in-chief, *Earthquakes of Northern Eurasia in 1996*, Moscow: Russian Academy of Sciences, Geophysical Survey, 245-252. (in Russian).
- Gusev, A. A. (1991). Intermagnitude relationships and asperity statistics. *Pure Appl. Geophys.*, **136**, 515-527.
- Gusev, A. A., and Melnikova, V. N. (1992). Relationships between magnitude scales for global and Kamchatkan earthquakes. *Volc. Seism.*, **12**, 723-733. (English translation of the original Russian publication: Гусев А. А., Мельникова, В. Н. (1990). Связи между магнитудами - среднемировые и для Камчатки. *Вулканология и сейсмология*, №6, 55-63.
- Gutenberg, B., and C. F. Richter (1956). Earthquake magnitude, intensity, energy, and acceleration (second paper), *Bull. Seism. Soc. Am.*, **46**, 105-145.
- Hanks, T. C., and H. Kanamori (1979). A moment magnitude scale, *J. Geophys. Res.*, **84**, 2348-2350.
- [Hanks, T. C., and Boore, D. M. \(1984\). Moment-magnitude relations in theory and practice. \*J. Geoph. Res.\* \*\*89\*\*, \*\*B7\*\*, 6229-6236.](#)
- Houston, H. (1999). Slow ruptures, roaring tsunamis. *Nature*, **400**, 4009-410.
- Houston, H., and Kanamori, H. (1986). Source spectra of great earthquakes: teleseismic constraints on rupture process and strong motion. *Bull. Seism. Soc. Am.*, **76**, 19-42.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf).
- IASPEI (2011) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf).
- Iida, K. (1959). *J. Earth Sci. Nagoya Univ.*, **7**, 2, 98-107.
- Kanamori, H. (1983). Magnitude scale and quantification of earthquakes, *Tectonophysics* **93**, 185-199.
- Kanamori, H., and E. E. Brodsky (2004). The Physics of earthquakes, *Reports on Progress in Physics*, **67**, 1429-1496.

- Kanamori, H., J. Mori, E. Hauksson, T. H. Heaton, L. K. Hutton, and L. M. Jones (1993). Determination of earthquake energy release and  $M_L$  using TERRAScope, *Bull. Seism. Soc. Am.*, **83** 330-346.
- Kondorskaya, N. V., and N. A. Shebalin (editors) (1977). *New Catalog of Strong Earthquakes in the Territory of the USSR from Ancient Times to 1975*, Moscow: Nauka, 534 pp. (in Russian)
- Kondorskaya, N. V., Z. I. Aranovich, O. N. Solov'eva, and N. V. Shebalin (Eds.) (1981). *Instrukciya o poryadke proizvodstva i obrabotki nabludenii na seismicheskikh stanciyach ESSN* (Instructions on the recording and processing of observations from the Unified System of Seismic Monitoring of the USSR). *Inst. Fiziki Zemli Akad. Nauk SSSR*, Moscow, 272 pp.
- Lemzikov, V. K., and A. A. Gusev (1991). Coda-based energy classification of near Kamchatka earthquakes. *Volc. Seis.*, **11**, 558-578. English translation of the original publication in Russian: Лемзиков В.К., Гусев А.А.(1989). Энергетическая классификация близких камчатских землетрясений по уровню кода-волн. *Вулканология и сейсмология*, №4., 83-97.
- Polet, J., and Kanamori, H. (2000). Shallow subduction zone earthquakes and their tsunamigenic potential. *Geophys. J. Int.*, **142**, 684-702.
- Pustovitenco, B. G., and V. E. Kul'chitsky (1974). About energy estimation for earthquakes of the Crimea – Black Sea region, in N. V. Kondorskaya, I. L. Nersesov, I. V. Forbunova, T. G. Rautian, and O. A. Korchagina (editors), *Magnitude and Energy Classification of Earthquakes*, **2**, Moscow: Academy of Sciences of the USSR, Institute of Physics of the Earth, 113-124. (in Russian)
- Rautian, T. G. (1958). Attenuation of seismic waves and the energy of earthquakes, I, *Trudy Institut Seismostoiikogo Stroitel'stva i Seismologii Akademii Nauk Tadzhikskoi SSR*, **7**, 41-85. (in Russian)
- Rautian, T. G. (1960). Energy of earthquakes, in Y. V. Riznichenko (editor), *Methods for the Detailed Study of Seismicity*, Moscow: Izdatel'stvo Akademii Nauk SSSR, 75-114. (in Russian)
- Rautian, T. G. (1964). About the determination of the energy of earthquakes at distances up to 3000 km. *Akademiya Nauk SSSR, Trudy Instituta Fiziki Zemli*, no. 32, p. 88-93. (in Russian)
- Rautian, T., and V. I. Khalturin (1978). The use of coda for determination of the earthquake source spectrum, *Bull. Seism. Soc. Am.*, **68**, 904-922.
- Rautian, T., Khalturin, V., Fujita, K., Mackey, K., and Kendall, A. (2007). Origins and methodology of the Russian energy K-class system and its relationship to magnitude scales: *Seis. Res. Lett.*, **78**, 579-590.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale, *Bull. Seism. Soc. Am.*, **25**, 1-32.
- Richter, C. F. (1958). *Elementary seismology*. W. H. Freeman and Company, San Francisco and London, viii + 768 pp.
- Riznichenko, Yu. V. (1992). *Problems of seismology*. Mir and Springer Publishers, Moscow and Berlin-Heidelberg-New York, 445 pp. (English translation of the original Russian publication of 1985).
- Sadovskiy, M. A., O. K. Kedrov, and I. P. Pasechnik (1986). On the question of an energetic classification of earthquakes. *Izvestiya, Earth Physics*, **22**, 83-89.

- Solonenko, A. V. (1977). The dependence of the attenuation of seismic waves from the energy of earthquakes and the energy classification of near earthquakes, in Gorshkov, G. P.(Ed.): Seismicity and seismogeology of Eastern Siberia: Moscow, Nauka, p. 74-82.
- Solov'ev, S. L., and O. N. Solov'eva (1967). The relationship between the energy class and magnitude of Kuril earthquakes, *Izvestiya, Academy of Sciences of the USSR, Earth Physics Series* **3**, 79-84.
- Tan, Y, and Helmberger, D. (2007). A new method for determining small earthquake source parameters using short-period P waves. *Bull. Seismol. Soc. Amer.*, **97**, 1176-1195.
- Tocher, D. (1958). Earthquake energy and ground breakage. *Bull. Seism. Soc. Am.*, **48**, 147-153.
- Vanek, J., A. Zatopek, Karnik, N. V. Kondorskaya, Y. V. Riznichenko, E. F. Savarensky, S. L. Solov'ev, and N. V. Shebalin (1962). Standardization of magnitude scales, *Bulletin (Izvestiya), Academy of Sciences, USSR, Geophysics Series* **6**, 108-111.
- Wyss, M., and Brune, J.N. (1968). Seismic moment, stress, and source dimensions for earthquakes in the California-Nevada regions. *J. Geophys. Res.*, **73**, 4681-4694.

| Topic   | Source parameters and moment-tensor solutions                                                         |
|---------|-------------------------------------------------------------------------------------------------------|
| Author  | Günter Bock (†), Formerly GeoForschungsZentrum Potsdam, Germany                                       |
| Version | September 2002; editorially adapted and amended July 2012 (see Note); DOI: 10.2312/GFZ.NMSOP-2_IS_3.8 |

**Note:** Late Dr. G. Bock has been unable to upgrade his former section 3.5 in Chapter 3 of the first NMSOP (2002) edition. It has been reproduced here for the electronic edition of NMSOP-2 with some editorial changes in format and revised web addresses until being complemented by an expanded up-to-date Information Sheet on the topic by Prof. F. Krüger.

|                                                     | Page |
|-----------------------------------------------------|------|
| <b>1 Introduction</b>                               | 1    |
| <b>2 Basic relations</b>                            | 2    |
| <b>3 An inversion scheme in the time domain</b>     | 5    |
| <b>4 Decomposition of the moment tensor</b>         | 7    |
| <b>5 Steps taken in moment-tensor inversion</b>     | 9    |
| <b>6 Some methods of moment-tensor inversion</b>    | 10   |
| 6.1 NEIC fast moment tensors                        | 10   |
| 6.2 Harvard CMT solutions                           | 10   |
| 6.3 EMSC rapid source parameter determinations      | 11   |
| 6.4 Relative moment-tensor inversion                | 11   |
| 6.5 NEIC broadband depths and fault-plane solutions | 11   |
| <b>Acknowledgments</b>                              | 12   |
| <b>Recommended overview readings</b>                | 12   |
| <b>References</b>                                   | 13   |

## 1 Introduction

The concept of first order *moment tensor* provides a complete description of equivalent body forces of a general seismic point source (see Figure 1). A source can be considered a point source if both the distance  $D$  of the observer from the source and the wavelength  $\lambda$  of the data are much greater than the linear dimension of the source. Thus, moment-tensor solutions are generally derived from low-frequency data and they are representative of the gross properties of the rupture process averaged over tens of seconds or more. The double-couple source model describes the special case of shear dislocation along a planar fault. This model has proven to be very effective in explaining the amplitude and polarity pattern of P, S and surface waves radiated by tectonic earthquakes. In the following, we briefly outline the relevant relations (in a first order approximation) between the moment tensor of a seismic source and the observed seismogram. The latter may be either the complete seismogram, one of its main groups (P, S or surface waves), or specific features of seismograms such as peak-to-peak amplitudes of body waves, amplitude ratios or spectral amplitudes. Then we outline a linear inversion scheme for obtaining the moment tensor using waveform data in the time domain. Finally, we will give an overview of some useful programs for moment-tensor analysis. Applications of moment-tensor inversions to the rapid (i.e., generally within 24 hours after the event) determination of source parameters after significant earthquakes will also be described.



## 2 Basic relations

Following Jost and Herrmann (1989), the displacement  $d$  on the Earth's surface at a station can be expressed, in case of a point source, as a linear combination of time-dependent moment-tensor elements  $M_{kj}(\xi, t)$  that are assumed to have the same time dependence convolved (indicated by the star symbol) with the derivative  $G_{skj}(\mathbf{x}, \xi, t)$  of the Green's functions with regard to the spatial  $j$ -coordinate:

$$u_s(\mathbf{x}, t) = M_{kj}(\xi, t) * G_{skj}(\mathbf{x}, \xi, t). \quad (1)$$

$u_s(\mathbf{x}, t)$ :  $s$  component of ground displacement at position  $\mathbf{x}$  and time  $t$

$M_{kj}(\xi, t)$ : components of 2nd order, symmetrical seismic moment tensor  $M$

$G_{skj}(\mathbf{x}, \xi, t)$ : derivative of the Green's function with regard to source coordinate  $\xi_j$

$\mathbf{x}$ : position vector of station with coordinates  $x_1, x_2, x_3$  for north, east and down

$\xi$ : position vector of point source with coordinates  $\xi_1, \xi_2, \xi_3$  for north, east and down

Eq. (1) follows from the representation theorem in terms of the Green's function (see Equations (21) and (38) in IS 3.1). The Green's function represents the impulse response of the medium between source and receiver and thus contains the various wave propagation effects through the medium from source to receiver. These include energy losses through reflection and transmission at seismic discontinuities, anelastic absorption and geometrical spreading. The  $M_{kj}(\xi, t)$  from Eq.(1) completely describes the forces acting in the source and their time dependence. The Einstein summation notation is applied in Eq. (1) and below, i.e., the repeated indices  $k$  and  $j = 1, 2, 3$  imply summation over  $x_1, x_2$  and  $x_3$ . In Eq. (1) the higher order terms of the Taylor expansion around the source point of the Green's functions  $G_{skj}(\mathbf{x}, \xi, t)$  have been neglected. Note that the source-time history  $s(t)$  (see 3.1, Figs. 3.4 and 3.9), which describes the time dependence of moment released at the source, is contained in  $c$ . If we assume that all the components of  $M_{kj}(\xi, t)$  have the same time dependence  $s(t)$  the equation can be written as:

$$u_s(\mathbf{x}, t) = M_{kj} [G_{skj}(\mathbf{x}, \xi, t) * s(t)] \quad (2)$$

with  $s(t)$ : source time history.

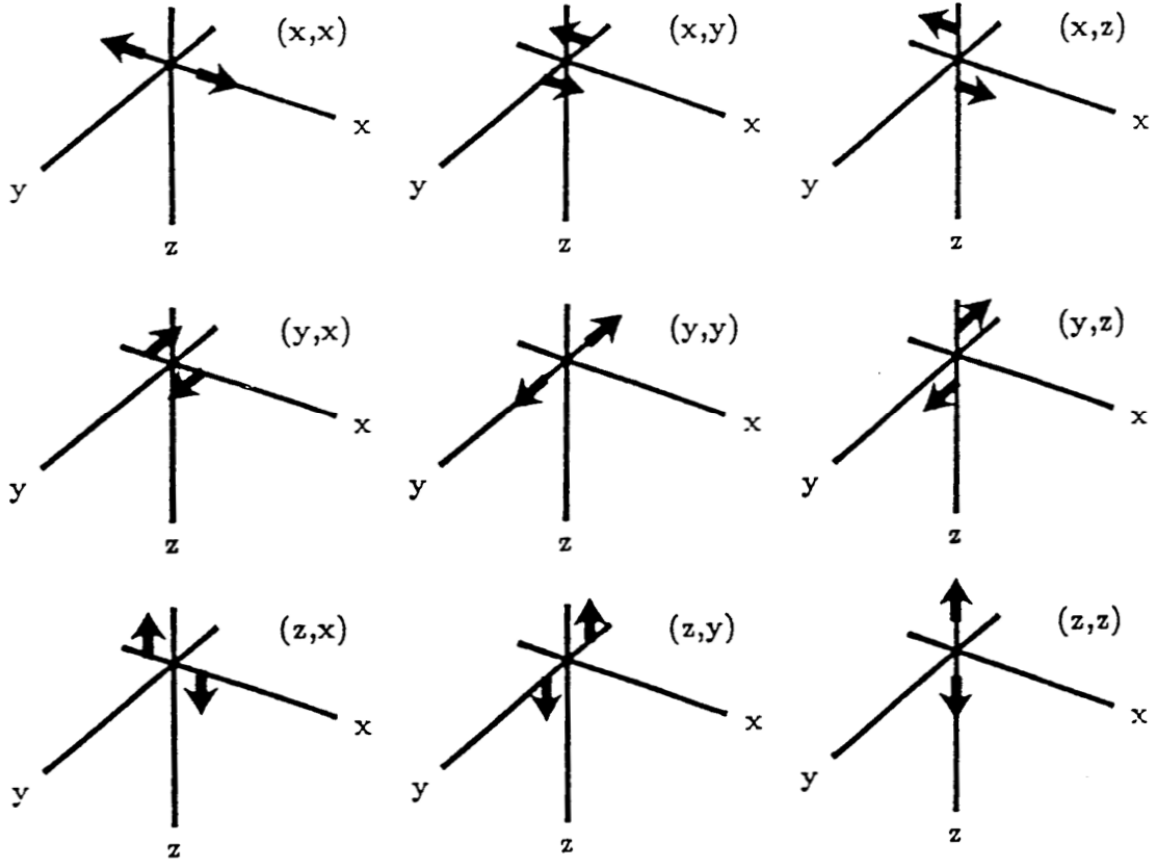
When determining  $M_{kj}(\xi, t)$  from seismic records,  $u_s(\mathbf{x}, t)$  is calculated by convolution of the observed seismogram components  $y_s(\mathbf{x}, t)$  with the inverse of the seismograph's displacement response function  $i(t)$ :

$$u_s(\mathbf{x}, t) = y_s(\mathbf{x}, t) * \text{Inv}\{i(t)\}$$

In the frequency domain (see Eq. (14) in IS 3.1) convolution is replaced by multiplication:

$$D_s(\mathbf{x}, \omega) = Y_s(\mathbf{x}, \omega) I(\omega)^{-1}$$

where  $\omega$  is circular frequency. The  $D_s(\mathbf{x}, \omega)$ ,  $Y_s(\mathbf{x}, \omega)$ , and  $I(\omega)^{-1}$  are the respective Fourier transforms of the time series  $d_s(\mathbf{x}, t)$ ,  $y_s(\mathbf{x}, t)$ , and  $i(t)^{-1}$  (see 5.2.7 where  $I(\omega)^{-1}$  is denoted as  $H_d(\omega)^{-1}$ ).



**Figure 1** The nine generalized couples representing  $G_{skj}(\mathbf{x}, \boldsymbol{\xi}, t)$  in Eq. (3.69). Note that force couples acting on the  $y$  axis in  $x$  direction or vice versa (i.e.,  $(x,y)$  or  $(y,x)$ ) will cause shear faulting in the  $x$  and  $y$  direction, respectively. Superimposition of vector dipoles in  $x$  and  $y$  direction with opposite sign, e.g.,  $(x,x) + (-y,-y)$  will also cause shear faulting but  $45^\circ$  off the  $x$  and  $y$  direction, respectively. Both representations are equivalent (reproduced from Jost and Herrmann, A student's guide to and review of moment tensors. *Seismol. Res. Lett.*, **60**, 2, 1989, Fig. 2, p. 39; ©Seismological Society of America).

In the following we assume that the source-time function  $s(t)$  is a delta function (i.e., a "needle" impulse). Then,  $M_{kj}(\boldsymbol{\xi}, t) = M_{kj}(\boldsymbol{\xi}) \cdot \delta(t)$ , and the right side of Eq.(2) simplifies to  $M_{kj}(\boldsymbol{\xi}) \cdot G_{skj}(t)$ . The seismogram recorded at  $\mathbf{x}$  can be regarded as product of  $G_{skj}$  and  $M_{kj}$ . (e.g., Aki and Richards, 1980 and 2002; Lay and Wallace, 1995; Udias, 1999). Thus, the derivative of  $G_{skj}$  with regard to the source coordinate  $\xi_i$  describes the response to a single couple with its lever arm pointing in the  $\xi_j$  direction (see Figure 1). For  $k = j$  we obtain a vector dipole; these are the couples  $(x,x)$ ,  $(y,y)$ , and  $(z,z)$  in Figure 1. A double-couple source is characterized by a moment tensor where one eigenvalue of the moment tensor vanishes (equivalent to the Null or B axis), and the sum of eigenvalues vanishes, i.e., the trace of the moment tensor is zero. Physically, this is a representation of a shear dislocation source without any volume changes.

Using the notation of Figure 1, double-couple displacement fields are represented by the sum of two couples such as  $(x,y) + (y,x)$ ,  $(x,x) + (y,y)$ ,  $(y,y) + (z,z)$ ,  $(y,z) + (z,y)$  etc. An explosion

source (corresponding to  $M_6$  in Eq. (8) and Figure 2) can be modelled by the sum of three vector dipoles  $(x,x) + (y,y) + (z,z)$ . A compensated linear vector dipole (CLVD, see section 4 below) can be represented by a vector dipole of strength 2 and two vector dipoles of unit strength but opposite sign in the two orthogonal directions.

The seismic moment tensor  $\mathbf{M}$  has, in general, six independent components which follows from the condition that the total angular momentum for the equivalent forces in the source must vanish. For vanishing trace, i.e., without volume change, we have five independent components that describe the deviatoric moment tensor. The double-couple source is a special case of the deviatoric moment tensor with the constraint that the determinant of  $\mathbf{M}$  is zero, i.e., that the stress field is two-dimensional.

In general,  $\mathbf{M}$  can be decomposed into an isotropic and a deviatoric part:

$$\mathbf{M} = \mathbf{M}^{\text{isotropic}} + \mathbf{M}^{\text{deviatoric}} \quad (3)$$

The decomposition of  $\mathbf{M}$  is unique while further decomposition of  $\mathbf{M}^{\text{deviatoric}}$  is not. Commonly,  $\mathbf{M}^{\text{deviatoric}}$  is decomposed into a double couple and CLVD:

$$\mathbf{M}^{\text{deviatoric}} = \mathbf{M}^{\text{DC}} + \mathbf{M}^{\text{CLVD}} \quad (4)$$

For a double-couple source, the Cartesian components of the moment tensor can be expressed in terms of strike  $\phi$ , dip  $\delta$  and rake  $\lambda$  of the shear dislocation source (fault plane), and the scalar seismic moment  $M_0$  (Aki and Richards, 1980):

$$\begin{aligned} M_{xx} &= -M_0(\sin\delta \cos\lambda \sin 2\phi + \sin 2\delta \sin\lambda \sin^2\phi) \\ M_{xy} &= M_0(\sin\delta \cos\lambda \cos 2\phi + 0.5 \sin 2\delta \sin\lambda \sin 2\phi) \\ M_{xz} &= -M_0(\cos\delta \cos\lambda \cos\phi + \cos 2\delta \sin\lambda \sin\phi) \\ M_{yy} &= M_0(\sin\delta \cos\lambda \sin 2\phi - \sin 2\delta \sin\lambda \cos^2\phi) \\ M_{yz} &= -M_0(\cos\delta \cos\lambda \sin\phi - \cos 2\delta \sin\lambda \cos\phi) \\ M_{zz} &= M_0 \sin 2\delta \sin\lambda \end{aligned} \quad (5)$$

As the tensor is always symmetric it can be rotated into a principal axis system such that all non-diagonal elements vanish and only the diagonal elements are non-zero. The diagonal elements are the *eigenvalues* (see Eq. (6) in Information Sheet 3.1) of  $\mathbf{M}$ ; the associated directions are the *eigenvectors* (i.e., the *principal axes*). A linear combination of the principal moment-tensor elements completely describes the radiation from a seismic source. In the case of a double-couple source, for example, the diagonal elements of  $\mathbf{M}$  in the principal axis system have two non-zero eigenvalues  $M_0$  and  $-M_0$  (with  $M_0$  the scalar seismic moment) whose eigenvectors give the direction at the source of the tensional (positive) T axis and compressional (negative) P axis, respectively, while the zero eigenvalue is in the direction of the B (or Null) axis of the double couple (for definition and determination of  $M_0$  see Exercise 3.4).

Eq. (2) describes the relation between seismic displacement and moment tensor in the time domain. If the source-time function is not known or the assumption of time-independent

moment-tensor elements is dropped, e.g., for reasons of source complexity, the frequency-domain approach is chosen:

$$u_s(x, f) = M_{kj}(f)G_{skj}(x, \xi, f) \quad (6)$$

where  $f$  denotes frequency. Procedures for the linear moment-tensor inversion can be designed in both the time and frequency domain using Eq. (2) or (6). We can write (2) or (6) in matrix form:

$$\mathbf{u} = \mathbf{G} \bar{\mathbf{m}}. \quad (7)$$

In the time domain, the  $\mathbf{u}$  is a vector containing  $n$  sampled values of observed ground displacement at various times, stations and sensor components, while  $\mathbf{G}$  is a  $6 \times n$  matrix and the vector  $\bar{\mathbf{m}}$  contains the six independent moment-tensor elements to be determined. In the frequency domain,  $\mathbf{u}$  contains  $k$  complex values of the displacement spectra determined for a given frequency  $f$  at various stations and sensor components.  $\mathbf{G}$  is a  $6 \times k$  matrix and is generally complex like  $\bar{\mathbf{m}}$ . For more details on the inversion problem in Eq. (7) the reader is referred to Chapter 6 in Lay and Wallace (1995), Chapter 12 in Aki and Richards (1980), or Chapter 19 of Udias (1999).

To invert Eq. (7) for the unknown  $\bar{\mathbf{m}}$ , one has to calculate the derivatives of the Green's functions. The calculation of the Green's functions constitutes the most important part of any moment-tensor inversion scheme. A variety of methods exists to calculate synthetic seismograms (e.g., Müller, 1985; Doornbos, 1988; Kennett, 1988). Some of the synthetic seismogram codes allow calculations for the moment-tensor elements as input source while others allow input for double-couple and explosive point sources. The general moment tensor can be decomposed in various ways using moment-tensor elements of double-couple and explosive sources so that synthetic seismogram codes employing these source parameterizations can also be used in the inversion of (7).

### 3 An inversion scheme in the time domain

In this section, we will describe in short the moment-tensor inversion algorithm of Kikuchi and Kanamori(1991), where the moment tensor is decomposed into elementary double-couple sources and an explosive source. Adopting the notation used by Kikuchi and Kanamori(1991), the moment tensor  $M_{kj}$  is represented by a linear combination of  $N_e = 6$  elementary moment tensors  $M_n$  (Figure 2):

$$M_{kj} = \sum_{n=1}^{N_e} a_n M_n \quad (8)$$

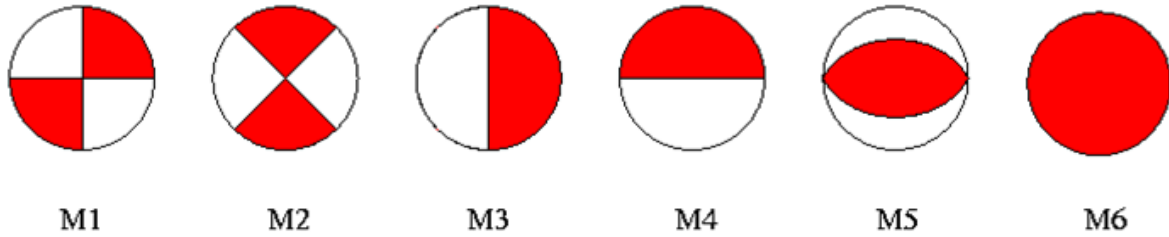
with

$$M_1 : \begin{bmatrix} 0 & 1 & 0 \\ 1 & 0 & 0 \\ 0 & 0 & 0 \end{bmatrix}; \quad M_2 : \begin{bmatrix} 1 & 0 & 0 \\ 0 & -1 & 0 \\ 0 & 0 & 0 \end{bmatrix}; \quad M_3 : \begin{bmatrix} 0 & 0 & 0 \\ 0 & 0 & 1 \\ 0 & 1 & 0 \end{bmatrix}$$

$$M_4: \begin{bmatrix} 0 & 0 & 1 \\ 0 & 0 & 0 \\ 1 & 0 & 0 \end{bmatrix}; M_5: \begin{bmatrix} -1 & 0 & 0 \\ 1 & 0 & 0 \\ 0 & 0 & 1 \end{bmatrix}; M_6: \begin{bmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{bmatrix}$$

The  $M_1$  and  $M_2$  represent pure strike-slip faults;  $M_3$  and  $M_4$  represent dip-slip faults on vertical planes striking N-S and E-W, respectively, and  $M_5$  represents a 45° dip-slip fault. The  $M_6$  represents an isotropic source radiating energy equally into all directions (i.e., an explosion).

## Elementary Moment Tensors



**Figure 2** Elementary moment tensors used in the inversion of the full moment tensor (after Kikuchi and Kanamori, 1991)

A pure deviatoric moment tensor ( $\text{trace}(M_{kj}) = 0$ ) is entirely represented by the five elementary moment tensors  $M_1$  to  $M_5$ . The following brief description of the linear inversion for the moment tensor (Kikuchi and Kanamori, 1991) is an example of an inversion performed in the time domain. It can be easily adopted for an inversion in the frequency domain by replacing the time series by their spectra. Let  $w_{sn}(t)$  denote the Green's function derivative at station  $s$  in response to the elementary moment *tensor*  $M_n$ , and let  $x_s(t)$  be the observed ground displacement as function of time at station  $s$ . The best estimate for the coefficients  $a_n$  in Eq. (8) can be obtained from the condition that the difference between observed and synthetic displacement functions be zero:

$$\begin{aligned} \Delta &= \sum_{s=1}^{N_s} \int \left[ x_s(t) - \sum_{n=1}^{N_e} a_n w_{sn}(t) \right]^2 dt \\ &= R_x - 2 \sum_{n=1}^{N_e} a_n G_n + \sum_{m=1}^{N_e} \sum_{n=1}^{N_e} R_{nm} a_n a_m \\ &= \text{Minimum} \end{aligned} \tag{9}$$

The  $N_e$  is the number of elementary moment tensors, and  $N_s$  is the number of displacement records used. The other terms in (9) are given by:

$$R_x = \sum_{s=1}^{N_s} \int [x_s(t)]^2 dt$$

$$R_{nm} = \sum_{s=1}^{N_s} \int [w_{sn}(t) w_{sm}(t)] dt$$

$$G_n = \sum_{s=1}^{N_s} \int [w_{sn}(t) x_s(t)] dt$$

Integration is carried out over selected portions of the waveforms. Evaluating  $\partial\Delta/\partial a_n = 0$  for  $n = 1, \dots, N_e$  yields the normal equations

$$\sum_{m=1}^{N_e} R_{nm} a_m = G_n \quad (10)$$

with  $n$  ranging from 1 to  $N_e$ . The solution for  $a_n$  is given by:

$$a_n = \sum_{m=1}^{N_e} R_{nm}^{-1} G_m \quad (11)$$

The inverse  $R_{nm}^{-1}$  of matrix  $R_{nm}$  can be obtained by the method of generalized least squares inversion (e.g., Pavlis, 1988). The resultant moment tensor is then given by

$$M_{kj} = \begin{bmatrix} a_2 - a_5 + a_6 & a_1 & a_4 \\ a_1 & -a_2 + a_6 & a_3 \\ a_4 & a_3 & a_5 + a_6 \end{bmatrix} \quad (12)$$

The variance of the elements  $a_n$  can be calculated under the assumption that the data are statistically independent:

$$\text{var}(a_n) = \sum_{m=1}^{N_e} (R_{nm}^{-1})^2 \sigma_m^2$$

where  $\sigma_m^2$  is the variance of the data  $G_n$ . In the case where the variance of the data is not known,  $\sum_{m=1}^{N_e} (R_{nm}^{-1})^2$  can be used as relative measure for the uncertainty.

#### 4 Decomposition of the moment tensor

Except for the volumetric and deviatoric components, the decomposition of the moment tensor is not unique. Useful computer programs for decomposition were written by Jost and distributed in Volume VIII of the Computer Programs in Seismology by Herrmann of Saint Louis University (<http://www.eas.slu.edu/People/RBHerrmann/ComputerPrograms.html>) or e-mail to R. W. Herrmann: [rbh@slueas.slu.edu](mailto:rbh@slueas.slu.edu)). The first step in the decomposition is the calculation of eigenvalues and eigenvectors of the seismic moment tensor. For this the

program *mteig* can be used. It performs rotation of the moment tensor  $\mathbf{M}$  into the principal axis system. The eigenvector of the largest eigenvalue gives the T (or tensional) axis; the eigenvector of the smallest eigenvalue gives the direction of the P (or compressional) axis, while the eigenvector associated with the intermediate eigenvalue gives the direction of the Null axis. The output of *mteig* is the diagonalized moment tensor

$$M = \begin{bmatrix} m_1 & 0 & 0 \\ 0 & m_2 & 0 \\ 0 & 0 & m_3 \end{bmatrix} \quad (13)$$

whose elements are input to another program, *mtdec*, which performs a moment-tensor decomposition. First, the moment tensor is decomposed into an isotropic and a deviatoric part (see Eq. 3):

$$M = \frac{1}{3} \begin{bmatrix} tr(M) & 0 & 0 \\ 0 & tr(M) & 0 \\ 0 & 0 & tr(M) \end{bmatrix} + \begin{bmatrix} m_1^1 & 0 & 0 \\ 0 & m_2^1 & 0 \\ 0 & 0 & m_3^1 \end{bmatrix} \quad (14)$$

with  $tr(\mathbf{M}) = m_1 + m_2 + m_3$  being the trace of  $\mathbf{M}$ . The isotropic part of  $\mathbf{M}$  is important in quantifying volume changes of the source, but it is usually difficult to resolve so that isotropic parts of less than 10% are often not considered to be significant. The deviatoric part of the moment tensor can be further decomposed. Options include decompositions into three vector dipoles, into three double couples, into 3 CLVD sources, into a major and minor double couple, and into a best double couple and a CLVD having the same principal axis system. The source mechanisms reported by Harvard and USGS are based on the decomposition of the moment tensor into a best double couple and a CLVD. In addition to the best double couple they also provide the moment-tensor elements. To estimate the double-couple contribution to the deviatoric moment tensor, the parameter

$$\varepsilon = \frac{|m_{\min}|}{|m_{\max}|}$$

is used (Dziewonski et al., 1981) where  $m_{\min}$  and  $m_{\max}$  are the smallest and largest eigenvalues of the deviatoric part of  $\mathbf{M}$ , respectively, both in absolute terms. For a pure double-couple source,  $\varepsilon = 0$  because  $m_{\min} = 0$ ; for a pure CLVD,  $\varepsilon = 0.5$ . The percentage double-couple contribution can be expressed as  $(1-2\varepsilon) \times 100$ . Significant CLVD components are often reported for large intermediate-depth and very deep earthquakes. In many cases, however, it can be shown that these are caused by superposition of several rupture events with different double-couple mechanisms (Kuge and Kawakatsu, 1990; Frohlich, 1995; Tibi et al., 1999).

Harvard and USGS publish the moment tensors using the notation of normal mode theory. It is based on spherical co-ordinates  $(r; \Theta; \Phi)$  where  $r$  is the radial distance of the source from the

center of the Earth,  $\Theta$  is co-latitude, and  $\Phi$  is longitude of the point source. The 6 independent moment-tensor elements in the  $(x, y, z) = (\text{north, east, down})$  coordinate system are related to the components in  $(r; \Theta; \Phi)$  by

$$M_{rr} = M_{zz}$$

$$M_{\Theta\Theta} = M_{xx}$$

$$M_{\Phi\Phi} = M_{yy}$$

$$M_{r\Theta} = M_{zx}$$

$$M_{r\Phi} = -M_{zy}$$

$$M_{\Theta\Phi} = -M_{xy}$$

## 5 Steps taken in moment-tensor inversion

Generally, the quality of moment-tensor inversion depends to a large extent on the number of data available and the azimuthal distribution of stations about the source. Dufumier (1996) gives a systematic overview for the effects caused by differences in the azimuthal coverage and the effects caused due to the use of only P waves, P plus SH waves or P and SH and SV waves for the inversion with body waves.

A systematic overview with respect to the effects caused by an erroneous velocity model for the Green function calculation and the effects due to wrong hypocenter coordinates can be found in Šílený et al. (1992), Šílený and Pšenčík (1995), Šílený et al. (1996) and Kravanja et al. (1999).

The following is a general outline of the various steps to be taken in a moment-tensor inversion using waveform data:

### 1) Data acquisition and pre-processing

- good signal-to-noise ratio
- unclipped signals
- good azimuthal coverage
- removing mean value and linear trends
- correcting for instrument response, converting seismograms to displacement
- low-pass filtering to remove high-frequency noise and to satisfy the point source approximation

### 2) Calculation of synthetic Green's functions dependent on

- Earth model
- location of the source
- receiver position

### 3) Inversion



- selection of waveforms, e.g., P, S H or full seismograms
- taking care to match waveforms with corresponding synthetics
- evaluation of Eqs. (8) and (9)
- decomposition of moment tensor, e.g., into best double couple plus CLVD

The inversion may be done in the time domain or frequency domain. Care must be taken to match the synthetic and observed seismograms. Alignment of observed and synthetic waveforms is facilitated by cross-correlation techniques. In most moment-tensor inversion schemes, focal depth is assumed to be constant. The inversion is done for a range of focal depths and as best solution one takes that with the minimum variance of the estimate.

## 6 Some methods of moment-tensor inversion

### 6.1 NEIC fast moment tensors

This is an effort by the U.S. National Earthquake Information Center (NEIC) in co-operation with the IRIS Data Management Center to produce rapid estimates of the seismic moment tensor for earthquakes with body-wave magnitudes  $\geq 5.8$ . Digital waveform data are quickly retrieved from “open” IRIS stations and transmitted to NEIC by Internet. These data contain teleseismic P waveforms that are used to compute a seismic moment tensor using a technique based on optimal filter design (Sipkin, 1982). Near real-time Current Fast Moment Tensor Solutions are available via <http://earthquake.usgs.gov/earthquakes/eqarchives/fm/>. One can also subscribe via <https://ssleearthquake.usgs.gov/ens/> to a free Earthquake Notification Service (ENS) that sends automated notification E-mails when earthquakes happen in the area of interest to the subscriber

### 6.2 Harvard and Lamont CMT solutions

The Harvard group developed and maintained an extensive catalog of centroid moment-tensor (CMT) solutions for strong (mainly  $M > 5.5$ ) earthquakes over the period from 1976-2006. Since then it is maintained and continued by the Global Centroid Moment Tensor Project (GCMT) at the Lamont-Doherty Earth Observatory (LDEO) of Columbia University. The main dissemination point for information and results from this project (e.g., description of the CMT procedure, global CMT Catalog Search, CMT catalog and quick CMT ACII files, special studies of particular earthquakes or sets of earthquakes) is now via the website <http://www.globalcmt.org>. The Harvard CMT method makes use of both very long-period ( $T > 40$  s) body waves (from the P wave onset until the onset of the fundamental modes) and so-called mantle waves at  $T > 135$  s that comprise the complete surface-wave train. Starting with earthquakes in 2004, the GCMT analysis includes now also intermediate-period surface waves in the moment-tensor inversion. This allows the globally uniform determination of moment tensors for earthquakes down to  $M_w = 5.0$  (Ekström et al., 2012).

Besides the moment tensor the iterative inversion procedure seeks a solution for the best point source location of the earthquake. This is the point where the system of couples is located in the source model described by the moment tensor. It represents the integral of the moment density over the extended rupture area. This centroid location may, for very large earthquakes, significantly differ from the hypocenter location based on arrival times of the first P-wave onsets. The hypocenter location corresponds to the place where rupture started. Therefore, the

offset of the centroid location relative to the hypocentral location gives a first indication on fault extent and rupture directivity. In case of the August 17, 1999 Izmit (Turkey) earthquake the centroid was located about 50 km east of the "P-wave" hypocenter. The centroid location coincided with the area where the maximum surface ruptures were observed.

### 6.3 EMSC and GFZ rapid source parameter determinations

This is an initiative of the European-Mediterranean Seismological Center (Bruyeres-le-Chatel, France, <http://www.emsc-csem.org/>) and the GEOFON Programs at the GeoForschungs-Zentrum Potsdam (<http://www.gfz-potsdam.de/geofon/>). The EMSC method uses a grid search algorithm to derive the fault-plane solutions and seismic moments of earthquakes ( $M > 5.5$ ) in the European-Mediterranean area. Solutions are derived within 24 hours after the occurrence of the event. The data used are P- and S-wave amplitudes and polarities. Figure 3 shows one of the early examples of the kind of output data produced. Nowadays near real-time CMT solutions also for earthquakes world-wide can be obtained via <http://geofon.gfz-potsdam.de/eqinfo/> and one may also **subscribe** receiving automatic earthquake notification e-mails by filling-in the Alert Mailing List Registration form of the GEOFON Rapid Earthquake Information Service (see link via <http://www.gfz-potsdam.de/geofon/>).

### 6.4 Relative moment-tensor inversion

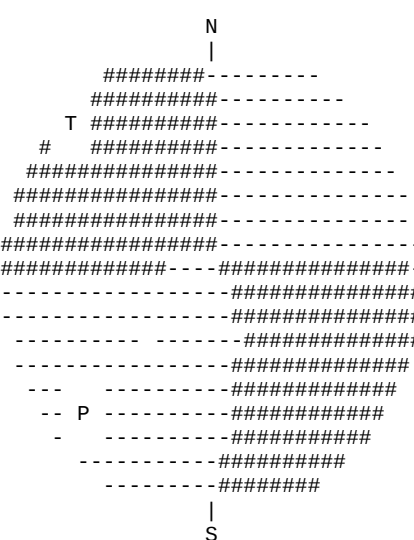
Especially for the inversion of local events so called relative moment-tensor inversion schemes have been developed (Oncescu, 1986; Dahm, 1996). If the sources are separated by not more than a wavelength, the Green's functions can be assumed to be equal with negligible error. In this case it is easy to construct a linear equation system that relates the moment-tensor components of a reference event to those of another nearby event. This avoids the calculation of high-frequency Green's functions necessary for small local events and all problems connected with that (especially the necessity of modeling site transfer functions in detail).

This is a very useful scheme for the analysis of aftershocks if a well determined moment tensor of the main shock is known. Moreover, if enough events with at least slightly different mechanisms and enough recordings are available, it is also possible to eliminate the reference mechanism from the equations (Dahm, 1996). This is interesting for volcanic areas where events are swarm-like and of similar magnitude, and where a reference moment tensor can not be provided (Dahm and Brandsdottir, 1997).

### 6.5 NEIC broadband depths and fault-plane solutions

Moment-tensor solutions, which are generally derived from low-frequency data, reflect the gross properties of the rupture process averaged over tens of seconds or more. These solutions may differ from solutions derived from high frequency data, which are more sensitive to the dynamic part of the rupture process during which most of the seismic energy is radiated. For this reason, beginning January 1996, the NEIC has determined, whenever possible, a fault plane solution and depth from broadband body waves for any earthquake having a magnitude greater than about 5.8 and it has published the source parameters in the Monthly Listings of

the PDE. The broadband waveforms that are used have a flat displacement response over the frequency range 0.01-5.0 Hz. (This bandwidth, incidentally, is also commensurate with that used by the NEIC to compute teleseismic  $E_s$ .) Initial constraints on focal mechanism are provided by polarities from P, pP and PKP waves, as well as by Hilbert-transformed body waves of certain secondary arrivals (e.g., PP), and from transversely polarized S waves. The fault-plane solution and depth are then refined by least-squares fitting of synthetic waveforms to teleseismically recorded P-wave groups (consisting of direct P, pP and sP). More information can be found under [http://neic.usgs.gov/neis/nrg/bb\\_processing.html](http://neic.usgs.gov/neis/nrg/bb_processing.html).

|                                                                                                                                                                                               |              |                |                |                 |                                                                                                                                                                                                       |  |  |  |  |
|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--------------|----------------|----------------|-----------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--|--|--|--|
| <b>European-Mediterranean Seismological Centre</b><br><b>Centre Sismologique Euro-Mediterraneen</b>                                                                                           |              |                |                |                 |                                                                                                                                                                                                       |  |  |  |  |
| Double-couple solution provided by GFZ Potsdam                                                                                                                                                |              |                |                |                 | Corner frequencies of bandpass filter: 0.020 and 0.100 Hz                                                                                                                                             |  |  |  |  |
| EMSC event parameters: 21-JUN-2000_00:51:46.6<br>63.88 N 20.69 W (Iceland)<br>Depth = 10 km (adopted in inversion)<br>Depth = 9 km (based on 32 depth phases)                                 |              |                |                |                 | First fault plane: Strike = 358 degrees<br>Rake = 185 degrees<br>Dip = 85 degrees                                                                                                                     |  |  |  |  |
| 32 stations used in inversion:                                                                                                                                                                |              |                |                |                 | Second fault plane: Strike = 268 degrees<br>Rake = -5 degrees<br>Dip = 85 degrees                                                                                                                     |  |  |  |  |
| <b>Station</b>                                                                                                                                                                                | <b>Delta</b> | <b>Azimuth</b> | <b>Takeoff</b> | <b>Polarity</b> | $M_0 = (4.3 \pm 2.1) \cdot 10^{18} \text{ N}\cdot\text{m}$<br>$M_w = 6.4$                                                                                                                             |  |  |  |  |
| adk                                                                                                                                                                                           | 63.06        | 343.55         | 20.3           | C               | Source duration = 4 s (from BB displacement seismograms)                                                                                                                                              |  |  |  |  |
| aqu                                                                                                                                                                                           | 29.12        | 121.60         | 27.6           | C               | <b>Principal axes</b> <b>Trend</b> <b>Plunge</b>                                                                                                                                                      |  |  |  |  |
| biny                                                                                                                                                                                          | 38.06        | 262.06         | 26.1           | x               | -----                                                                                                                                                                                                 |  |  |  |  |
| brg                                                                                                                                                                                           | 22.41        | 109.49         | 33.8           | C               | P                      223                      7                                                                                                                                                     |  |  |  |  |
| cart                                                                                                                                                                                          | 28.87        | 146.49         | 27.7           | C               | N                      43                      83                                                                                                                                                     |  |  |  |  |
| cmb                                                                                                                                                                                           | 60.57        | 296.59         | 20.9           | C               | T                      313                      0                                                                                                                                                     |  |  |  |  |
| cm1a                                                                                                                                                                                          | 26.31        | 188.62         | 28.2           | D               |                                                                                                                                                                                                       |  |  |  |  |
| cor                                                                                                                                                                                           | 56.07        | 302.72         | 22.0           | C               |                                                                                                                                                                                                       |  |  |  |  |
| css                                                                                                                                                                                           | 43.52        | 105.32         | 24.9           | C               |                                                                                                                                                                                                       |  |  |  |  |
| dug                                                                                                                                                                                           | 55.68        | 291.95         | 22.1           | C               |                                                                                                                                                                                                       |  |  |  |  |
| eil                                                                                                                                                                                           | 48.77        | 107.33         | 23.7           | C               |                                                                                                                                                                                                       |  |  |  |  |
| ffc                                                                                                                                                                                           | 39.71        | 296.00         | 25.8           | C               |                                                                                                                                                                                                       |  |  |  |  |
| furi                                                                                                                                                                                          | 68.86        | 114.32         | 19.0           | C               |                                                                                                                                                                                                       |  |  |  |  |
| hgn                                                                                                                                                                                           | 19.27        | 120.81         | 34.9           | C               |                                                                                                                                                                                                       |  |  |  |  |
| incn                                                                                                                                                                                          | 75.70        | 26.33          | 17.4           | D               |                                                                                                                                                                                                       |  |  |  |  |
| kev                                                                                                                                                                                           | 19.21        | 51.83          | 34.9           | D               |                                                                                                                                                                                                       |  |  |  |  |
| kmbo                                                                                                                                                                                          | 77.57        | 119.84         | 17.0           | C               |                                                                                                                                                                                                       |  |  |  |  |
| kbs                                                                                                                                                                                           | 17.85        | 20.07          | 40.1           | D               |                                                                                                                                                                                                       |  |  |  |  |
| kwp                                                                                                                                                                                           | 27.08        | 101.36         | 28.0           | C               |                                                                                                                                                                                                       |  |  |  |  |
| morc                                                                                                                                                                                          | 24.75        | 106.92         | 28.4           | C               |                                                                                                                                                                                                       |  |  |  |  |
| mrni                                                                                                                                                                                          | 46.08        | 104.64         | 24.3           | C               |                                                                                                                                                                                                       |  |  |  |  |
| mte                                                                                                                                                                                           | 24.76        | 155.63         | 28.4           | C               |                                                                                                                                                                                                       |  |  |  |  |
| pas                                                                                                                                                                                           | 63.05        | 292.64         | 20.3           | C               |                                                                                                                                                                                                       |  |  |  |  |
| pet                                                                                                                                                                                           | 58.94        | 331.77         | 21.3           | C               |                                                                                                                                                                                                       |  |  |  |  |
| rgn                                                                                                                                                                                           | 19.51        | 103.10         | 34.8           | C               |                                                                                                                                                                                                       |  |  |  |  |
| selv                                                                                                                                                                                          | 28.58        | 151.01         | 27.7           | C               |                                                                                                                                                                                                       |  |  |  |  |
| sfuc                                                                                                                                                                                          | 28.62        | 155.24         | 27.7           | C               |                                                                                                                                                                                                       |  |  |  |  |
| sjg                                                                                                                                                                                           | 55.09        | 235.67         | 22.2           | D               |                                                                                                                                                                                                       |  |  |  |  |
| sspa                                                                                                                                                                                          | 40.16        | 262.47         | 25.7           | D               |                                                                                                                                                                                                       |  |  |  |  |
| suw                                                                                                                                                                                           | 24.19        | 93.73          | 28.5           | C               |                                                                                                                                                                                                       |  |  |  |  |
| tns                                                                                                                                                                                           | 20.68        | 118.00         | 34.5           | C               |                                                                                                                                                                                                       |  |  |  |  |
| tuc                                                                                                                                                                                           | 61.55        | 285.54         | 20.7           | C               |                                                                                                                                                                                                       |  |  |  |  |
| Data provided by:<br>IRIS/USGS, MedNet, USNSN, GRSN, UCM/SFO/GEOFON,<br>IRIS/IDA, GEOFON, GII/GEOFON, KNMI, IRIS/GEOFON,<br>IRIS/AWI/GEOFON, TERASCOPE, GRSN/GEOFON, IAG,<br>GTSN, U. Arizona |              |                |                |                 |                                                                                                                   |  |  |  |  |
|                                                                                                                                                                                               |              |                |                |                 | Done by G. Bock, GeoForschungsZentrum Potsdam.<br>Visit the GFZ-EMSC web page under <a href="http://www.gfz-potsdam.de/pb2/pb24/emsc/emsc.html">http://www.gfz-potsdam.de/pb2/pb24/emsc/emsc.html</a> |  |  |  |  |

**Figure 3** Example of output data produced by the routine procedure for rapid EMSC source parameter determinations by the GEOFON group at the GFZ Potsdam.

## Acknowledgments

The author acknowledges with thanks helpful reviews by A. Udias and W. Brüstle.

## Recommended overview readings

Aki and Richards (1980 and 2002)  
 Ben Menahem and Singh (2000)  
 Das and Kostrov (1988)  
 Lay and Wallace (1995)  
 Udias (1999)

## References

- Aki, K., and Richards, P. G. (1980). Quantitative seismology. *Freeman*, San Francisco, Vol. I and II, 932 pp.
- Aki, K., and Richards, P. G. (2002). Quantitative seismology. Second Edition, ISBN 0-935702-96-2, *University Science Books*, Sausalito, CA., xvii + 700 pp.
- Ben-Menahem, A. and S.J. Singh (2000). *Seismic Waves and Sources*. 2<sup>nd</sup> edition, Dover Publications, New York.
- Dahm, T. (1996). Relative moment tensor inversion based on ray theory: theory and synthetic tests. *Geophys. J. Int.*, **124**, 245-257.
- Dahm, T., and Brandsdottir, B. (1997). Moment tensors of micro-earthquakes from the Eyjafjallosjokull volcano in South Iceland, *Geophys. J. Int.*, **130**, 183-192.
- Das, S., and Kostrov, B. V. (1988). Principles of earthquake source mechanics. *Cambridge University Press*.
- Doornbos, D. J. (Ed.) (1988a). Seismological algorithms, Computational methods and computer programs. *Academic Press*, New York, xvii + 469 pp.
- Dufumier, H. (1996). On the limits of linear moment tensor inversion of teleseismic body wave spectra. *Pageoph*, **147**, 467-482.
- Dziewonski, A. M., Chou, T.-A., and Woodhouse, J. H. (1981). Determination of earthquake source parameters from waveform data for studies of global and regional seismicity. *J. Geophys. Res.*, **86**, 2825-2852.
- Ekström, G., Nettles, M., and Dziewonski, A. M. (2012). The global CMT project 2004-2010: Centroid-moment tensors for 13,017 earthquakes. *Phys. Earth Planet Int.*, **200-201**, 1-9.
- Frohlich, C. (1995). Characteristics of well-determined non-double-couple earthquakes in the Harvard CMT catalog. *Phys. Earth Planet. Inter.*, **91**, 213-228.
- Kennett, B. L. N. (1988). Systematic approximations to the seismic wavefields, In: Doornbos (1988a), 237-259.
- Kikuchi, M., and Kanamori, H. (1991). Inversion of complex body waves – III. *Bull. Seism. Soc. Am.*, **81**, 2335-2350.
- Kravanja, S., Panza, G. F., and Šilený, J. (1999). Robust retrieval of a seismic point-source function. *Geophys. J. Int.*, **136**, 385-394.

- Kuge, K., and Kawakatsu, H. (1990). Analysis of a deep “non double couple” earthquake using very broadband data, *Geophys. Res. Lett.*, **17**, 227-230.
- Lay, T., and Wallace, T. C. (1995). *Modern global seismology*. ISBN 0-12-732870-X, Academic Press, 521 pp.
- Müller, G. (1985). The reflectivity method: A tutorial. *J. Geophys.*, **58**, 153-174.
- Oncescu, M. C. (1986). Relative seismic moment tensor determination for Vrancea intermediate depth earthquakes, *Pageoph*, **124**, 931-940.
- Šilený, J., Panza, G. F., and Campus, P. (1992). Waveform inversion for point source moment retrieval with variable hypocentral depth and structural model. *Geophys. J. Int.*, **109**, 259-274.
- Šilený, J., and Pšenčík, I. (1995). Mechanisms of local earthquakes in 3-D inhomogeneous media determined by waveform inversion. *Geophys. J. Int.*, **121**, 459-474.
- Šilený, J., Campus, P., and Panza, G. F. (1996). Seismic moment tensor resolution by waveform inversion of a few local noisy records - I. Synthetic tests. *Geophys. J. Int.*, **126**, 605-619.
- Sipkin, S. A. (1982). Estimation of earthquake source parameters by the inversion of waveform data: synthetic waveforms. *Phys. Earth Planet. Inter.*, **30**, 242-259.
- Tibi, R., Bock, G., Xia, Y., Baumbach, M., Grosser, H., Milkereit, C., Karakisa, S., Zünbül, S., Kind, R., and Zschau, J. (2001). Rupture processes of the 1999 August 17 Izmit, and November 12 Düzce (Turkey) earthquakes. *Geophys. J. Int.*, **144**, F1-F7.
- Udías, A. (1999). *Principles of Seismology*. Cambridge University Press, United Kingdom, 475 pp.

| Topic   | Moment tensor inversion and moment tensor interpretation                                                                                                                                                                                                                                                                                                                                                                                                                                         |
|---------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p><b>Torsten Dahm</b>, Helmholtz Centre Potsdam GFZ German Research Centre for Geosciences and University of Potsdam, Institute of Earth and Environmental Sciences, Germany; E-mail: <a href="mailto:torsten.dahm@gfz-potsdam.de">torsten.dahm@gfz-potsdam.de</a></p> <p><b>Frank Krüger</b>, University of Potsdam, Institute of Earth and Environmental Sciences, 14476 Potsdam, Germany; E-mail: <a href="mailto:Frank.Krueger@geo.uni-potsdam.de">Frank.Krueger@geo.uni-potsdam.de</a></p> |
| Version | January 2014; DOI:10.2312/GFZ.NMSOP-2_IS_3.9                                                                                                                                                                                                                                                                                                                                                                                                                                                     |

|                                                                                      | page |
|--------------------------------------------------------------------------------------|------|
| <b>1 Introduction</b>                                                                | 1    |
| 1.1 Nomenclature and terms                                                           | 2    |
| <b>2 Moment tensors: Interpretation and decomposition</b>                            | 3    |
| 2.1 Elementary physical sources                                                      | 5    |
| 2.1.1 The isotropic source (explosion and implosion)                                 | 5    |
| 2.1.2 The general dislocation source                                                 | 5    |
| 2.1.3 The shear dislocation source                                                   | 6    |
| 2.1.4 The tensile crack                                                              | 7    |
| 2.2 Decomposition into elementary sources                                            | 7    |
| 2.2.1 Decomposition into strike slip, dip slip DC sources and a vertical CLVD        | 8    |
| 2.2.2 Decomposition into a best double couple and a CLVD                             | 9    |
| 2.2.3 Decomposition into an isotropic source and a mixed-mode shear-tensile crack    | 10   |
| 2.3 Computer programs for decomposition                                              | 10   |
| 2.4 Displaying different moment tensor components                                    | 12   |
| <b>3 Moment tensor representations: basic relations</b>                              | 13   |
| 3.1 Resolving the time dependency of moment tensors                                  | 16   |
| 3.2 Representation of the centroid location (centroid moment tensor inversion)       | 18   |
| 3.3 Representation in a layered Earth model                                          | 20   |
| 3.4 Representation for body waves (peak amplitudes and amplitude ratios)             | 22   |
| 3.5 Relative moment tensor inversion using multiple events (body wave method)        | 24   |
| 3.6 Full waveform method using amplitude spectra                                     | 24   |
| 3.7 Moment tensor representation using eigen-vibrations and eigen-modes of the Earth | 25   |
| 3.8 Codes to calculate Green functions for moment tensor inversion                   | 26   |
| <b>4 Moment tensor inversions: schemes in practice</b>                               | 27   |
| <b>5 Free packages for moment tensor inversion</b>                                   | 28   |
| <b>6 Moment tensor catalogues</b>                                                    | 30   |
| <b>7 Acknowledgments</b>                                                             | 33   |
| <b>8 References</b>                                                                  | 33   |

## 1 Introduction

The following brief guide to moment tensor inversion discusses the characteristics of moment tensors, its physical interpretation and the different ways to decompose moment tensors. It reviews the basic equations used in different inversion schemes, clarifies the role of Green's

functions, and summarizes different practical aspects of moment tensor inversions for different types of waves and different distance ranges (local, regional and global earthquakes). Typical resolution problems of moment tensor components are discussed. The IS provides some guidelines to start moment tensor studies, and thereby introduces a modern, python based moment tensor toolbox (Kiwi Tools) that is flexible to adapt to all specific source studies and incorporates pre-calculated Green's function databases. The complementing exercise EX 3.6 provides a practical on moment tensor inversion using the Kiwi Tools. The IS does not derive the fundamental concepts and the theory of source representations, since this is given in chapter 3 and IS3.1 of the NMSOP-2.

The material is based in large parts on a review paper by Jost and Herrmann (1989) and on lecture notes and practicals we developed during several training courses on moment tensors between 2000 and 2012 (e.g. ESC and IUGG training courses, ICTP workshops in Trieste, Winterschool Sudelfeld). The material updates and extends IS3.8 in NMSOP by Günther Bock on moment tensor determination and decomposition.

### 1.1 Nomenclature and terms

The following nomenclature is used (unit in parentheses):

**Table 1:** Nomenclature

| Parameter                                       | Explanation                                                        | Type                                                      |
|-------------------------------------------------|--------------------------------------------------------------------|-----------------------------------------------------------|
| $t$ and $\tau$                                  | Time, referenced to receiver and source, respectively              | Scalars (s)                                               |
| $\omega$                                        | Angular frequency                                                  | Scalar (rad)                                              |
| $\mathbf{x}$ and $\xi$                          | Spatial coordinates of source and receivers                        | Vectors (m)                                               |
| $\mathbf{M}$ and $\mathbf{m}$                   | Moment tensor and moment tensor density                            | 2 <sup>nd</sup> rank tensors, (Nm and Nm/m <sup>2</sup> ) |
| $\mathbf{G}$                                    | Green's tensor                                                     | 3 <sup>rd</sup> rank tensor (m)                           |
| $\mathbf{u}$ and $\mathbf{D}$                   | Displacement and dislocation across fault                          | Vectors (m)                                               |
| $\mathbf{c}$                                    | Elasticity tensor                                                  | 4 <sup>th</sup> rank tensor (Pa)                          |
| $A$ and $V$                                     | Surface and volume of source                                       | Scalars (m <sup>2</sup> and m <sup>3</sup> )              |
| $\phi, \varphi, \lambda, \delta, \alpha, \beta$ | Angles                                                             | Scalars (rad)                                             |
| $\delta, \delta$ or $\delta_{ij}$               | Kronecker symbol, see e.g. IS3.1                                   | Scalar or 2 <sup>nd</sup> rank tensor                     |
| $N, \eta, K$                                    | 1 <sup>st</sup> and 2 <sup>nd</sup> Lamé's constant and bulk modul | Scalars (Pa)                                              |
| $\mathbf{F}$ and $\mathbf{f}$                   | Point force and force density                                      | Vector (N or N/m <sup>3</sup> )                           |
| $r$                                             | Distance                                                           | Scalar (m)                                                |
| $\gamma$                                        | Direction cosine                                                   | Vector (rad)                                              |
| $\mathbf{n}, \mathbf{l}$                        | Unit vectors (e.g. fault normal or slip direction)                 |                                                           |
| $\mathbf{s}$                                    | Slowness vector                                                    | Vector (s/m)                                              |
| $\mathbf{e}$                                    | Eigenvalue vector of moment tensor                                 | Vector (Nm)                                               |
| $h(t)$                                          | Source time function of the point source                           | Dimensionless                                             |

We use bold face letters (e.g.  $\mathbf{x}$ ) to express vectors and tensors. Scalars are denoted in normal (not bold face) letters. Components of vectors and tensors are denoted by  $x_i$  and  $M_{ij}$ , respectively, where indices usually vary between 1, 2, and 3. We further apply the Einstein summation convention, meaning that repeating indices in one term imply summation over the index, i.e.  $e_k n_k = \sum e_k n_k = e_1 n_1 + e_2 n_2 + e_3 n_3$ . Indices separated by comma “,” denote partial

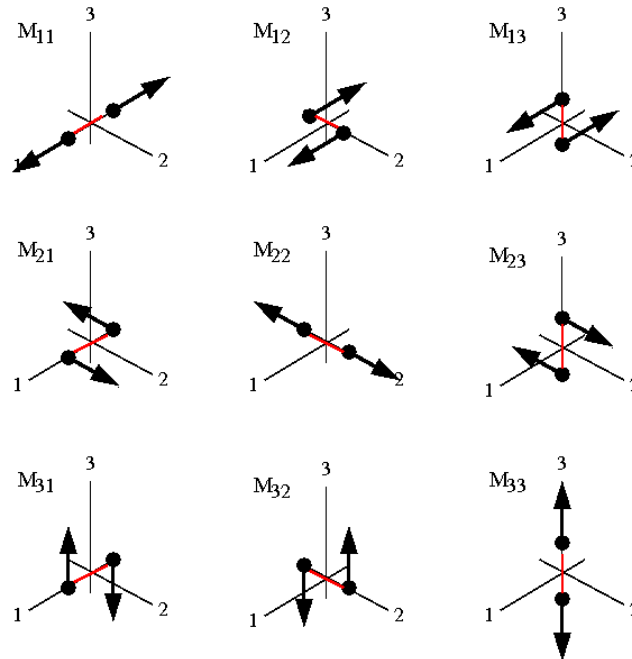
derivatives to space coordinates, i.e.  $G_{ij,k} = \partial G_{ij} / \partial x_k$ . Note that the summation convention will also apply to terms like  $M_{kk}$  or  $M_{kj} G_{ik,j}$ . The \* symbol denotes a **temporal convolution integral** and we partly drop the implicit time variable in the convolution formulas, e.g.,

$$\int_{\tau=-\infty}^t M(\xi, \tau) G(x, \xi, t - \tau) d\tau = M(\xi, t) * G(x, \xi, t) = M(\xi) * G(x, \xi)$$

## 2 Moment tensors: interpretation and decomposition

Moment tensors provide a general theoretical framework to describe seismic sources based on generalized force couples (Fig. 1). The moment tensor description is not restricted to earthquake sources, but covers also other types of seismic sources such as explosions, implosions, rock falls, landslides, meteorite terminal explosions (e.g. atmospheric), and mixed mode ruptures driven by fluid and gas injections. Thus, the concept of moment tensors is quite general and flexible making moment tensor inversions a very important tool in seismic source characterization. Moment tensors have the potential to substitute other, more traditional, source parameter estimations, such as e.g. magnitude or focal solutions from first motion polarities.

We denote a point source moment tensor by  $\mathbf{M}$ , a moment tensor density by  $\mathbf{m}$ . A moment tensor  $\mathbf{M}$  defines the strength of a seismic source in terms of its seismic moment, usually denoted by the scalar quantity  $M_0$ , and the radiation pattern of seismic waves. The moment tensor (3x3 matrix) is symmetric, i.e. it has six independent components. The diagonal elements represent linear vector dipoles, the off diagonal elements represent the force couples with arms (moment) (Fig. 1).



**Figure 1** The system of force couples representing the components of a Cartesian moment tensor. Diagonal elements of the moment tensor represent linear vector dipoles, while off-diagonal elements represent force couples with moment.



The moment tensor has components  $M_{ij}$  where  $i, j = 1, 2$ , or  $3$ . Often, a local geographic coordinate system is used to define a Cartesian system tensor, with positive  $x_1$  being north-, positive  $x_2$  eastward, and positive  $x_3$  being downward (NED). The NED coordinate system is used below to derive relations between the angles of a ruptured fault (strike and dip angle) and the dislocation direction on the rupture plane (rake angle).

A system commonly used in free-oscillation analysis is a local  $r$ - $\theta$ - $\phi$ -system, with  $r$ ,  $\theta$ ,  $\phi$  pointing upward, southward, and eastward (USE), respectively. This system is used for the routinely reported Global Centroid Moment tensors (Global CMT, formerly Harvard CMT). The geographic Cartesian tensor (NED) transforms by:

$$\begin{aligned} M_{rr} &= +M_{zz}, & M_{\theta\theta} &= +M_{nn}, & M_{\phi\phi} &= +M_{ee}, \\ M_{r\theta} &= +M_{nz}, & M_{r\phi} &= -M_{ez}, & M_{\theta\phi} &= -M_{ne}. \end{aligned} \quad (1)$$

Additional coordinate systems in use are ENU (e.g. Jost and Herrmann, 1989) and NWU (e.g. Box 8.3 in Lay and Wallace, 1995 and section 4.2.1 in Stein and Wysession, 2003). Different definitions lead to a different ordering and polarities of individual matrix components of the tensor. A wrong association may lead to mis-interpretation of components and fault directions. Therefore, the coordinate system should always be published together with the moment tensor solution.

With the exception of microseisms (ambient noise), earthquakes or explosions are the most common localized sources for seismic waves in the Earth. Other source such as tensile cracks, rock bursts or mass slope are significant sources close to the Earth's surface or in regions under high fluid overpressure (e.g. at volcanoes). All internal sources can be represented by a specific combination of generalized force couples (moment tensor components), which radiate at low frequencies the same waves as, for instance, the dislocation process on the geological fault. Having estimated a moment tensor, a decomposition and interpretation of the moment tensor is often needed to find the most appropriate geological or physical source process. In general, two concepts are followed when decomposing moment tensors:

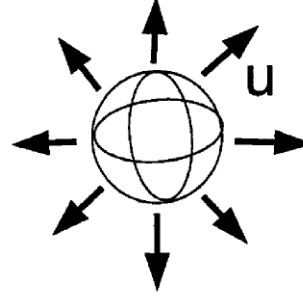
- (1) the decomposition into physical (geological) source components to aid interpretation, for instance explosions, shear and tensile cracks or
- (2) the purely mathematical decomposition as a technique to simplify the inversion or numerical analysis.

We briefly introduce the source representations of the most elementary physical sources and then discuss different decompositions.

## 2.1 Elementary physical sources

### 2.1.1 The isotropic source (explosion and implosion)

$$\mathbf{M}_{iso} = \Delta V (\eta + 2N) \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix}.$$



**Figure 2** Shape and radiation of the isotropic (volumetric) moment tensor

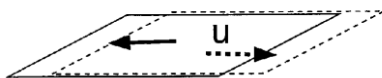
An explosion in a drill hole at depth is associated with abrupt volume change. In terms of pressure and force dipoles it is represented by an abrupt, isotropic pressure change and by means of three orthogonal linear force dipoles, respectively (see Fig. 2). The explosion (or implosion) radiates, theoretically, P waves with equal amplitude and polarity in all directions (isotropic source). S waves are not excited by the buried isotropic source. Most often, the polarity of the P wave radiation is plotted to visualize the radiation of the source. The lower hemisphere of a virtual sphere around the source is projected in a horizontal plane; examples and explanations are given in, e.g., chapter 3 of the NMSOP-2. The radiation pattern plot of an isotropic source is then a uni-colored sphere (black for explosion and white for implosion).

The only free parameter of the isotropic moment tensor,  $\mathbf{M}_{iso}$ , is the pre-factor of the unity matrix in Fig. 2, which defines the strength, or moment  $M_0 = tr = (M_{11} + M_{22} + M_{33})/3$ , of the isotropic source. Knowing  $M_0$  and the elastic constants  $N$  and  $\eta$  (see table 1) of the rocks at the source, the volume change  $\Delta V$  of the explosion or implosion source can be estimated by

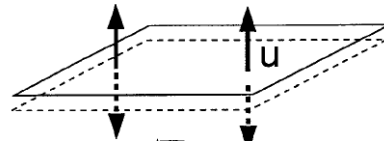
$$tr = (M_{11} + M_{22} + M_{33}) / 3 = \Delta V (\eta + 2N) \quad (\text{explosion}) \quad (2)$$

Note that dislocation sources can also generate isotropic components in the radiation pattern of the source. The volume change associated with these sources is, however, different to equation (2) (see Müller, 1973, 2001) and given below. Note further that the moment of an explosion in the atmosphere has to be treated differently (e.g., Heimann et al., 2013).

### 2.1.2 The general dislocation source



**Figure 3a:** Dislocation of a shear crack with  $\mathbf{n}=(0,0,1)$  and  $\mathbf{D}_S=(0,-1,0)$ .



**Figure 3b:** Dislocation of a tensile crack with  $\mathbf{n}=(0,0,1)$  and  $\mathbf{D}_N=(0,0,1)$ .

A source associated with a planar crack and a dislocation vector  $\mathbf{D}$  in an arbitrary direction (shear as well as opening mode, see Fig. 3) may be termed a "general dislocation source". While tectonic earthquakes rupture in pure shear mode, rupture under the presence of fluids with high overpressure may involve an additional opening of the rupture plane. This may occur during magma- or fluid-intrusions, although the validity of observations of mixed-mode rupture are often debated among experts. The moment tensor of a general dislocation source is (eq. 3.21 in Aki and Richards, 2002)

$$M_{pq} = N A (n_p D_q + n_q D_p) + \eta A n_k D_k \delta_{pq} \quad (3)$$

The general dislocation source has five independent parameters. Equation (3) can be described in terms of strike ( $\phi$ ), dip ( $\delta$ ), rake ( $\lambda$ ), and the magnitudes of the shear ( $D_S$ ) and opening ( $D_N$ ) dislocation. If the moment tensor is defined in a NED system the relation to these parameters is (e.g., Dahm, 1998)

$$\begin{aligned} M_{11}/A &= -D_S N (\sin 2\phi \sin \delta \cos \lambda + \sin^2 \phi \sin 2\delta \sin \lambda) + D_N (\eta + 2N \sin^2 \phi \sin^2 \delta) \\ M_{12}/A &= +D_S N (\cos 2\phi \sin \delta \cos \lambda + 0.5 \sin 2\phi \sin 2\delta \sin \lambda) - D_N N \sin 2\phi \sin^2 \delta \\ M_{13}/A &= -D_S N (\cos \phi \cos \delta \cos \lambda + \sin \phi \cos 2\delta \sin \lambda) + D_N N \sin \phi \sin 2\delta \\ M_{22}/A &= +D_S N (\sin 2\phi \sin \delta \cos \lambda - \cos^2 \phi \sin 2\delta \sin \lambda) + D_N (\eta + 2N \cos^2 \phi \sin^2 \delta) \\ M_{23}/A &= -D_S N (\sin \phi \cos \delta \cos \lambda - \cos \phi \cos 2\delta \sin \lambda) - D_N N \cos \phi \sin 2\delta \\ M_{33}/A &= +D_S N \sin 2\delta \sin \lambda + D_N (\eta + 2N \cos^2 \delta) \end{aligned} \quad (4)$$

### 2.1.3 The shear dislocation source

The shear dislocation source is a special case of (4) if  $D_N=0$  and is usually associated with a tectonic earthquake. It has four independent parameters. The earthquake rupture is an expression of the rapid dislocation (described by the rake angle and slip) of the two surfaces of a geological fault (described by strike and dip angles) relative to each other. The rupture plane of an earthquake may become quite large if the earthquake is strong. For instance, a magnitude  $M_W$  6 earthquake may rupture a fault of several kilometer length. The ruptured portion of the geological fault is usually approximated by a plane of area  $A$ , e.g. with length  $L$  and width  $W$ . The earthquake source is therefore idealized by the physical model of a "planar shear crack" or "shear dislocation" source. In terms of forces, a shear crack can be represented by two perpendicular force dipoles with zero angular momentum. Therefore, the shear crack source is often termed "double couple" (DC). Fig. 3a gives an example of a shear dislocation on a horizontal plane with the associated moment tensor,

$$\mathbf{M} = A D_S N \begin{bmatrix} 0 & 0 & 0 \\ 0 & 0 & -1 \\ 0 & -1 & 0 \end{bmatrix}, \quad (5)$$

where  $D_S$  is the average shear dislocation over the plane  $A$ . The seismic moment of the shear crack depends on the product between dislocation, the area of the ruptured plane and the shear modulus (see equation 3 and, e.g., equation 43 in IS3.1)

$$M_0 = \sqrt{(0.5 M_{pq} M_{pq})} = A D_S N \quad (6)$$

### 2.1.4 The tensile crack

For  $D_S=0$  in equation (4) we have a pure tensile crack. The tensile crack has only three independent parameters, since the direction of the dislocation vector is always perpendicular to the rupture plane (no rake angle defined). The principal axis moment tensor of the tensile crack of Fig. 3b, for instance, is given by

$$AD_N \begin{bmatrix} \eta & 0 & 0 \\ 0 & \eta & 0 \\ 0 & 0 & \eta + 2N \end{bmatrix} = (\eta + \frac{2}{3}N)AD_N \begin{bmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{bmatrix} + \frac{2}{3}AD_N N \begin{bmatrix} -1 & 0 & 0 \\ 0 & -1 & 0 \\ 0 & 0 & 2 \end{bmatrix} \quad (7)$$

The strength of the tensile crack may be defined in a similar manner as before. Calculating the volume change  $\Delta V$  can be of interest, for instance to estimate the volume of the intruded material. The decomposition in equation (7) shows how the volume change is related to the isotropic component of the tensile crack,  $\mathbf{M}_{\text{iso}}$ . Note this relation is different to equation (2) of the explosion source,

$$tr = (M_{11} + M_{22} + M_{33}) / 3 = \Delta V (\eta + 2N/3) \quad (\text{tensile crack}). \quad (8)$$

## 2.2 Decomposition into elementary sources

The decomposition of  $M_{pq}$  into physical sources is not unique. Several decompositions have been introduced depending on the source region and range of interpretations considered. The review paper by Jost and Herrmann (1989) gives an excellent overview and introduction to decompositions in use today. The paper describes the concepts of moment tensor decomposition, and how they can be coded by using dyadics. A brief summary is given below, and we refer to the original publication for details.

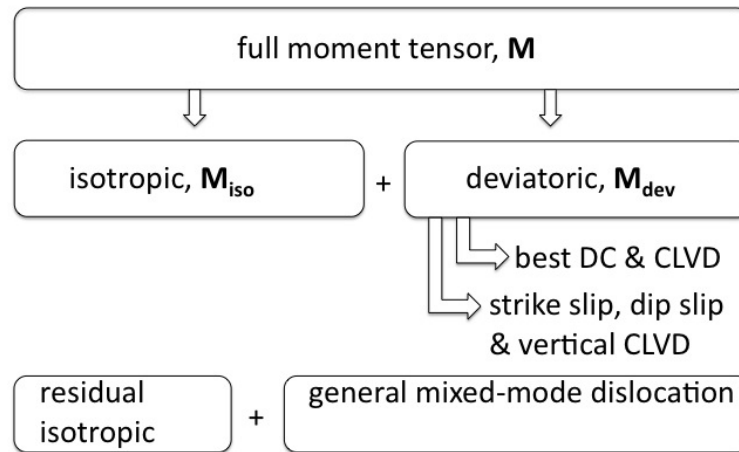
The moment tensor is first decomposed in its principal axis system

$$\mathbf{M} = \begin{pmatrix} a_{1x} & a_{2x} & a_{3x} \\ a_{1y} & a_{2y} & a_{3y} \\ a_{1z} & a_{2z} & a_{3z} \end{pmatrix} \begin{pmatrix} e_1 & 0 & 0 \\ 0 & e_2 & 0 \\ 0 & 0 & e_3 \end{pmatrix} \begin{pmatrix} a_{1x} & a_{1y} & a_{1z} \\ a_{2x} & a_{2y} & a_{2z} \\ a_{3x} & a_{3y} & a_{3z} \end{pmatrix} \quad (9)$$

Here,  $\mathbf{a}_i$  represent eigenvectors associated to eigenvalues  $e_i$ . The next step is to split the full tensor into an isotropic and a deviatoric tensor. If  $\mathbf{M}'$  represents the moment tensor in its principal axis system with eigenvalues  $e_i$  the decomposition is

$$\begin{pmatrix} e_1 & 0 & 0 \\ 0 & e_2 & 0 \\ 0 & 0 & e_3 \end{pmatrix} = \begin{pmatrix} tr & 0 & 0 \\ 0 & tr & 0 \\ 0 & 0 & tr \end{pmatrix} + \begin{pmatrix} e_1 - tr & 0 & 0 \\ 0 & e_2 - tr & 0 \\ 0 & 0 & e_3 - tr \end{pmatrix} \quad (10)$$

with  $tr = (e_1 + e_2 + e_3)/3$  being 1/3 of the trace of  $\mathbf{M}'$  and  $e'_k = e_k - tr$  being the eigenvalues of the deviatoric tensor.



**Figure 4** The two possibilities to decompose a full moment tensor, either into an isotropic and deviatoric tensor or into a general mixed-mode shear crack and a residual isotropic component. The deviatoric component can be further decomposed into different elementary sources.

The decomposition into  $\mathbf{M}_{iso}$  and  $\mathbf{M}_{dev}$  is unique. The  $\mathbf{M}_{iso}$  tensor radiates only P waves, Rayleigh waves and spheroidal normal modes with no directional preference. The deviatoric tensor ( $\mathbf{M}_{dev}$ ) often has no direct geological meaning and is therefore further decomposed into different geologically reasonable sources. Several decompositions have been suggested for the deviatoric tensor, for instance (e.g. Jost and Herrmann, 1989 and Fig. 4):

- a best double-couple (DC) and a CLVD,
- a strike slip, a pure dip slip and a vertical compensated linear vector dipole (vsCLVD),
- an isotropic source and a mixed mode double couple and tensile crack on the same crack (general dislocation source).

Geologically, the first decomposition maximizing the DC part is the most relevant one in most cases.

The first step in the decomposition is the calculation of eigenvalues and eigenvectors of the seismic moment tensor. If by definition a positive isotropic tensor is associated with volume expansion, the eigenvector of the largest eigenvalue gives the direction of the T (or tensional) axis; the one of the smallest eigenvalue the direction of the P (or compressional) axis, and the eigenvector of the intermediate eigenvalue gives the direction of the null axis. The demeaned absolute values of the eigenvalues can be used to estimate the proportional strength of the different source components of the deviatoric tensor (see Jost and Herrmann, 1989).

### 2.2.1 Decomposition into strike slip, dip slip DC sources and a vertical CLVD

Elementary tensors that fulfil orthogonality are sometimes used instead of Cartesian components (see e.g. Aki and Richards, 2002, Box 4.4, p. 112). One possible reason is to calculate Green's functions if the available codes that cannot handle single component moment tensor excitation. Another reason might be to easily incorporate constraints during inversion.

A possible decomposition into elementary tensors are: (1) a strike-slip double-couple ( $\mathbf{M}_1$ ), (2) a 90° dip-slip double-couple ( $\mathbf{M}_2$ ), (3) an isotropic tensor and (4) a vertical CLVD ( $\mathbf{M}_{v-clvd}$ ),

$$\mathbf{M} = \mathbf{M}_1(\phi_S) + \mathbf{M}_2(\phi_D) + \mathbf{M}_{iso} + \mathbf{M}_{v-clvd} \quad \text{with (in Cartesian system)}$$

$$\begin{aligned} \begin{pmatrix} M_{11} & M_{12} & M_{13} \\ M_{21} & M_{22} & M_{23} \\ M_{31} & M_{32} & M_{33} \end{pmatrix} &= \begin{pmatrix} -\frac{1}{2}(M_{22} - M_{11}) & M_{12} & 0 \\ M_{21} & \frac{1}{2}(M_{22} - M_{11}) & 0 \\ 0 & 0 & 0 \end{pmatrix} \\ &+ \begin{pmatrix} 0 & 0 & M_{13} \\ 0 & 0 & M_{23} \\ M_{31} & M_{32} & 0 \end{pmatrix} \\ &+ \frac{1}{3}(M_{11} + M_{22} + M_{33}) \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix} \\ &+ \frac{1}{3} \left( \frac{1}{2}(M_{22} + M_{11}) - M_{33} \right) \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & -2 \end{pmatrix}, \end{aligned} \quad (11)$$

The first and second term represent a strike-slip and pure dip-slip source, respectively, with strike angles  $\phi_S$  and  $\phi_D$ . The third term in equation (22) is the isotropic component and the forth term the vertical CLVD source. The strike slip source,  $\mathbf{M}_1$ , has a  $\pi$ -periodic azimuthal radiation pattern, the dip-slip  $\mathbf{M}_2$  a  $2\pi$ -periodic, and  $\mathbf{M}_{v-clvd}$  and  $\mathbf{M}_{iso}$  an azimuth-independent radiation pattern. SH- waves cannot resolve  $\mathbf{M}_{v-clvd}$  and  $\mathbf{M}_{iso}$ , SV-waves are not sensitive to  $\mathbf{M}_{iso}$ , while P-waves theoretically can resolve all elementary sources. The decomposition in equation (22) can be expressed in terms of strike, dip, and rake of the best double couple, the null-eigenvalue, and the trace of the tensor (Dahm, 1993).

### 2.2.2 Decomposition into a best double couple and a CLVD:

The decomposition, including the isotropic component of the full tensor, is commonly expressed by

$$\begin{bmatrix} e_1 & 0 & 0 \\ 0 & e_2 & 0 \\ 0 & 0 & e_3 \end{bmatrix} = tr \begin{bmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{bmatrix} + (e_2 - tr) \begin{bmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & -2 \end{bmatrix} + (e_1 - e_2) \begin{bmatrix} 1 & 0 & 0 \\ 0 & 0 & 0 \\ 0 & 0 & -1 \end{bmatrix} \quad (12)$$

If  $e_{min}$  and  $e_{max}$  are the smallest and largest absolute values of  $e'_k$  (assume  $|e'_3| \geq |e'_1| \geq |e'_2|$ ) and with  $F = e_{min} / e_{max} = e'_2 / e'_3$ , one may define  $100(1-2F)$  as the percentage of the double couple component (between 0% and 100%, see e.g. Dziewonski et al., 1981).

The DC component (last term in 12) has been described above. A CLVD source (middle term in 12) has no verified geological representation and is for crustal earthquakes often interpreted as source component representing the residual radiation from the best double couple source. The residual radiation may result from noisy data or from simplifying assumptions, as for instance from neglecting wave effects of a 3D Earth during the inversion. Therefore, a small or zero CLVD component is sometimes interpreted as a confirmation of the model assumption.

The decomposition of the deviatoric component into best DC and CLVD is commonly used in seismology. For instance, the source mechanism reported by nearly all seismic services are based on the decomposition into a best double couple and a CLVD component. A computer program for such a decomposition is described below.

### 2.2.3 Decomposition into an isotropic source and a mixed-mode shear-tensile crack:

The decomposition of the principal axis tensor into  $M_{iso}$  and a general dislocation crack is

$$\begin{bmatrix} e_1 & 0 & 0 \\ 0 & e_2 & 0 \\ 0 & 0 & e_3 \end{bmatrix} = tr \begin{bmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{bmatrix} + (e_2 - tr) \begin{bmatrix} \frac{3}{2} \cos 2\phi - \frac{1}{2} & 0 & \frac{3}{2} \sin 2\phi \\ 0 & 1 & 0 \\ \frac{3}{2} \sin 2\phi & 0 & -\frac{3}{2} \cos 2\phi - \frac{1}{2} \end{bmatrix} \\ - \frac{3}{2}(e_2 - tr) \tan 2\phi \begin{bmatrix} \sin 2\phi & 0 & \cos 2\phi \\ 0 & 0 & 0 \\ \cos 2\phi & 0 & -\sin 2\phi \end{bmatrix}, \quad (13)$$

with  $\tan^2 \phi = \frac{-3(e_2 - tr) + e_3 - e_1}{3(e_2 - tr) + e_3 - e_1}, e_2 - tr \neq 0$ .

where the angle  $\phi$  defines the angle of the dislocation plane measured against the T axis in the plane given by the T and P axis (P and T are interchanged in case of contraction) (e.g. Dahm, 1993; Dahm and Brandsdottir, 1997). Another formulation of mixed-mode crack decomposition is given by Vavryčuk (2001).

## 2.3 Computer programs for decomposition

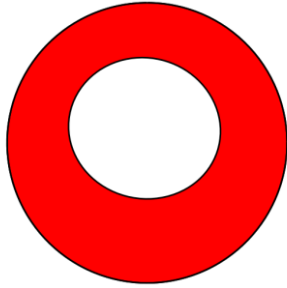
Computer programs for decomposition were written by Jost and distributed in Volume VIII of the software package “Computer Programs in Seismology” (see R. Herrmann, Saint Louis University <http://www.eas.slu.edu/People/RBHerrmann/ComputerPrograms.html>). Another computer package is provided with the python tool ”MoPaD” (Krieger and Heimann, 2012; <http://www.larskrieger.de/mopad/>), which can be used for decomposition, projection and the plotting of moment tensors. MoPaD can be easily combined with GMT plotting tools (<http://gmt.soest.hawaii.edu>).

The following examples are derived using MOPAD. On 19 March 2013 a  $M_L$  4.2 earthquake struck the Polish mine district close to the Rudna copper mine. 19 mine workers were trapped in the mine at 850 m depth for several hours due to rock bursts affecting parts of the underground tunnels. A centroid moment tensor inversion (0.07-0.1 Hz) using broadband regional stations retrieved a seismic moment of  $M_W$  3.8 ( $M_0=5.3E14$  Nm) at a source depth of

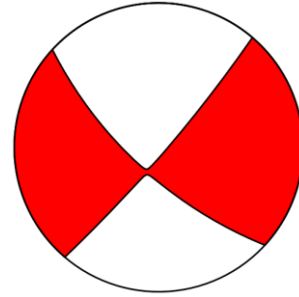
about 1 km (Whidden et al., 2013; Rudzinski and Lizurek, pers. communication). The NED Cartesian moment tensor is decomposed into an isotropic and deviatoric part,

$$\begin{aligned}\hat{\mathbf{M}} &= 5.3 \cdot 10^{14} Nm \begin{pmatrix} -0.40 & +0.01 & -0.10 \\ +0.01 & -0.43 & +0.01 \\ -0.10 & +0.01 & -1.00 \end{pmatrix} \\ &= 5.3 \cdot 10^{14} Nm \begin{pmatrix} -0.61 & & \\ & -0.61 & \\ & & -0.61 \end{pmatrix} + 5.3 \cdot 10^{14} Nm \begin{pmatrix} +0.21 & +0.01 & -0.10 \\ +0.01 & +0.18 & +0.01 \\ -0.10 & +0.01 & -0.39 \end{pmatrix}\end{aligned}$$

The radiation pattern is plotted in a lower hemispherical projection (Fig. 5a, and chapter 3 of NMSOP-2 for explanations on the type of plots). The isotropic component is negative (implosion) and covers 60% of the moment of the full tensor.



**Figure 5a** Lower hemispherical projection of the P-wave radiation (stereographic) of the Rudna mine 2013  $M_W$  3.8 moment tensor solution



**Figure 5b** Lower hemispherical projection (stereographic) of the 13 Oct. 2013  $M_W$  6.6 Gulf of California moment tensor solution.

Apparently, the relatively slow collapse of void space in the mine building dominates the long period wave radiation and gives the major contribution of the seismic moment. The 40% deviatoric source component can be decomposed in a best double couple (DC) and a CLVD component, where the DC covers only about 13% of the size of the deviatoric tensor. The low frequency radiation pattern can therefore be interpreted in terms of an isotropic and CLVD component, where the CLVD is roughly vertically oriented. The source process may be interpreted in terms of a tensile crack as given in equation (23), by e.g.

$$AD_N \begin{pmatrix} \eta & 0 & 0 \\ 0 & \eta & 0 \\ 0 & 0 & \eta + 2N \end{pmatrix} \approx 5.3 \cdot 10^{14} Nm \begin{pmatrix} -0.4 & 0 & 0 \\ 0 & -0.4 & 0 \\ 0 & 0 & -1.00 \end{pmatrix}$$

and thus



$$\begin{aligned}
AD_N &= 5.3 \cdot 10^{14} \text{ Nm} \cdot 0.5(-1.0 + 0.4)/25 \cdot 10^9 \text{ N/m}^2 \approx 6360 \text{ m}^3 \\
\eta &= -0.4 \times 5.3 \cdot 10^{14} \text{ Nm} / -6360 \text{ m}^3 \approx 33 \text{ GPa} \\
\nu &= \frac{\eta}{2(\eta + N)} \approx 0.28
\end{aligned}$$

The collapsed volume  $AD_N$  can be used to estimate the area affected by the collapse if the average height of the galleries in the mine is specified. For instance, assuming the average height of the gallery is 5m, the estimated area is about 35m x 35m. The estimated Poisson ratio of 0.28 and the derived Lamé's parameter may be used to check the consistency of the source interpretation with expected rock properties. Ford et al. (2008) studied a mine collapse event of a coal mine in Utah and estimated a vertical crack component of comparable relative strength.

The second example concerns the moment tensor solution of an  $M_w$  6.6 earthquake in the Gulf of California that occurred on 13 October 2013 at a depth of 16 km. The global moment tensor solution in an USE system is (GFZ solution) is

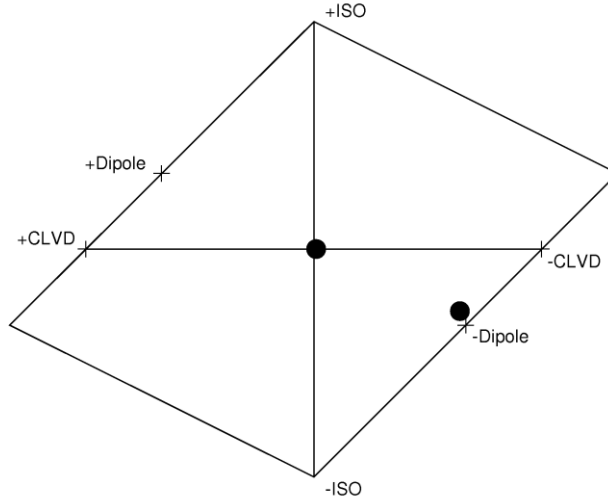
$$\hat{\mathbf{M}} = 1 \cdot 10^{18} \text{ Nm} \begin{pmatrix} -0.59 & -2.52 & -1.72 \\ -2.52 & -7.58 & -0.95 \\ -1.72 & -0.95 & +8.16 \end{pmatrix}$$

The isotropic component is constraint to zero during inversion. The seismic moment is estimated to as 9.1E18 Nm. The DC component of the solution is 99%, meaning that the rupture can be interpreted as pure shear crack, in this case a dominant left lateral strike slip mechanism (see Fig. 5b) with strike, dip and rake of the two nodal planes of

$$\begin{aligned}
\text{plane 1: } & \Phi = 133^\circ, \quad \delta = 69^\circ, \quad \lambda = -173^\circ \\
\text{plane 2: } & \Phi = 40^\circ, \quad \delta = 84^\circ, \quad \lambda = -20^\circ
\end{aligned}$$

## 2.4 Displaying different moment tensor components

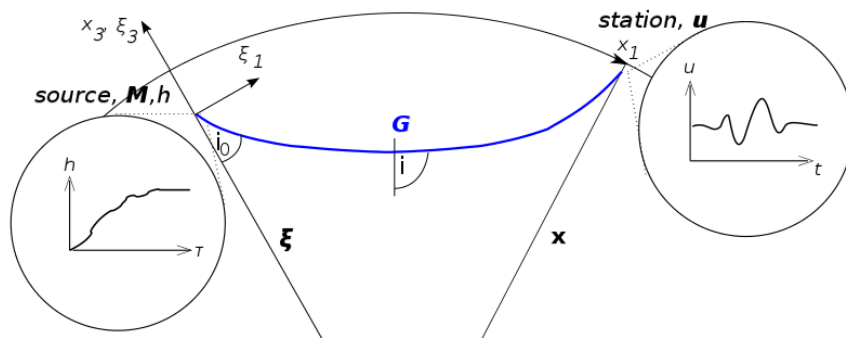
Displaying the elementary source components in a simple graph is not trivial. Hudson et al. (1989) and Riedesel and Jordan (1998) suggested diagrams for this purpose. The Hudson et al. (1989) plot is now often used. See Tape and Tape (2012) for a geometric comparison of source-type plots. Figure 6 displays the Hudson-type plot for the two moment tensors discussed in 2.3. The pure double couple moment tensor is centered at the origin and indicates a pure DC moment tensor. The Rudna mine solution is localized in the lower half-space indicating negative isotropic source components. It is very close to a pure linear dipole solution, which further shows that the source is close to a pure collapsing crack solution. Opening cracks would appear in the positive half-space close to the positive dipole marker. Pure explosions and isotropic implosions events appear on the vertical axis of the plot frame at the points marked by +ISO and -ISO, respectively.



**Figure 6** Hudson plot for displaying source components. The filled circles represent the source components of the two examples in 2.3.

### 3 Moment tensor representations: basic relations

A moment tensor representation is a formula to explain the theoretical relation between ground motions at the stations and the moment tensor at the source. This usually considers the dynamic ground motion and the wave radiation pattern in the far field. However, the theory also explains static and near field (near source) ground motions as well as source-generated strain and stress variations in the Earth. Knowing the moment tensor representation (the forward problem),  $\mathbf{M}$  can be obtained by inverting one or several of the observed field variables. Moment tensor inversion requires the availability digital time series of observed data, e.g. full seismograms or at least P and S phases, measured un-clipped on linear and predictable acquisition systems (e.g. calibrated systems), and the calculation or availability of accurate synthetic seismograms of the Earth (i.e. Green's functions, denoted by  $\mathbf{G}$ ). Therefore, moment tensor inversion is more demanding from an observational and computational point of view.



**Figure 7** Geometry of the problem. The Green's tensor  $\mathbf{G}$  describes the impulse response of the Earth at station  $\mathbf{x}_0$  for a force excitation at  $\xi_0$ . The point source is described by moment tensor  $\mathbf{M}$  and source time function  $h(\tau)$ . The ground motion at the station,  $u(t)$ , differs from the source excitation because of the filter effect of the Earth. The blue line indicates the ray path. The ray angle  $i$  is measured against the vertical; at the source  $i_0$  defines the take-off angle. Note that the global coordinate system is source centered and that the 1-direction of both systems is defined by the source station path.

The basic theory and representation of seismic sources is outlined in chapter 3 of the NMSOP-2 and in IS3.1. We build on this theory and concentrate on the point source moment tensor representation of ground displacement (see equation 21 in IS3.1, neglecting the single force term)

$$u_n(\mathbf{x}, t) = M_{pq}(\boldsymbol{\xi}, t) * \frac{\partial}{\partial \xi_q} G_{np}(\mathbf{x}, \boldsymbol{\xi}, t)|_{\xi_0} = M_{pq}(\boldsymbol{\xi}, t) * G_{np,q}(\mathbf{x}, \boldsymbol{\xi}, t) \quad (14)$$

Equation (14) may alternatively be given in the frequency domain. The components of the ground displacement at location  $\mathbf{x}$  are declared by  $u_n$ .  $\mathbf{M}$  is the moment tensor of the seismic point source at location  $\boldsymbol{\xi}$ , and  $G_{np,q}$  are spatial derivatives with respect to  $\xi_k$  of components of the Green's tensor  $\mathbf{G}$ . The Green's tensor can be viewed as a structural term defined between  $\boldsymbol{\xi}$  and  $\mathbf{x}$  which describes all wave propagation effects including the elastostatic response of the Earth due to a single force delta-excitation at point  $\boldsymbol{\xi}$  measured at point  $\mathbf{x}$  (Fig. 7). The first index of  $G_{np}$  gives the direction of the ground motion at the station, and the second index gives the direction of the force at the source. It is common practice to denote the two independent coordinate systems of the receiver  $(\mathbf{x}, t)$  and the source point  $(\boldsymbol{\xi}, \tau)$  as independent variables in  $\mathbf{G}$ . Note that the following equivalence is used  $G(\mathbf{x}, t; \boldsymbol{\xi}, \tau) = G(\mathbf{x}, t - \tau; \boldsymbol{\xi}, 0)$  (e.g., Aki and Richards, 2002). The components of the Green's tensors are often named Green's functions and may represent full seismograms. Later, we will introduce elementary seismograms as a special component of derivatives of the Green's tensor.

Equation (14) can be used to invert for moment tensors. Necessary condition is that observed seismograms have been deconvolved to ground displacement, velocity or acceleration (see IS11.1) and that Green's functions have been calculated between the source and receiver point assuming an appropriate Earth model. We will give examples of Green's function calculation below. Equation (14) can be written in matrix form, e.g. by considering the discrete form of the multichannel convolution integral (e.g., Krieger, 2011). We introduce notations  $\mathbf{d}$ ,  $\mathbf{G}'$  and  $\mathbf{y}$  to indicate that the data and model vectors are not directly the tensor components defined in (14). They represent channels from more than one station ( $R$  is the maximal channel number), either in displacement, velocity or acceleration, and the derivatives or Green's functions. In the time domain,  $\mathbf{d}$ ,  $\mathbf{G}'$  and  $\mathbf{y}$  further contain individual time samples (denoted by subscript index of time variables) at the different components and at different stations. We find

$$\begin{bmatrix} d_1(t_0) \\ d_2(t_0) \\ d_3(t_0) \\ \vdots \\ d_R(t_0) \\ d_1(t_1) \\ d_2(t_1) \\ \vdots \\ \vdots \end{bmatrix} = \begin{bmatrix} G'_{11}(t_0, \tau_0) & G'_{21}(t_0, \tau_0) & \dots & G'_{11}(t_0, \tau_1) & G'_{21}(t_0, \tau_1) & \dots \\ G'_{12}(t_0, \tau_0) & G'_{22}(t_0, \tau_0) & \dots & G'_{12}(t_0, \tau_1) & G'_{22}(t_0, \tau_1) & \dots \\ G'_{13}(t_0, \tau_0) & G'_{23}(t_0, \tau_0) & \dots & G'_{13}(t_0, \tau_1) & G'_{23}(t_0, \tau_1) & \dots \\ \vdots & \vdots & \ddots & \vdots & \vdots & \ddots \\ G'_{1R}(t_0, \tau_0) & G'_{2R}(t_0, \tau_0) & \dots & G'_{1R}(t_0, \tau_1) & G'_{2R}(t_0, \tau_1) & \dots \\ G'_{11}(t_1, \tau_0) & G'_{21}(t_1, \tau_0) & \dots & G'_{11}(t_1, \tau_1) & G'_{21}(t_1, \tau_1) & \dots \\ G'_{12}(t_1, \tau_0) & G'_{22}(t_1, \tau_0) & \dots & G'_{12}(t_1, \tau_1) & G'_{22}(t_1, \tau_1) & \dots \\ \vdots & \vdots & \ddots & \vdots & \vdots & \ddots \\ \vdots & \vdots & \ddots & \vdots & \vdots & \ddots \end{bmatrix} \cdot \begin{bmatrix} y_1(\tau_0) \\ y_2(\tau_0) \\ \vdots \\ y_6(\tau_0) \\ y_1(\tau_1) \\ y_2(\tau_1) \\ \vdots \\ \vdots \end{bmatrix}$$

or equivalently

$$\mathbf{d} = \mathbf{G}' \mathbf{y} . \quad (15)$$

Equation (15) is given in the form of a linear system of equations, and represents a discrete form of the multichannel convolution integral.  $\mathbf{d}$  and  $\mathbf{y}$  are the ‘data’ and ‘model’ vector, respectively, and  $\mathbf{G}'$  represents the coefficient matrix. Dimension of  $\mathbf{d}$ ,  $\mathbf{y}$  and  $\mathbf{G}'$  are determined by the length of the data seismograms, the number of channels and the length of the assumed source time function ( $N_\tau$ ), and can easily grow large. The data vector  $\mathbf{d}$  consists of individual seismograms measured at different stations. These may be full seismograms, usually band-pass filtered, or windowed traces of specific wave phases. The coefficient matrix  $\mathbf{G}'$  is build from derivatives of the Green’s functions (filtered and processed in the exact same way as the data). The components of the model vector  $\mathbf{y}$  contain the  $6 \times N_\tau$  unknowns and are formed by holomorphic mapping of the moment tensor components into a model vector  $\mathbf{y}$ , e.g.

$$\begin{aligned} y_1 &= M_{11}, & y_2 &= M_{12}, & y_3 &= M_{22}, \\ y_4 &= M_{13}, & y_5 &= M_{23}, & y_6 &= M_{33}, \end{aligned} \quad (16)$$

Equation (15) is typically solved by the least squares (LS) inversion technique, either in the time domain (e.g. multichannel Wiener filtering described, e.g., in Wiggins and Robinson, 1965; Sipkin, 1982) or in the frequency domain (e.g. generalized inverse using singular value decomposition, e.g., Menke 1989).

The system (15) is over-determined, i.e. it has more independent equations (data) than unknowns, if for instance enough ground motion measurements are available from different distances and station azimuths, such that the azimuthal data coverage is sufficient to resolve the radiation pattern of the source. Also, the assumed source time duration should be much smaller than the length of the seismograms (or is parameterized by simple functions). Equation (15) is written for 27 independent components of the spatial derivatives of the Green’s tensor (3<sup>rd</sup> order tensor, i.e.  $3 \times 3 \times 3$  components) if three component seismograms are inverted. Green’s functions in this general formulation would consider near and far field terms. The time history of each moment tensor component is assumed to be independent. Altogether, this leads to a large number of unknowns, which are possibly difficult to resolve. Additionally, the choice of waves types selected (near field, far field, body wave phases, surface waves, etc.), the frequency range, the type of applied filters as well as the weighting scheme applied to the equations may influence the outcome of the inversion. Therefore, the equations implemented in practice are often modified in comparison to equation (15). For instance, the number of unknowns is reduced, e.g. by simplifying the time history of the moment tensor or by specifying the type of source mechanism (e.g. non-explosive). The number of Green’s functions is reduced by taking advantage of Earth model symmetries and/or by neglecting near field terms. Additionally, damping and linear or non-linear constraints may be considered. This leads to modified schemes and equations.

We discuss representations of the source time and moment rate function, the concept of joint inversion point source moment tensors and centroid locations, the simplified representation formulas for a layered Earth model and for the far field, formulas for amplitude spectra and specific representations if only body waves or peak amplitudes are used. Some words are given for the relative moment tensor inversion, the normal mode summation and how to calculate Green’s functions in practice.

### 3.1 Resolving the time dependency of moment tensors

The components of the moment tensor are time dependent and are related to the time history of slip (or rupture) at the source. Equation (15), or the equivalent form in the frequency domain, invert for the independent time function of each moment tensor component. The problem is often ill-posed and source-time functions vary a lot between the components. Different approaches have been suggested to stabilize the source time function inversion.

Often, a homogeneous rupture process is assumed, so that each moment tensor component has the same time history, e.g.  $h(t)$ . Then, it is useful to factorize the time history and leave only the magnitude of the time-independent scaling factors of the moment tensor components, e.g. denoted by  $\hat{M}_{pq}$ , leading to

$$M_{pq}(t) = h(t)\hat{M}_{pq} \quad \text{with} \quad h(t) = \begin{cases} 0 & \text{for } t \leq 0 \\ 1 & \text{for } t > T_d > 0 \end{cases} \quad (17)$$

where  $T_d$  is the duration of the ramp-like source time.

A common simplification is to use only far-field recordings. The far field displacement scales with the time derivative of  $h$ , i.e.  $\dot{h} = dh/dt$ , often denoted as moment rate function.

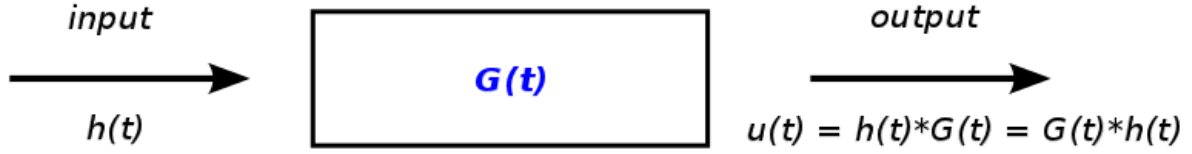
Near field Green's functions represent the component of the Green's functions, which remain zero at all times far from the source. Near and far field terms can be derived by equating the derivative of  $\mathbf{G}$  with respect to the source coordinate  $\xi_k$ , and separating the terms attenuating faster and slower. Lets assume  $\xi$  as independent variable. Both other intrinsic variables of  $G$ ,  $x$  and  $t-\tau$ , depend on the distance to the source, which itself depends on  $\xi$ . The derivative of  $G$  is equated by

$$\frac{\partial}{\partial \xi_q} G_{np}(\mathbf{x}, t') = \frac{\partial G_{np}}{\partial x_l} \frac{\partial x_l}{\partial \xi_q} + \frac{\partial G_{np}}{\partial t} \frac{\partial t'}{\partial \xi_q} = G_{npq}^{(n)} + \dot{G}_{npq}^{(f)} \quad (18)$$

$t'$  is the retarded time at the station. For body waves,  $\partial t'/\partial \xi_k$  are components of the slowness vector of the phase. The first and second term are associated with the near-field and far-field term of  $\mathbf{G}$ , associated by superscripts (n) and (f). Inserting (17) and (18) into (14) leads to the far field moment tensor representation as

$$u_n^{(f)}(\mathbf{x}, t) \approx \hat{M}_{pq} h(t) * \dot{G}_{npq}^{(f)}(\mathbf{x}, t) = \hat{M}_{jk} \dot{h}(t) * G_{npq}^{(f)}(\mathbf{x}, t) \quad (19)$$

Equation (19) shows that displacement in the far field depends on moment rate function and the far field Green's functions excited by a unit step at the point source. This is a general result, which explains that the far field body wave pulses from earthquakes are related to time derivatives of the source time function at the source. If the source time function is an ideal step function, the far field pulse is a pulse like. If it is a ramp function of duration  $T_d$ , the expected body wave pulse has the shape of a boxcar function of duration  $T_d$ .



**Figure 8** The Earth is described as a linear system with impulse response  $G(t)$  (Green function). The ground motion response  $u(t)$  at the station depends on  $G(t)$  and the excitation  $h(t)$  at the seismic source (input). The shape of the source radiation, i.e. the moment tensor, has been set unity in the 1D sketch (i.e. representing an explosive source as special case).

The interrelation between source time function  $h(t)$  and displacement  $u(t)$  can also be understood as a filtering effect (Fig. 8).  $h(t)$  is the input signal. The linear filter is represented by  $G(t)$  as the elastic response of the Earth. The ground displacement at the station,  $u(t)$ , is the filter output, and is equated as a convolution integral between the input and the filter.

Before we give some examples on how to estimate moment rate functions in practice, we use equation (19) to derive and explain the centroid time  $\tau_0$ . The centroid time is a point source parameter, defining the centre time of the wave radiation process. It is defined from the long period approximation of (19), which is valid if the period is  $T > T_d$ . A Taylors series expansion of the Green's tensor in (19) around the centroid  $\tau_0$  leads to

$$u_n^{(f)}(\mathbf{x}, t) \approx \hat{M}_{pq} G_{npq}^{(f)}(\mathbf{x}, t - \tau_0) + \hat{M}_{pq} \dot{G}_{npq}^{(f)}(\mathbf{x}, t - \tau_0) \int_0^\infty \dot{h}(\tau)(\tau - \tau_0) d\tau + \dots$$

The convolution integral is written in explicit form. The centroid time  $\tau_0$  is then defined in such a way that the integral on the right-hand side vanishes (see Fig. 9 and Nabelek, 1984). For low-pass filtered seismograms the other higher order terms can be omitted, which gives the temporal (and spatial) point source representation of ground displacement as

$$u_n^{(f)}(\mathbf{x}, t) = \hat{M}_{pq} G_{npq}^{(f)}(\mathbf{x}, t - \tau_0) \quad . \quad (20)$$

Note that the far field displacements in (20),  $u^{(f)}$ , becomes zero a long time after waves have passed through the observing point since Green's functions in the far field are zero for large  $t$ .

Although equation (20) is appropriate if the seismograms are low-pass filtered, many authors try to resolve the time history  $h(t)$  or, at least, the source duration. Although we have not yet strictly defined the spatial point source, we note that attempts to retrieve the source time functions of a spatial point source are not always useful. This is because a spatial point source moment tensor representation requires wavelengths larger than the size of the rupture plane. Resolving the source time is then only possible if the rise-time of the slip function at a single slip patch is longer than the rupture duration.

A common method to resolve the time dependency of  $h(t)$  uses the linear representation in (15) and a two step inversion approach. Examples are given by Vasco (1989) or Cesca and Dahm (2007). The first inversion step estimates the independent time function of each

moment tensor component  $y_k$ . This may be performed either in the time or in the spectral domain. The second inversion step aims to retrieve the dominant common time history  $h(t)$  presented in each component. A singular value decomposition of the matrix formed by the time dependent moment tensors is calculated. The matrix of orthogonal eigenvectors are weighted by the eigenvalues. The dominant time history common to all source components is then represented by the system of equations, which considers only the largest eigenvalue and sets all other eigenvalues to zero. More details can be found in e.g. Vasco (1989), Cesca and Dahm (2007) or Vavryczuk and Kühn (2012).

Other methods use smooth parameterizations of the moment rate function and solve directly the non-linear inverse problem for  $dh/dt$  (equation 19). The two most common parameterizations are a series of boxcars or overlapping triangles (e.g., Tsai et al., 2005, for the Great Sumatra-Andaman earthquake). The boxcar parameterization and the iterative, linearized least-squares inversion is described in, e.g., Tanioka and Ruff (1997) or Lay and Wallace (1995).

Tocheport et al. (2007) follow a different approach, which uses a stack of time-shifted and weighted body wave pulses at different stations minimized against an average source time function. The time shifts and weights are parameter of the inversion, which is solved by simulated annealing. Since the method requires isolated body waves, the method can only be applied to deep and intermediate earthquakes.

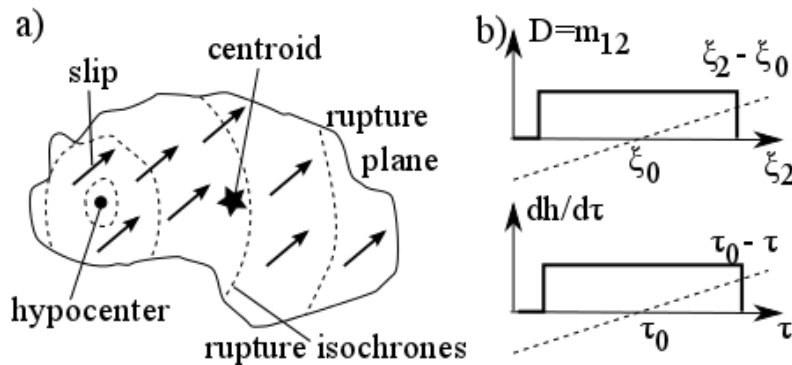
If a point source inversion is performed and seismograms are lowpass filtered, the details of the source time function of most earthquakes cannot be resolved. In such a case, often, only a rise time of the source time function is estimated (equivalent to the duration of a boxcar moment rate function, or other simple shapes). For instance, different rise times are tested in a grid search approach and the one leading to the least residuals in the waveform inversion is finally chosen (e.g., Heimann, 2011). The grid search approach can be used to determine station-related, apparent source durations after a best point source moment tensor is estimated. This may be of interest for a quick and rough analysis of rupture directivity, i.e. to separate fault and auxiliary plane and to distinguish the direction of rupture. For instance, Cesca et al. (2010b) suggest to keep the seismic moment and source mechanism of the point source solution and to invert for the source duration at each station. The apparent source time functions can then be interpreted in terms of a simplified kinematic rupture model. Zhang et al (2013) use an iterative deconvolution and stacking method to retrieve apparent source time functions for the purpose of rapid source imaging.

### 3.2 Representation of the centroid location (centroid moment tensor inversion)

Equation (20) was already defined for a centroid time  $\tau_0$  as origin time to calculate Green's functions. Similarly to the derivation of  $\tau_0$  spatial centroid coordinates  $\xi_0$  can be derived by means of a Taylors series expansion of Green's functions. The problem is sketched in Fig. 9, a detailed derivaton can be found in, e.g., Nabelek (1984). Often, the hypocenter from the location of first phase arrivals is taken as a first approximation of the centroid. However, the hypocenter, which represents the point of rupture initiation (nucleation), may be tens or hundred kilometer away from the slip centroid if the earthquake was very large. Thus, the Green's functions calculated for the parameters of the hypocenter may be wrong or at least have large phase shifts relative to observed data. This may bias the moment tensor results. Three approaches are in use to reduce the possible bias.

The first workaround, instead of estimating a best centroid position, is to introduce ad hoc static phase shifts to the observed waveforms before inversion, for instance by cross correlation between observed and synthetic data. To first order the phase shifts may correct the mislocation of the centroid and leave the amplitudes of seismograms unchanged. However, the estimation of phase shifts is a tricky task and may easily lead to undesired (incorrect) shifts, especially if automatic processing is performed.

The more rigorous approach is to estimate the centroid during the inversion, which is named a centroid moment tensor inversion. The inverse problem has four additional unknowns (moment tensor components plus spatial and temporal coordinates) and becomes nonlinear. The centroid inversion can be setup similar to the Geiger method of a standard location (IS11.1). The nonlinear equations are approximated by the first linear terms of a Taylor series expansion around a starting solution. For instance, the hypocenter is used as starting location. The inversion for changes of source parameters is linear and solved employing a standard LS method. The solution is used to improve the starting solution, and the process is iteratively repeated until the changes become very small or the residuals are minimal. Since the equations for moment tensor inversion are linear, the moment tensor inversion itself is straight forward will always find an optimal solution. However, Green's functions have to be re-calculated for the updated centroid location during each iteration. The centroid location problem is nonlinear and will have the same problems of non-uniqueness and non-convergence as standard earthquake location, especially since waveforms instead of first motions are fitted. In practice, depth centroid moment tensor inversion is thus often estimated by a grid search over a depth range in fixed steps combined with additional empirical static phase shifts at each station (limited to a fixed range of allowed phase shifts). Considering static phase shifts in addition to centroid depth location has the advantage that the impacts from incorrect Earth models or unaccounted 3D structure are reduced.



**Figure 9** a) Schematic sketch of an earthquake rupture and the definition of hypocenter, centroid, rupture plane and slip. Rupture and slip is parameterized by the moment tensor density,  $\mathbf{m}$ , and its time history  $h(t)$ . b) 1D visualization of spatial and temporal centroid coordinates to balance the third order spatial ( $\int_V (\xi_k - \xi_{0k}) m_{ij} dV = 0$ ) and temporal moments ( $\int_\tau (\tau - \tau_0) (dh/d\tau) d\tau = 0$ ) to zero (principle is comparable to finding the center of gravity).

The third approach to handle an unknown centroid location is to invert amplitude spectra in the first iteration. Amplitude spectra discard the phase information and are therefore less affected by incorrect centroid locations and Earth models. However, the amplitude spectra



cannot resolve the polarity of the source mechanism. This means an amplitude spectra inversion alone cannot resolve whether the seismic source was an explosion or an implosion, or whether the earthquake was a thrust or a normal faulting event. However, strike and dip of the fault and auxiliary planes are resolved. In order to resolve the polarity ambiguity either additional information is added to the inversion (e.g. first motion polarities), or the inversion is performed in a two-step approach, where amplitude spectra are inverted in the first iteration, and time-shifted full waveforms are considered in addition in a 2<sup>nd</sup> iteration. The 2<sup>nd</sup> step may keep strike and dip of the first solution fixed. We will further discuss this third approach below.

### 3.3 Representation in a layered Earth model

If a layered Earth model is assumed, only 10 instead of 27 Green's functions are needed. If additionally only far field seismograms (see IS3.1) are analyzed, the number of required Green's functions is further reduced to 8. If only body wave phases are inverted (ray-theoretical description), the number of Green's functions further reduces to 3 greatly simplifying the inverse problem. The reduced set of specific Green's functions are often combinations of the original Green's functions defined under specific azimuth angles and are called elementary seismograms. We briefly outline the redundancy of Green's tensors and the definition of elementary seismograms.

Fig. 7 indicates two coordinate systems, a local one defined at the source location and a global one for the source-station geometry. Sensor components of seismometers are usually measured in an east–north–up system. The sensor components in the representation formulas and in Fig. 7 should be defined in the curvilinear, global source-centered system, e.g. by radial, transversal and down directions (distance  $r$ , azimuth  $\phi$  and depth  $z$ ). However, the moment tensor is usually defined for a local Cartesian system. At the pole, the local NED system, with  $x=N$ ,  $y=E$  and  $z=down$ , has the same directions as the local spherical system. Only in the specific case where the great circle path between source and receiver is along a meridian, i.e. the station is under azimuth  $\phi'=0$ , the local and global directions would be identical. We denote the Green's tensor in this specific case by  $\mathbf{G}^0$ . If the media is invariant to azimuthal rotations around  $z$ , e.g. layered Earth models, the Green's tensor can be given as a rotated version of  $\mathbf{G}^0$ . We define unprimed and primed indices to define the unrotated (NED) and rotated azimuth directions.  $R(\phi')$  declares a matrix describing rotation with angle  $\phi'$  around the  $z$  direction. Using tensor rotation the Green's tensor in the source-station system can be equated from  $\mathbf{G}^0$

$$G_{n'p',q'} = R_{p'p}(\phi') R_{q'q}(\phi') G_{n'p,q}^0 = R_{p'p}(\phi') G_{n'p,q}^0 R_{qq'}(-\phi') \quad (21)$$

At  $\phi'=0$  the P-SV motion can only be excited by moment tensor components  $M_{xx}$ ,  $M_{zz}$ ,  $M_{xz}$ ,  $M_{zx}$  and  $M_{yy}$  (note that  $M_{yy}$  comes into play only for near field displacements). Contrary, SH motion can only be excited by moment tensor components  $M_{xy}$  and  $M_{yx}$ . This reduces the zero azimuth tensor to

$$\mathbf{G}_r^0 = \begin{pmatrix} G_{rx,x}^0 & 0 & G_{rx,z}^0 \\ 0 & G_{ry,y}^0 & 0 \\ G_{rz,x}^0 & 0 & G_{rz,z}^0 \end{pmatrix}$$

$$\mathbf{G}_\varphi^0 = \begin{pmatrix} 0 & G_{\varphi x,y}^0 & 0 \\ G_{\varphi y,x}^0 & 0 & G_{\varphi y,z}^0 \\ 0 & G_{\varphi z,y}^0 & 0 \end{pmatrix}$$

$$\mathbf{G}_z^0 = \begin{pmatrix} G_{zx,x}^0 & 0 & G_{zx,z}^0 \\ 0 & G_{zy,y}^0 & 0 \\ G_{zz,x}^0 & 0 & G_{zz,z}^0 \end{pmatrix}$$

Thus, in layered media, equation (14) reduces to

$$\begin{aligned} u_{r'} &= +g_1 \star (M_{xx} \cos^2 \varphi + M_{yy} \sin^2 \varphi + M_{xy} \sin 2\phi) \\ &\quad +g_9 \star ((M_{xx} \sin^2 \varphi + M_{yy}) \cos^2 \varphi - M_{xy} \sin 2\phi) \\ &\quad +g_2 \star (M_{xz} \cos \phi + M_{yz} \sin \phi) \\ &\quad +g_3 \star M_{zz} \\ u_{\varphi'} &= +g_4 \star \left( \frac{1}{2} (M_{yy} - M_{xx}) \sin 2\varphi + M_{xy} \cos 2\varphi \right) \\ &\quad +g_5 \star (M_{yz} \cos \varphi - M_{xz} \sin \varphi) \\ u_{z'} &= +g_6 \star (M_{xx} \cos^2 \varphi + M_{yy} \sin^2 \varphi + M_{xy} \sin 2\phi) \\ &\quad +g_{10} \star ((M_{xx} \sin^2 \varphi + M_{yy}) \cos^2 \varphi - M_{xy} \sin 2\phi) \\ &\quad +g_7 \star (M_{xz} \cos \phi + M_{yz} \sin \phi) \\ &\quad +g_8 \star M_{zz} \end{aligned} \tag{22}$$

with the following elementary seismograms

$$\begin{aligned} g_1 &= G_{r'x,x}^0 & g_2 &= G_{r'x,z}^0 + G_{r'z,x}^0 & g_3 &= G_{r'z,z}^0 \\ g_4 &= G_{\varphi'x,y}^0 + G_{\varphi'y,x}^0 & g_5 &= G_{\varphi'y,z}^0 + G_{\varphi'z,y}^0 & & \\ g_6 &= G_{z'x,x}^0 & g_7 &= G_{z'x,z}^0 + G_{z'z,x}^0 & g_8 &= G_{z'z,z}^0 \\ g_9 &= G_{r'y,y}^0 & g_{10} &= G_{z'y,y}^0 & & \end{aligned} \tag{23}$$

Equation (22) and (23) represent the system of equations for a layered Earth. Only ten independent elementary seismograms  $g_1 - g_{10}$  are needed, of which components  $g_9$  and  $g_{10}$  contain only near field terms (e.g. Müller, 1985) and are neglected in most moment tensor studies of earthquakes.

### 3.4 Representation for body waves (peak amplitudes and amplitude ratios)

The number of independent Green's functions is further reduced if a ray theoretical approximation is valid, for instance, if we consider only far field body wave phases associated with a specific ray path (e.g., qP and qS phases). The ray direction (ray angle  $i$ , see Fig. 7), is related to the horizontal and vertical components of the slowness vector by

$$(s_x, s_y, s_z) = (s_x, 0, s_z) = |s| (\sin(i), 0, \cos(i)). \quad (24)$$

If we consider in the far field that  $M_{pq}(\xi_0) * G_{np,q}(x, \xi_0) \approx dM(\xi_0)/dt * G_{np}(x, \xi_0) s_q + O(\text{near field terms})$  (e.g. Dahm, 1996), the rotation of the zero azimuth Green's tensor leads to further simplified moment tensor representation for the three modes of body waves phases:

for qP

$$\begin{aligned} G_{z'x}^0 s_x &= g_6 = I_e^P \sin^2 i \\ G_{z'x}^0 s_z + G_{z'z}^0 s_x &= g_7 = I_e^P 2 \sin i \cos i \\ G_{z'z}^0 s_z &= g_8 = I_e^P \cos^2 i, \end{aligned}$$

for qSV

$$\begin{aligned} G_{r'x}^0 s_x &= g_1 = I_e^{SV} \cos i \sin i \\ G_{r'x}^0 s_z + G_{r'z}^0 s_x &= g_2 = I_e^{SV} (\cos^2 i - \sin^2 i) \\ G_{r'z}^0 s_z &= g_3 = -I_e^{SV} \sin i \cos i, \end{aligned} \quad (25)$$

and for qSH

$$\begin{aligned} G_{\varphi'x}^0 s_y + G_{\varphi'y}^0 s_x &= g_4 = I_e^{SH} \sin i \\ G_{\varphi'y}^0 s_z + G_{\varphi'z}^0 s_y &= g_5 = I_e^{SH} \cos i, \end{aligned}$$

where  $I_e^P$ ,  $I_e^{SV}$  and  $I_e^{SH}$  now define elementary wave mode phases (e.g. seismogram wavelets containing specific body wave phases) which consider the geometrical divergence, the attenuation and the travel time of the phase along the ray path between source and receiver. Note that  $G_{\varphi'x}^0 s_y = 0$  and  $G_{\varphi'z}^0 s_y = 0$  (see Dahm, 1996). Insertion in equation (14) leads to (see also Box 9.13 in Aki and Richards, 2002)

P wave:

$$\begin{bmatrix} u_{r'}^P \\ u_{z'}^P \end{bmatrix} = \begin{bmatrix} (I_e^P)_{r'} \\ (I_e^P)_{z'} \end{bmatrix} \star \left[ (M_{xx} \cos^2 \varphi + M_{yy} \sin^2 \varphi + M_{xy} \sin 2\phi) \sin^2 i \right. \\ \left. + (M_{xz} \cos \phi + M_{yz} \sin \phi) 2 \sin i \cos i \right. \\ \left. + M_{zz} \cos^2 i \right]$$

SV wave:

$$\begin{bmatrix} u_{r'}^{SV} \\ u_{z'}^{SV} \end{bmatrix} = \begin{bmatrix} (I_e^{SV})_{r'} \\ (I_e^{SV})_{z'} \end{bmatrix} \star \left[ (M_{xx} \cos^2 \varphi + M_{yy} \sin^2 \varphi + M_{xy} \sin 2\phi) \sin i \cos i \right. \\ \left. + (M_{xz} \cos \phi + M_{yz} \sin \phi) (\cos^2 i - \sin^2 i) \right. \\ \left. - M_{zz} \sin i \cos i \right] \quad (26)$$

SH wave:

$$u_{\varphi'}^{SH} = (I_e^{SH})_{\varphi'} \star \left[ \left( \frac{1}{2} (M_{yy} - M_{xx}) \sin 2\varphi + M_{xy} \cos 2\varphi \right) \sin i \right. \\ \left. + (M_{yz} \cos \varphi - M_{xz} \sin \varphi) \cos i \right]$$

Note that for each body wave phase of qP, qSV or qSH type only one elementary “phase-wavelet” enters the moment tensor representation. Equation (26) forms the basis for the well known polarity method, where for each phase only the polarity is considered to constrain the focal solution of a double couple source (see chapter 3). It is also the basis for a moment tensor inversion based on generalized rays (e.g. Nabelek, 1984), peak amplitude (e.g. Ebel and Bonjer, 1990), and amplitude ratios or the relative moment tensor inversion (e.g. Dahm, 1996).

Equation (26) can be used for body wave peak amplitude moment tensor inversion of weak and moderate events with short source duration. For instance, peak amplitudes or waveform moments (=displacements integrated over the first pulse) of low pass filtered P and S phases may be measured and used as input data for inversion. Low pass filtering with a passband below the corner frequency of the event has to be performed to avoid amplitude distortion from rupture directivity. Seismograms need to be deconvolved to ground motion to estimate true seismic moments. The geometrical ray parameters, i.e. the ray azimuth and take-off angle, are typically taken from a location program where ray tracing is performed. Although elementary seismograms  $I_e$  are reduced to a scalar factor comprising geometrical and intrinsic attenuation, and therefore might be estimated by ray tracers during location, it is recommended to extract  $I_e$  from complete synthetic seismograms which are filtered in the same way as the observed seismograms. This may avoid bias from apparent polarity flipping, a problem that may occur if short period stations were used. The method has been used occasionally for short period local and teleseismic earthquakes studies but is now of lesser relevance since broadband data are often available. Data problems were often difficult to identify since observed and synthetic waveforms were not directly compared. In order to further stabilize the method, pure double couple constraints (shear crack constraints) can be introduced.

A derivative of the peak amplitude method is the amplitude ratio method to determine focal solutions for double couple sources. Amplitude ratios between P, SV and SH phases are used as input data, together with P phase polarities, and the associated relations between the

elementary seismograms  $I_e$  of the different wave modes are considered from ray theory. A commonly used Fortran package is *focmec* (by A. Snoke, see <http://www.iris.edu/pub/programs/focmec>).

### 3.5 Relative moment tensor inversion using multiple events (body wave method)

To locate earthquakes multiple event inversions were developed. Well-known examples are the joint hypocenter, the master event location and the hypoDD methods (IS11.1). A moment tensor approach similar to the master event location was developed by e.g. Dahm (1996). The method is of interest if earthquake clusters are analyzed. Spatially clustered seismicity may occur for example during aftershock sequences, for earthquakes swarms in volcanic and geothermal areas, or at specific structures at tectonic faults or subducted plates (e.g., deep earthquake clusters). The concept of the relative moment tensor inversion is to use a reference mechanism (reference event) from the cluster, which is fixed during inversion, and to estimate the moment tensors of the studied events relative to the reference event. Since the studied and the reference earthquake from a tiny cluster have similar elementary seismograms  $I_e$ , the  $I_e$  factors can be eliminated from equation (26). Geometrical and intrinsic attenuation, amplification from site effects or unknown instrument functions have thus no or only a minor influence on the moment tensor result. The method is further described in Dahm (1996) and Dahm and Brandsdottir (1997). Necessary pre-requisites for application are that

- (a) the mechanism of the reference event is known a priori,
- (b) the events are tightly clustered and seismograms are low-pass filtered, so that events can be regarded as temporal and spatial point sources within approximately one wavelength from the reference event,
- (c) seismograms are low-pass filtered prior to measuring peak amplitudes. The corner frequency of the low-pass filter should be in the range of the corner frequency of the largest studied event of the cluster (potentially limiting the size of the smallest event that can be analyzed).

Dahm (1996) describes an extension of the method, for which the necessity of a fixed reference mechanism can be dropped when the radiation patterns within the earthquake cluster differ sufficiently.

### 3.6 Full waveform method using amplitude spectra

Full waveform methods considering broadband seismograms are nowadays routinely in use to study moderate and strong earthquakes at regional and teleseismic distances. As discussed in the previous paragraphs unconsidered phase shift between Green functions and data may be present if the centroid coordinates are incorrect or if the Earth model to calculate Green's functions is too simplified. Amplitude spectra inversions neglect the phase information. In the frequency domain, equation (15) can be written as

$$\tilde{d}(\omega) = \tilde{G}'_{nk}(\omega) y_k(\omega) = \left[ \text{Re}\{\tilde{G}'_{nk}\} + j\text{Im}\{\tilde{G}'_{nk}\} \right] y_k \quad , \quad (27)$$

where  $\text{Re}\{\}$  and  $\text{Im}\{\}$  are the real and imaginary part of the complex spectrum and  $j$  is the imaginary unit. The absolute value of the spectral seismogram amplitudes are then

$$|\tilde{d}| = \left[ \left( \text{Re}\{\tilde{G}'_{nk}\} y_k \right)^2 + \left( \text{Im}\{\tilde{G}'_{nk}\} y_k \right)^2 \right]^{1/2} = f(\tilde{G}', y) \quad . \quad (28)$$

The amplitude spectrum (28) is a non-linear function  $f$  of the moment tensor components  $y$ . The problem can be linearized and solved by an iterative inversion scheme (e.g., the gradient method). Equation (28) is linearized by means of the linear term of a Taylor's series expansion around a starting moment tensor solution. During each iteration, changes of moment tensor components are solved by minimizing the residual data vector. Local minima can be easily avoided if different starting models are considered by means of a grid search, e.g. represented by double couple starting solutions with different orientations. Examples are provided in e.g. Dahm et al. (1999 and Cesca et al. (2006).

Amplitude spectra inversion is in generally more robust and independent of phase misalignments as long as the time windows taken for inversion contain the phases modeled by Green's functions. The most serious drawback of amplitude spectra inversion is that the polarity of the source mechanism is not used resulting in a set of two indistinguishable moment tensor solutions with common nodal planes but different polarities. Therefore, amplitude spectra inversion is often implemented as a first step inversion, which is refined in a second step by considering additional constraints. For instance, polarity information of first onsets of body wave phases may be used to resolve the polarity ambiguity. A more elegant waveform approach is to use the results from the first inversion to estimate phase shifts at selected high quality stations, e.g. by means of cross correlation between observed and predicted waveforms, and to add these time domain traces in a second inversion step (e.g. Heimann, 2011; Cesca et al., 2010). The second step may either use time traces of phase corrected waveforms only or may be run as a weighted mixed input inversion considering amplitude spectra at all stations and time traces at selected key stations only. Since computation time and memory is not an issue any more we recommend the two-step inversion, which can easily be automated and reduces the problem of misaligned phases due to incorrect Earth models or earthquake locations.

### 3.7 Moment tensor representation using eigen-vibrations and eigen-modes of the Earth

The moment tensor representation calculates ground displacement as a sum of weighted Green functions (elementary seismograms), where weights are represented by moment tensor components. How the Green's functions are calculated is not specified. Most often, surface or body waves are used for moment tensor inversion. Body and surface waves in any finite body can be understood as a sum over a large number of eigenvibrations of this body. Therefore, Green's functions and synthetic seismograms may be calculated by summing up normal modes of the Earth. Such an approach was used by the first systematic global centroid moment tensor program led by A. Dziewonski in the early 1980's at Harvard University inverting fundamental modes and long period surface waves ( $> 135$  s). Other types of representation formulas may be derived for the normal mode theory, which are not further specified here (see, e.g., Dahlen and Tromp, 1998). Normal mode theory, for a long time, has been the most complete approach to calculate synthetic seismograms and Green's functions, since self gravitation, Earth rotation, Earth ellipticity and pre-stress are all included.

### 3.8 Codes to calculate Green functions for moment tensor inversion

Depending on the mode and type of selected data different computer codes can be recommended to calculate synthetic Green functions. The list below is incomplete and represents a selection based on the author's experience.

**Table 2** Computer codes to calculate synthetic seismograms

| Input data                                                                    | Type of code                                                                                                                                                           | Name of code          | Link, reference                                                                                                                                                                                                                                    |
|-------------------------------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------|-----------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Local and regional full waveforms (body and surface waves), high frequencies. | Reflectivity, layered Earth models, anisotropy & intrinsic attenuation, receiver at depth                                                                              | Qseis (fortran code)  | Wang (1999), <a href="ftp://ftp.gfz-potsdam.de/pub/home/turk/wang/qseis2006-code+input.zip">ftp://ftp.gfz-potsdam.de/pub/home/turk/wang/qseis2006-code+input.zip</a>                                                                               |
| Global and regional full waveforms                                            | Numerical integration of the SODE (integration of minors)                                                                                                              | GEMINI (fortran code) | Friederich and Dalkomo (1995), <a href="http://www.quest-itn.org/library/software/gemini-greens-function-of-the-earth-by-minor-integration">http://www.quest-itn.org/library/software/gemini-greens-function-of-the-earth-by-minor-integration</a> |
| Global and regional full waveforms, free oscillations, tsunami, infrasound    | Hybrid method (analytical and numerical integration) for layered Earth models. self gravitation, ocean layer, atmospheric layer, Earth normal modes, long time series. | QSSP (fortran code)   | Wang, R. (1997), <a href="ftp://ftp.gfz-potsdam.de/pub/home/turk/wang/qssp2010-code+input.rar">ftp://ftp.gfz-potsdam.de/pub/home/turk/wang/qssp2010-code+input.rar</a>                                                                             |
| Free oscillations, long period surface waves                                  | Numerical integration, layered Earth models, self gravitation, anisotropy, attenuation.                                                                                | MINEOS (fortran)      | <a href="http://www.geodynamics.org/ci/g/software/mineos">http://www.geodynamics.org/ci/g/software/mineos</a>                                                                                                                                      |
| Full waveforms                                                                | Numerical approach, 3D Earth models, ellipticity, rotation, self gravitation, attenuation.                                                                             | SPECFEM3D             | <a href="http://www.geodynamics.org/ci/g/software/specfem3d-globe">http://www.geodynamics.org/ci/g/software/specfem3d-globe</a>                                                                                                                    |

Green's function can be pre-calculated for specific Earth models (e.g. standard models) and stored in Green's function databases. They can then be shared over the internet, e.g. on the KINHERD web page (<http://kinherd.org>), from where they can be downloaded by users, for instance if computing power or facilities are not available to build own databases. Exercise EX 3.6 gives an example, referring to moment tensor inversion based on the Pyrocko package (<http://emolch.github.com/pyrocko>) or the Kiwi Tools (<http://kinherd.org/kiwitools>) (e.g. RAPIDINV, see <http://kinherd.org/rapidinv>), which can directly use these pre-calculated databases.

## 4 Moment tensor inversions schemes in practice

Solving equation 3 in the time or in the frequency domain includes several steps from data preparation and Green's function calculation to solving a system of linear equations. The reliability and evaluation of the outcome, i.e. a moment tensor, depends not only on the datasets and a formal error estimation but on many additional factors, where the most important are listed below.

### 1) Data acquisition and pre-processing

- selection of seismograms with good signal-to-noise ratio
- unclipped signals
- good azimuthal coverage
- removing mean value and linear trends, detection of data gaps
- correcting for instrument response, converting seismograms to displacement, velocity or acceleration
- selection of waveforms, e.g., P, S and surface wave windows or full seismograms
- low-pass filtering appropriate to the dataset to remove high-frequency noise and estimating the corner frequency of the source to safely satisfy the point source approximation

### 2) Calculation of accurate synthetic Green functions for specific

- earth model
- location of the source
- receiver position

### 3) Inversion and interpretation of the inversion result

- decision for an inversion algorithm to solve eq. 3 or eq. 12
- test of resolution for all free parameters of the source model
- decomposition of the moment tensor, e.g., into best double couple plus CLVD

The inversion may be performed in the time or frequency domain either using a singular value decomposition to calculate the generalized inverse matrix or, e.g., by LU decomposition. We do not provide details: the reader may find more information and software in Press et al. (2007) (chapter 2, especially 2.3). A general introduction and overview on data inversion can be found in the textbooks of Menke (1989) and Aster et al. (2013). Linearization and iterative solution strategies using one specific or a set of start models are required for instance to solve eq. 12 (amplitude spectra based solution).

Care must be taken to match the synthetic and observed seismograms. Alignment of observed and synthetic waveforms is facilitated by cross-correlation techniques. In most moment-tensor inversion schemes, focal depth is assumed to be constant. The inversion is done for a range of focal depths and the best solution is assumed to be the with the lowest misfit.

Dufumier and Cara (1995) and Dufumier (1996) give a systematic overview for the how  $M_{ij}$  results are affected by differences in the azimuthal coverage and by using only P waves, P plus SH waves or P and SH and SV waves for the inversion with body waves. Their results indicate the most important factor is good azimuthal coverage, followed by the combined use of P and S phases (and Love and Rayleigh waves). A systematic overview with respect to the effects caused by an erroneous velocity model for the Green function calculation and the



effects due to wrong hypocenter coordinates can be found in Šilený et al. (1992), Šilený and Pšenčík (1995), Šilený et al. (1996) and Kravanja et al. (1999).

The dynamical boundary conditions on the solid free surface imply that the moment tensor elements  $M_{xz}$  ( $= M_{13}$ ) and  $M_{yz}$  ( $= M_{23}$ ) vanish as the source location approaches the Earth's free surface or the seafloor. As a result, these vertical dip-slip components are not well constrained for shallow-focus sources. A discussion regarding the related uncertainties for the determination of focal mechanisms, the centroid depth but also the scalar seismic moment can be found in Bukchin et al. (2010).

Depending on the dataset the isotropic component,  $M_{iso}$ , may be difficult to resolve. Therefore, an isotropic component with strength of less than 10% of the strength of the full moment tensor is often considered to be insignificant. A better approach is to test the statistical significance, e.g. by means of an F-test.

Significant CLVD components are often reported for large intermediate-depth and very deep earthquakes. In many cases, however, it can be shown that these may be caused by the superposition of several sub-events with different double-couple mechanisms (Kuge and Kawakatsu, 1990; Frohlich, 1995; Tibi et al., 2001).

How good is a specific solution? Several standard approaches exist to get an answer to this question. The numerical stability of the inversion can, for instance, be checked with the condition number. This number is defined as the ratio between the square root of the largest eigenvalue to the smallest eigenvalue of the generalized inverse and is an outcome of calculating the generalized inverse using a singular value decomposition method. It indicates the overall sensitivity of the solution to errors in the data. For iterative inversion schemes straightforward error estimation techniques are not applicable. One way to estimate the errors in the solutions is the use of many start models within reasonable limits and subsequent evaluation of the scatter in the resulting model parameters. Jackknifing and bootstrapping (see, e.g., Shao and Tu, 1995) are also widely used methods to test the stability of solutions, especially when non-linear iterative schemes are used (e.g., Heimann, 2011; Zahradník and Custódio, 2012). See also Valentine and Trampert (2012) for a recent assessment of uncertainties for centroid moment tensor determinations.

## 5 Free software packages for moment tensor inversion

Here we list freely available software packages for moment tensor inversion. We mention briefly differences regarding the distance range and details of the inversion outcome.

- RAPIDINV (Cesca, 2010), based on the Kiwi Tools (Heimann, 2011): Time and frequency domain inversion code in the regional and teleseismic distance range using full waveform Green's function databases. Includes inversion for point source, simple finite sources and extensive resolution testing, see <http://kinherd.org/rapidinv> and exercise 3.6 of the NMSOP-2: „A practical in moment tensor inversion using the Kiwi tools“ (<http://gfzpublic.gfz-potsdam.de/pubman/item/escidoc:130023:6>)
- VOLPIS (Cesca and Dahm, 2007). A stand-alone frequency domain inversion code (fortran) for the study of time-dependent source parameters for long period volcanic

signals. Single forces and dipole forces (moment tensors) can both be inverted (<http://kinherd.org/volpis>).

- Time-Domain Moment Tensor INVerse Code (TDMT-INVC) by Dreger (2003), for a description and software download see [www.orfeus-eu.org/pub/software](http://www.orfeus-eu.org/pub/software)
- Regional moment tensor inversion code together with tutorial and examples by B. Herrmann see [www.eas.slu.edu/eqc/eqccps.html](http://www.eas.slu.edu/eqc/eqccps.html)
- ISOLA-GUI which is a matlab GUI for regional moment tensor inversion written by J.Zahradnik and E. Sokos, see <http://seismo.geology.upatras.gr/isola>
- mtinvers (Dahm et al., 2007) is a stand-alone fortran code for regional and teleseismic point source inversion. Inversion can be constrained to opening/closing cracks, double couple source, shear source (<http://kinherd.org/mtinvers>).
- Mtnv package written by C. Ammon and G. Randall for point source moment tensor inversion in time domain in the regional distance range, see <http://eqseis.geosc.psu.edu/~cammon>
- MT5 package (McCaffrey et al., 1991) is a PC based software package for the inversion of teleseismic body waves for moment tensor solutions or double couple fault plane solutions, see <http://ees2.geo.rpi.edu/rob/mt5> and can also be found in the IASPEI Software Library, number 4.

Differences between the available packages arise from the implementation of constraints in the inversion either to stabilize the solution or for special purposes. As was mentioned in the subsection on problematic sources and moment tensor decomposition it is for many problems desirable to constrain the moment tensor to its deviatoric part by setting the trace of the moment tensor equal to zero. In this case only 5 moment tensor elements are left and the corresponding linear equation system can be found by recombining terms in the Eqs.s (3) and (6,7). This problem is still a linear one. Other constraints used are: double couple, tensile crack or a mixed situation. In these cases the inversion becomes non-linear and iterative inversion schemes have to be used.

The offered codes differ in several aspects. In the teleseismic distance range differences arise mainly regarding the computation of full wavefield Greens functions and the use of complete seismograms and/or if the inversion is performed for specific body wave phases like P and S and if surface waves are included. Here again an important detail is if the flat Earth transformation (Chapman 1973; Müller, 1977; Bhattacharya, 1996) is used to account for the sphericity of the Earth or if full wavefield Green's functions are calculated directly for the spherical Earth or if normal modes are used (see chapter 2.5 for freely available codes for the computation of full wavefield Green's functions).

In the regional distance range more codes are available to calculate full wavefield synthetics for layered media. The use of a regionally valid velocity model is critical (e.g. Donner et al., 2013). It should be mentioned that many velocity models used for location may show problems when used to produce longperiod surface waveforms because they may have been tuned to fit the crustal and upper mantle P-wave velocities not the S-wave velocities in the uppermost crustal layers.

Some codes use a full grid search regarding centroid time, centroid location and centroid depth: most codes at least search for a best fitting centroid depth.

Exercise EX 3.6 of the NMSOP-2 (DOI: 10.2312/GFZ.NMSOP-2\_EX\_3.6) provides a tutorial on how to setup Green's function databases and the MT inversion by using the Kiwi Tools.

## 6 Moment tensor catalogues

In the late 1970s and early 1980s the group headed by A. Dziewonski at Harvard University began routine determination of the seismic source parameters of all earthquakes with magnitude  $\geq 5.5$  using the centroid moment tensor (CMT) method (Dziewonski et al., 1981). This inversion tries to fit simultaneously the very long-period body wave train from the P arrival until the onset of the fundamental modes and mantle waves with periods longer than 135 s using synthetics produced by normal mode summation (see chapter 2.6). In addition the best point-source parameters (centroid epicentral coordinates, centroid depth and origin time) and the six independent moment tensor elements were published.

The Harvard group maintained, from 1981 to 2006, an extensive catalog of centroid moment-tensor (CMT) solutions for strong (mainly  $M \geq 5.5$ ) earthquakes since 1976. This catalog is now maintained and continued by the Global Centroid Moment Tensor Project at Lamont-Doherty Earth Observatory (LDEO) of Columbia University. Their solutions, as well as quick CMT solutions of recent events, down to moment magnitudes  $M_W \approx 4.5$  can be viewed at <http://www.globalcmt.org>. The products are also distributed by IRIS (see <http://www.iris.edu/spud/momenttensor>) and other data centers like the USGS. Harvard CMT solutions for the very largest, gigantic earthquakes (e.g.,  $M_W$  9.3 Sumatra, 26 December 2004) turned out to be systematically deficient (Stein and Okal, 2005; Tsai et al., 2005), and provisions have now been made to use longer periods for mega earthquakes.

As mentioned above the CMT inversion procedure seeks a solution for the centroid location of the earthquake. This centroid location may, for very large earthquakes, significantly differ from the hypocenter location based on arrival times of the first P-wave onsets. This initial hypocenter location corresponds to the place where rupture started. Therefore, the offset of the centroid location relative to the hypocentral location gives a first indication on fault extent and rupture directivity. In the case of the August 17, 1999 Izmit (Turkey) earthquake the centroid was located about 50 km east of the P-wave hypocenter. The centroid location coincided with the area where the maximum surface ruptures were observed.

Several institutions publish in recent years moment tensors determined using the W-phase (Kanamori, 1993; Kanamori and Rivera, 2008; Duputel et al., 2012). The ultra-longperiod W phase follows direct P on very broadband seismograms. Its use has the advantage to provide correct spectral levels for scalar moment determination also for slow ruptures what is important due to their high tsunami potential. See [eost.u-strasbg.fr/wphase](http://eost.u-strasbg.fr/wphase) for a description of the method, the catalog and a list of further publications.

Several other institutions offer regularly moment tensor solutions. The following list is not complete and what is described may change in detail.

One of the most complete regional archives is published by Berkeley Seismological Laboratory (BSL) for Northern California based on inversion of full waveform synthetics mainly of mid- and longperiod surface waveforms (Dreger and Helmberger (1993) and Dreger (2003). The catalog reaches back till 1993. More references to methodological work and example cases can be found on the webpage [seismo.berkeley.edu/mt](http://seismo.berkeley.edu/mt).

Another regional catalog for Southern California is published by the Southern California Earthquake Data Center (SCEC) since 2000 and can be found at <http://www.data.scec.org>.

GEOAZUR-SCARDEC method. GEOAZUR offers near realtime moment tensor solutions for larger worldwide events using a body wave deconvolution method (Vallee et al., 2011).

An estimate for a point source source time function and plots showing the fit of synthetics and observed waveforms for P and SH are provided.

See <https://geoazur.oca.eu/spip.php?article1236>

GFZ worldwide moment tensor solutions are distributed since 2011. Solutions are based on a modification of the code of Dreger and Helmberger (1993) and Dreger (2003) and are given in text or graphical form. A catalogue can be found at <http://geofon.gfz-potsdam.de/eqinfo/list.php?mode=mt>

A fully automatic global moment tensor catalogue, since 2012, is provided by GFZ using the Kiwi Tools package. Uncertainties are evaluated using bootstrapping methods. Kinematic source solutions of the extended source are calculated for selected events (see Heimann, 2011). Catalogue can be found under <http://kinherd.org/quakes/>

The National Observatory of Athens (NOA) provides fast moment tensor solutions for events in the Mediterranean Sea region down to a magnitude of about Mw 3.5. The ISOLA software (Sokos and Zahradník, (2008, 2013). is used and a grid search to find the best centroid parameters is performed. Plots showing the fit of the synthetics to the waveforms are provided. See also

<http://bbnet.gein.noa.gr/HL/seismicity/moment-tensors>

INGV uses the Dreger (2003) time domain method to invert for moment tensors of earthquake in Italy and the surrounding regions. Solutions are provided in form of pdf files showing moment tensor solutions in text and graphical form. The quality of the solution can be judged from the fit of longperiod synthetic and data waveforms together with maps showing the azimuthal distribution of stations. See <http://cnt.rm.ingv.it/tdmt.html>

INGV Bologna provides a different catalog for earthquakes larger than Mw 4.0 for events in Italy and larger than Mw 4.5 for events in the wider region based on an adaption of the centroid moment tensor method to model mid-period surface waves (Pondrelli et al, 2006, 2011 ). Solutions since 1976 are provided in text and graphical form.

The SED/ETHZ provides an automatic regional moment tensor catalogue for the entire European-Mediterranean region. Until 2010 the method was based on Braunmiller et al. (2002) and Bernardi et al. (2004). The reviewed catalog covers 1999 to 2006, the automatic MT catalog 2002-2010 (see <http://www.seismo.ethz.ch/prod/tensors/index>). The current regional MT determination covers Switzerland and vicinity and is based on the Dreger code and is described in Clinton et al., 2006).

Regional moment tensor solutions for Spain and its surroundings are provided by IGN, see e.g. <http://www.ign.es/ign/layoutIn/sismoPrincipalTensorZonaAnio.do> Solutions are provided in text (Spanish language) and graphical form. A number for the variance reduction and waveform fits are provided.

The USGS provides several types of moment tensor solutions. The „fast“ moment tensor solutions are based on body wave inversion only. The USGS centroid moment tensors are resulting from longperiod full waveform inversion. Additionally the USGS publishes W-phase moment tensors from inversion of the ultra-longperiod W phase following direct P on very broadband seismograms. The latter has the advantage to provide correct spectral levels

for scalar moment determination also for slow ruptures important due to their tsunami potential. See the USGS homepage (in the moment under reconstruction)

An interesting approach using a systematic grid search over possible time and spatial centroid positions in the regional distance range and subsequent moment tensor inversion on longperiod time domain waveforms is used by H. Kawakatsu (ERI Tokyo University) and is published on [http://www.eri.u-tokyo.ac.jp/GriD\\_MT/](http://www.eri.u-tokyo.ac.jp/GriD_MT/).

The following table summarizes specific aspects of the published catalogs.

**Table 3** list of published moment tensor catalogs

| Service   | Range                                                  | Data, bandwidth                                                                                                                                                                                                                                                | Greens functions          | Constraints | Additional parameter       | Error bounds                                          | Inversion type         |
|-----------|--------------------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|---------------------------|-------------|----------------------------|-------------------------------------------------------|------------------------|
| CMT       | worldwide since 1976<br>$M > 5$                        | long-period BW and SW,<br>40-150 s                                                                                                                                                                                                                             | normal mode summation     | deviatoric  | centroid location and time | provided for MT elements                              | manually revised       |
| GFZ       | worldwide since 2011<br>$M > 4.5$                      | complete seismograms<br>$4.5 < M < 5.5$<br>BW and SW 30-80 s<br>$5 < M < 6.5$<br>BW 40-100 s, SW 60-150 s<br>$6 < M < 7$<br>BW 60-150, SW 80-200 s<br>$M > 6.5$<br>BW and SW 90-300 s<br>$M > 7.5$<br>BW and SW 200-600 s<br>$M > 8.5$<br>BW and SW 200-1000 s | reflectivity method       | deviatoric  | centroid depth             |                                                       | automatic plus revised |
| GFZ-KIWI  | worldwide since 2012<br>$M > 5.5$                      | BW and SW<br>20-100 s                                                                                                                                                                                                                                          | Gemini & QSSP             | deviatoric  | centroid location and time | yes                                                   | automatic              |
| USGS      | worldwide                                              | Longperiod BW                                                                                                                                                                                                                                                  | ray synthetics            | deviatoric  | centroid depth             | variance reduction                                    | automatic and revised  |
| USGS CMT  | worldwide                                              | Complete longperiod waveforms                                                                                                                                                                                                                                  | full waveform synthetics  | deviatoric  | centroid location and time | variance reduction                                    | automatic and revised  |
| USGS, ERI | worldwide since 1990<br>$M > 6.5$                      | W-phase<br>$5.5 < M < 6.5$ 50-150 s<br>$6.5 < M < 7$ 100-500 s<br>$7 < M < 7.5$ 100-600 s<br>$7.5 < M < 8$ 200-600 s<br>$M \geq 8.0$ 200-1000 s                                                                                                                | full waveform synthetics  | deviatoric  | centroid location and time | variance reduction                                    | automatic and revised  |
| BSL       | regional Northern California) since 1993<br>$M > 3.5$  | complete long-period waveforms<br>$M > 3.5$ , 10-50 s<br>$3.5 < M < 5$ 20-50 s<br>$M > 5$ 20-100 s                                                                                                                                                             | full waveform synthetics  | deviatoric  | centroid location and time | variance reduction, comparison of data and synthetics | Automatic and revised  |
| SCEC      | regional (southern California) since 2000<br>$M > 3.5$ | complete long-period waveforms                                                                                                                                                                                                                                 | full wave form synthetics | deviatoric  | Centroid location and time | variance reduction, comparison of data and synthetics | automatic and revised  |
| INGV      | regional (Mediterranean) $M > 3.5$                     | SW spectra and waveforms                                                                                                                                                                                                                                       | full waveform synthetics  | deviatoric  | centroid location and time | comparison of data and synthetics                     | automatic and revised  |

|     |                                                                 |                                                                                             |                          |            |                            |                                                       |           |
|-----|-----------------------------------------------------------------|---------------------------------------------------------------------------------------------|--------------------------|------------|----------------------------|-------------------------------------------------------|-----------|
| SED | regional<br>(Switzerland and vicinity)<br>since 1999<br>M > 3.0 | complete long-period waveforms                                                              | full waveform synthetics | deviatoric | centroid depth             | comp. of data and synthetics                          | automatic |
| IGN | regional<br>(Spain and surroundings)<br>since 2002<br>M > 3.5   | complete long-period waveforms<br>M>3.5<br>10-50 s<br>3.5<M<5<br>20-50 s<br>M>5<br>20-100 s | full waveform synthetics | deviatoric |                            | variance reduction, comparison of data and synthetics | automatic |
| ERI | regional<br>(Japan)<br>since 2003<br>M > 3.5                    | long-period complete waveforms<br>LP at 10 s                                                | full waveform synthetics | deviatoric | centroid location and time | variance reduction                                    | automatic |

## 7 Acknowledgments

Many students, course participants and colleagues helped improving the material over the years. We thank Peter Bormann for his persistent requests and reminders to deliver and improve the information sheets. We acknowledge the suggestions for improvement and careful reviews of Jochen Braunmiller, Daniela Kühn, Peter Bormann and Stefanie Donner.

## 8 References

- Aki, K., and Richards, P. G. (2002). Quantitative seismology. Second Edition, ISBN 0-935702-96-2, *University Science Books*, Sausalito, CA., 700 pp.
- Aster, R. C., Borchers, B., and Thurber, C. H., (2013). *Parameter estimation and inverse problems*, 2nd edition, Elsevier, ISBN: 978-0-12-385048-5.
- Bhattacharya, S., 1996. Earth-flattening transformation for P-SV waves (short note). *Bull. Seism. Soc. Am.*, **86**,1979–1982.
- Ben-Menahem, A., and Singh, S.J. (2000). *Seismic Waves and Sources*. 2<sup>nd</sup> edition, Dover Publications, New York.
- Bernardi, F., Braunmiller, J., Kradolfer U., and Giardini, D. (2004). Automatic regional moment tensor inversion in the European-Mediterranean region. *Geophys. J. Int.* , **157** (2): 703-716; doi: 10.1111/j.1365-246X.2004.02215.x.
- Bukchin, E., Clevede, E., and Mostinskiy, A. (2010) Uncertainty of moment tensor determination from surface wave analysis for shallow earthquakes. *J. Seismol.*; doi: 10.1007/s10950-009-9185-8.
- Braunmiller J., Kradolfer U., Baer M., and Giardini D. (2002). Regional moment tensor determination in the European-Mediterranean region-initial results. *Tectonophysics*, **356**, 5–22.
- Cesca, S., Bufo, E., and Dahm, T, (2006). Amplitude spectra moment tensor inversion of shallow earthquakes in Spain. *Geophys. J. Int.*; doi:10.1111/j.1365-246X.2006.03073.
- Cesca, S., and Dahm, T. (2007). A frequency domain inversion code to retrieve time-dependent parameters of very long period volcanic sources. *Computer and Geosciences*; doi: 10.1017/j.cageo.2007.03.018.

- Cesca, S., Heimann, S., Stammler, K., and Dahm, T. (2010). Automated procedure for point and kinematic source inversion at regional distances. *J. Geophys. Res.*, doi:10.1029/2009JB006450
- Cesca, S. and Heimann, S., and Dahm, T. (2010b). Rapid directivity detection by azimuthal spectra inversion. *J. Seismol.*, 15(1), 147-164; doi: 10.1007/s10950-010-9217-.
- Chapman, C. (1973). The earth flattening transformation in body wave theory. *Geophys. J. R. Astron. Soc.*, 35, 55–70.
- Clinton, J. F., Hauksson, E., and Solanki, K. (2006). An evaluation of the SCSN moment tensor solutions: Robustness of the  $M_W$  magnitude scale, style of faulting and automation of the method. *Bull. Seism. Soc. Am.*, 96 (5), 1689-1705.
- Dahlen, F. A., and Tromp, J. (1998). *Theoretical Global Seismology*. Princeton University Press, pp. 944..
- Dahm, T. (1993). Relative moment tensor inversion to determine the radiation pattern of seismic sources (in German). PhD thesis, Institute for Geophysics, University of Karlsruhe, Germany, 121 pp.
- Dahm, T. (1996). Relative moment tensor inversion based on ray theory: theory and synthetic tests. *Geophys. J. Int.*, 124, 245-257.
- Dahm, T., and Brandsdottir, B. (1997). Moment tensors of micro-earthquakes from the Eyjafjallajökull volcano in South Iceland *Geophys. J. Int.*, 130, 183-192.
- Dahm, T., and Manthei, G. and Eisenblätter, J. (1999). Automated moment tensor inversion to estimate source mechanisms of hydraulically induced micro-seismicity in salt rock. *Tectonophysics.*, 306, 1-17.
- Dahm, T., Krüger, F., Stammler, K., Klinge, K., Kind, R., Wylegalla, K., Grasso, J.-R. (2007). The 2004  $M_w$  4.4 Rotenburg, Northern Germany, earthquake and its possible relationship with gas recovery. *Bull. Seism. Soc. Am.*, 97(3), 691-704.
- Das, S., and Kostrov, B. V. (1988). *Principles of earthquake source mechanics*. Cambridge University Press.
- Donner, S., Rößler, D., Krüger, F., Ghods, A. and Strecker, M. (2013). Segmented seismicity of the  $M_w$  6.2 Baladeh earthquake sequence (Alborz mountains, Iran) revealed from regional moment. *J. Seismol.*, 17, 925-959; doi: 10.1007/s10950-013-9362-7.
- Doornbos, D. J. (Ed.) (1988). *Seismological algorithms, computational methods and computer programs*. Academic Press, New York, xvii + 469 pp.
- Dreger, D. S., and Helmberger, D. V. (1993). Determination of source parameters at regional distances with three-component sparse network data, *J. Geophys. Res.*, 98, no. B5, 8107–8126.
- Dreger, D. S. (2003). 85.11 TDMT\_INV: Time Domain Seismic Moment Tensor INVersion, In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.). *International Handbook of Earthquake and Engineering Seismology*, Academic Press, Amsterdam, Part B, p 1627.
- Dufumier, H., and Cara, M. (1995). On the limits of linear moment tensor inversion of surface wave spectra. *Pageoph*, 145, 235-257.
- Dufumier, H. (1996). On the limits of linear moment tensor inversion of teleseismic body wave spectra. *Pageoph*, 147, 467-482.
- Duputel, Z., Rivera, L., Kanamori, H., and Hayes, G. (2012). W-phase fast source inversion for moderate to large earthquakes (1990-2010), *Geophys. J. Int.*, 189(2), 1125-1147.

- Dziewonski, A. M., Chou, T.-A., and Woodhouse, J. H. (1981). Determination of earthquake source parameters from waveform data for studies of global and regional seismicity. *J. Geophys. Res.*, **86**, 2825-2852.
- Ebel, J. E., and Bonjer, K. P. (1990). Moment tensor inversion of small earthquakes in southwestern Germany for the fault plane solution. *Geophys. J. Int.* **101**(1), 133-146.
- Ekström, G., Nettles, M., and Dziewonski, A. M. (2012). The global CMT project 2004-2010: Centroid-moment tensors for 13,017 earthquakes. *Phys. Earth Planet Int.*, **200-201**, 1-9.
- Ford, S., Dreger, D., and Walter, W. (2008). Source characterization of the August 6, 2007 Crandell Canyon Mine seismic event. *Seismol. Res. Lett.*, **79**, 637-644.
- Friederich, W., and Dalkolmo, J. (1995): Complete synthetic seismograms for a spherically symmetric earth by a numerical computation of the Green's function in the frequency domain. *Geophys. J. Int.*, **122**, 537-550.
- Frohlich, C., (1995). Characteristics of well-determined non-double-couple earthquakes in the Harvard CMT catalog. *Phys. Earth Planet. Inter.*, **91**, 213-228.
- Heimann, S. (2011). A Robust Method to estimate kinematic earthquake source parameters, Universität Hamburg, PhD thesis.
- Heimann, S., González, A., Cesca, S., Wang, R., Dahm, T. (2013). Seismic characterization of the terminal explosion of the Chelyabinsk meteor. *Geophys. Res. Lett.*, **84**(6), 1021-1025; doi: 10.1785/0220130042.
- Hudson, J. A., Pearce, R. G., and Rogers, R. M. (1989). Source type plot for inversion of the moment tensor. *J. Geophys. Res.*, **94** B1, 765-774.
- Jost, M. L., and Hermann, R. B. (1989). A students guide to and review of moment tensors. *Seism. Res. Lett.*, **60**, 37-57.
- Kanamori, H. (1993). W Phase, *Geophys. Res. Lett.*, **20**, 1691-1694.
- Kanamori, H., and Rivera, L. (2008). Source inversion of the W phase: speeding tsunami warning. *Geophys. J. Int.*, **175**, 222-238.
- Kennett, B. L. N., (1988). Systematic approximations to the seismic wavefields, In: Doornbos (1988), 237-259.
- Kikuchi, M., and Kanamori, H. (1991). Inversion of complex body waves – III. *Bull. Seism. Soc. Am.*, **81**, 2335-2350.
- Kravanja, S., Panza, G. F., and Šilený, J. (1999). Robust retrieval of a seismic point-source function. *Geophys. J. Int.*, **136**, 385-394.
- Krieger, L. (2011). Automated inversion of long period signals from shallow volcanic and induced seismic events. Dissertation, Fakultät für Mathematik, Informatik und Naturwissenschaften, Universität Hamburg.
- Krieger, L. and Heimann, S. (2012). MoPaD – Moment tensor plotting and decomposition: A tool for graphical and numerical analysis of seismic moment tensors. *Seism. Res. Lett.*, **83**(3), 589-595, doi: 10.1785/gssrl.83.3.589.
- Kuge, K., and Kawakatsu, H. (1990). Analysis of a deep “non double couple” earthquake using very broadband data, *Geophys. Res. Lett.*, **17**, 227-230.
- Lay, T., and Wallace, T. C. (1995). *Modern global seismology*. ISBN 0-12-732870-X, Academic Press, 521 pp.
- Menke, W. (1989). *Geophysical data analysis: discrete inverse theory*. Revised edition, Academic Press Inc., Orlando, Florida, 260 pp..



- Müller, G. (1973). Seismic moment and long period radiation of underground nuclear explosions. *Bull. Seism. Soc. Am.*, **3**, 847-857.
- Müller, G. (1977). Earth-flattening approximation for body-waves derived from geometric ray theory – Improvements, corrections and range of applicability. *J. Geophys.*, **42**, 429– 436.
- Müller, G. (1985). The reflectivity method: A tutorial. *J. Geophys.*, **58**, 153-174.
- Müller, G.(2001). Volume change of seismic sources from moment tensors. *Bull. Seism. Soc. Am.*, **91**, 880-884.
- Nabelek, J. (1984). Determination of earthquake source parameters from inversion of body waves. Ph.D. thesis, Mass. Inst. of Tech.
- Oncescu, M. C. (1986). Relative seismic moment tensor determination for Vrancea intermediate depth earthquakes, *Pageoph*, **124**, 931-940.
- Pondrelli, S., Salimbeni, S., Ekström, G., Morelli, A., Gasperini, P., and Vannucci, G. (2006). The Italian CMT dataset from 1977 to the present, *Phys. Earth Planet. Int.* **159**(3-4), 286-303; doi:10.1016/j.pepi.2006.07.008
- Pondrelli S., Salimbeni S., Morelli A., Ekström G., Postpischl L., Vannucci G., and Boschi E. (2011). European-Mediterranean Regional Centroid Moment Tensor Catalog: solutions for 2005-2008. *Phys. Earth Planet. Int.*; doi: 10.1016/J.PEPI.2011.01.007.
- Press, W. H., Teukolsky, S. A., Vetterlin, W. T., and Flannery, B. P. (2007). *Numerical Recipes: The Art of Scientific Computing, Third Edition*, Cambridge University Press (ISBN-10: 0521880688, or ISBN-13: 978-0521880688).
- Riedesel, M., and Jordan, Th. H. (1998). Display and assessment of seismic moment tensors. *Bull. Seism. Soc. Am.*, **79**(1), 85-100.
- Shao, J., and Tu, D. (1995). The Jackknife and Bootstrap. *Springer Series in Statistics*.
- Šílený, J., Panza, G. F., and Campus, P. (1992). Waveform inversion for point source moment retrieval with variable hypocentral depth and structural model. *Geophys. J. Int.*, **109**, 259-274.
- Šílený, J., and Pšenčík, I. (1995). Mechanisms of local earthquakes in 3-D inhomogeneous media determined by waveform inversion. *Geophys. J. Int.*, **121**, 459-474.
- Šílený, J., Campus, P., and Panza, G. F. (1996). Seismic moment tensor resolution by waveform inversion of a few local noisy records - I. Synthetic tests. *Geophys. J. Int.*, **126**, 605-619.
- Sipkin, S. A. (1982). Estimation of earthquake source parameters by the inversion of waveform data: synthetic waveforms. *Phys. Earth Planet. Inter.*, **30**, 242-259.
- Sokos, E., and Zahradník, J., (2008). ISOLA: a Fortran code and a Matlab GUI to perform multiple-point source inversion of seismic data, *Computers and Geosciences*, **34**, 967-977.
- Sokos, E., and Zahradník, J. (2013). Evaluating centroid moment tensor uncertainty in new version of ISOLA software. *Seismol. Res. Letters*, **84**, 656-665.
- Stein, S., and Okal, E. (2005). Speed and size of the Sumatra earthquake. *Nature* **434**, 581-582
- Stein, S., and Wysession, M. (2003). *An introduction to seismology, earthquakes and Earth structure*. Blackwell Publishing, Oxford.
- Tanioka, Y., and Ruff, L. (1997). Source time functions. *Seism. Res. Lett.*, **68**, 386-400.

- Tocheport, A., Rivera, L., and Chevrot, S. (2007). A systematic study of source time functions and moment tensors of intermediate and deep earthquake. *J. Geophys. Res.*; doi: 10.1029/2006JB004534
- Tibi, R., Bock, G., Xia, Y., Baumbach, M., Grosser, H., Milkereit, C., Karakisa, S., Zünbül, S., Kind, R., and Zschau, J. (2001). Rupture processes of the 1999 August 17 Izmit, and November 12 Düzce (Turkey) earthquakes. *Geophys. J. Int.*, **144**, F1-F7.
- Tsai, V. C., Nettles, M., Ekström, G., and Dziewonski, A., 2005. Multiple CMT source analysis of the 2004 Sumatra earthquake. *Geophysical Research Letters*, **32**; L17304, doi: 10.1029/2005GL023813.
- Tape, W., and Tape, C. (2012). A geometric comparison of source-type plots. *Geophysical J. Int.*, **190**, 499-510; doi: 10.1111/j.1365-246X.2012.05490.x.
- Udías, A. (1999). *Principles of Seismology*. Cambridge University Press, United Kingdom, 475 pp.
- Valentine, A.P., and Trampert, J. (2012). Assessing the uncertainties on seismic source parameters: Towards realistic error estimates for centroid-moment-tensor determinations, *Phys..Earth Planet. Inter.*, **210**, 36-49.
- Vallée, M., Charléty, J., Ferreira, A. M. G., Delouis, B., and Vergoz, J. (2011). SCARDEC: a new technique for the rapid determination of seismic moment magnitude, focal mechanism and source time functions for large earthquakes using body wave deconvolution, *Geophys. J. Int.*, **184**, 338-358.
- Vasco, D.W., (1989). Deriving source-time functions using principle component analysis. *Bull. Seism. Soc. Am.*, **79**, 711-730.
- Vavryčuk, V. (2001). Inversion for parameters of tensile earthquakes. *J. Geophys. Res.*, **106**(B8), 16339–16355; doi:10.1029/2001JB000372.
- Vavryčuk, V., and Kuehn, D. (2012). Moment tensor inversion of waveforms: a two-step time-frequency approach. *Geophys. J. Int.*, **190**, 1761-1776.
- Wang, R. (1997). Tidal response of the solid earth. In H. Wilhelm, W. Zürn and H. G. Wenzel (Eds.). *Tidal phenomena, Lecture Notes in Earth Sciences*, Vol. **66**, pp. 27-57, Springer-Verlag, Berlin/Heidelberg.
- Wang, R. (1999). A simple orthonormalization method for stable and efficient computation of Green's functions. *Bull. Seism. Soc. Am.*, **89**(3), 733-741.
- Wiggins, R. A., and Robinson, E. A. (1965). Recursive solution to the multichannel filtering problem. *J. Geophys. Res.*, **70**, 8, 2156-2202; doi: 10.1029/JZ070i008p01885.
- Whidden, K. M., Rudzinski, L., Lizurek, G., Pankow, K. L. (2013). Regional, local and in-mine moment tensors for the 2013 Rudna mine collapse, Poland. AGU 2013 fall meeting, 5210-2396
- Zahradník, J., and Custódio, S. (2012). Moment tensor resolvability: Application to Southwest Iberia. *Bull. Seism. Soc. Am.*, **102**, 1235-1254.
- Zhang, Y., Wang, R., Zschau, J., Chen, Y.-T., Parolai, S., Dahm, T. (2014). Automatic imaging of earthquake rupture processes by iterative deconvolution and stacking of high-rate GPS and strong motion. *J. Geophys. Res.*, doi: 10.1002/2013JB010469

| Name        | <b>pdemagcomp.f</b>                                                                                                                    |
|-------------|----------------------------------------------------------------------------------------------------------------------------------------|
| Contributor | James Dewey, U. S. Geological Survey/NEIC, MS 966, Denver, CO 80225-0046; E-mail: <a href="mailto:jdewey@usgs.gov">jdewey@usgs.gov</a> |
| Versions    | November 2011; DOI: 10.2312/GFZ.NMSOP-2_PD_3.1                                                                                         |

The following FORTRAN program, *pdemagcomp.f*, is based on a code for computing *mb* magnitude in the PDE bulletin of the USGS/NEIC. The original author is unknown. The program incorporates a procedure, *mbmag*, that has been used for decades to compute *mb* magnitude for the USGS/NEIC. It is included to provide the source code for *mbmag* and to show how *mbmag* is used. Program *pdemagcomp.f* was written by James W. Dewey, USGS/NEIC, for a limited research objective.

### **pdemagcomp.f.**

- c program pdemagcomp.f
- c calculates a Gutenberg-Richter *mb*, given epicentral distance,
- c event focal depth, amplitude in nanometers and period in seconds
- c requires 'qfile', the table of Gutenberg-Richter Q values

```
common q(108,17)
character*1 mbflg
real mb
```

```
open(unit = 23, status = 'old', file = 'qfile')
read (23,32,end = 33) ((q(i,j),j=1,17),i=1,108)
32 format(17(x,f4.2))
33 continue
```

```
type *, 'enter event focal depth, f5.1'
read *, depth
type *, 'enter station epicentral distance, degrees, f6.2'
read *, delta
type *, 'enter amplitude in nanometers, f7.1'
read *, amp
type *, 'enter period in seconds, f3,1'
read *, t
type *, 'enter P-wave travel-time residual in seconds, f5.1'
read *, res
call mbmag(t,amp,mbflg,mb,depth,delta,res)
type *, 'mb = ', mb, 'mbflag = ', mbflg
stop
end
```

```
subroutine mbmag(t,amp,mbflg,mb,depth,delta,res)
```

- c explanation of parameters (J. Dewey, 3/31/99)
- c t = P-wave period
- c amp = amplitude in nanometers
- c mbflg = 'n' -- computed magnitude not reported for this station
- c = 'R' or 'r' -- computed magnitude reported for this station,
- c but not used to compute average event magnitude
- c mb = short-period P-wave magnitude
- c depth = focal depth in km
- c delta = epicentral distance in degrees

```

c  q = magnitude calibration function for Gutenberg-Richter mb.  tabulated
c    from 2 to 109 degrees, at depths of 0,25,50,75,100,150,...,650,700 km.
c  res = travel-time residual, in seconds, of P-wave with which
c    this magnitude is associated

      common q(108,17)
      character*1 mbflg
      real mb

c
      mb=0.
      if(t.lt..1.or.t.gt.9.) go to 11
      if(amp.eq.0.) go to 11
      if(delta.lt.5..or.delta.gt.109.) go to 11
c    if(delta.lt.5..and.depth.ne.0.) go to 11
      if(abs(res).gt.10..or.depth.gt.700.) go to 11
c
      do 7 k=1,17
      if(k.gt.5) go to 6
      dep=(k-1)*25
      if(depth.ge.dep) go to 7
      s1=(depth+25.0-dep)*0.04
      go to 8
6    dep=(k-3)*50
      if(depth.gt.dep) go to 7
      s1=(depth+50.0-dep)*0.02
      go to 8
7    continue
      k=17
      s1=0.0
8    do 9 j=1,108
      del=j+1
      if(delta.ge.del) go to 9
      s2=delta+1.0-del
      go to 10
9    continue
      j=108
      s2=0.0
10   q1=q(j-1,k-1)+s1*(q(j-1,k)-q(j-1,k-1))
      q2=q(j,k-1) +s1*(q(j,k) -q(j,k-1))
      qval=q1+s2*(q2-q1)
      mb=alog10(amp*.001/t)+qval !change amp to microns for use with q tables
c    if((t.gt.3..or.(delta.lt.15..and.depth.lt.100.)).and.
c    1 mbflg.ne.'R') mbflg='r'
      if((t.gt.3..or.delta.lt.15.).and.mbflg.ne.'R') mbflg='r'
      return
c
      11 mbflg='n'
      return
      end
=

```

The following Q-table is used at the USGS/NEIC to compute the short-period body-wave magnitude *mb* that is published in the PDE bulletin of the USGS/NEIC. For distances from 5° to 109°, it represents a "digitized" version of Figure 5 of B. Gutenberg and C.F. Richter, 1956, "Magnitude and Energy of Earthquakes", *Annali di Geofisica*, v. IX, n. 1, p. 1-15. The original figure is reproduced as Figure VIII-6 in Richter, C.F., 1958, *Elementary Seismology*, Freeman, San Francisco, 768 p. The origin of the Q values at distance less than 5° is not documented. The USGS/NEIC does not compute *mb* for all distance ranges represented in the table.

| dist | h0  | h25 | h50 | h75 | h100 | h150 | h200 | h250 | h300 | h350 | h400 | h450 | h500 | h550 | h600 | h650 | h700 |
|------|-----|-----|-----|-----|------|------|------|------|------|------|------|------|------|------|------|------|------|
| 2    | 5.8 | 0   | 0   | 0   | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    |
| 3    | 5.8 | 0   | 0   | 0   | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    |
| 4    | 6.1 | 0   | 0   | 0   | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    | 0    |
| 5    | 6.4 | 6.3 | 6   | 6   | 6    | 6    | 6.9  | 6.7  | 6.8  | 6.7  | 6.7  | 6.8  | 6.8  | 6.8  | 6.7  | 6.7  | 6.7  |
| 6    | 6.5 | 6.5 | 6.2 | 6.1 | 6    | 6    | 6    | 6.7  | 6.7  | 6.7  | 6.8  | 6.9  | 6.8  | 6.8  | 6.8  | 6.7  | 6.7  |
| 7    | 7   | 6.8 | 6.5 | 6.4 | 6.3  | 6.2  | 6    | 6.8  | 6.7  | 6.8  | 6.9  | 6.9  | 6.9  | 6.9  | 6.8  | 6.7  | 6.7  |
| 8    | 7   | 7   | 6.6 | 6.5 | 6.4  | 6.4  | 6    | 6.8  | 6.8  | 6.9  | 6    | 6.1  | 6    | 6.9  | 6.8  | 6.8  | 6.7  |
| 9    | 7.2 | 7   | 6.8 | 6.6 | 6.6  | 6.4  | 6.1  | 6.8  | 6.8  | 6    | 6.1  | 6.1  | 6.1  | 6    | 6.9  | 6.8  | 6.7  |
| 10   | 7.3 | 7.1 | 6.9 | 6.8 | 6.7  | 6.4  | 6.1  | 6.9  | 6.9  | 6.1  | 6.2  | 6.2  | 6.2  | 6.1  | 6.8  | 6.8  | 6.7  |
| 11   | 7.2 | 7   | 6.9 | 6.7 | 6.6  | 6.4  | 6.2  | 6.9  | 6    | 6.2  | 6.2  | 6.2  | 6.2  | 6.1  | 6.9  | 6.8  | 6.8  |
| 12   | 7.1 | 7   | 6.8 | 6.7 | 6.5  | 6.3  | 6.2  | 6    | 6    | 6.2  | 6.2  | 6.2  | 6.2  | 6.2  | 6    | 6.8  | 6.8  |
| 13   | 7   | 6.9 | 6.7 | 6.5 | 6.4  | 6.2  | 6.2  | 6    | 6.1  | 6.2  | 6.2  | 6.2  | 6.3  | 6.2  | 6    | 6.9  | 6.8  |
| 14   | 6.6 | 6.5 | 6.5 | 6.1 | 6    | 6.1  | 6.2  | 6.1  | 6.1  | 6.2  | 6.2  | 6.3  | 6.3  | 6.2  | 6.1  | 6.9  | 6.8  |
| 15   | 6.3 | 6.1 | 6   | 6   | 6    | 6.1  | 6.2  | 6.2  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.3  | 6.1  | 6.9  | 6.8  |
| 16   | 5.9 | 5.9 | 5.9 | 5.9 | 6    | 6.1  | 6.2  | 6.2  | 6.2  | 6.1  | 6.2  | 6.3  | 6.4  | 6.3  | 6.2  | 6    | 6.9  |
| 17   | 5.9 | 5.9 | 5.9 | 6   | 6    | 6.1  | 6.2  | 6.2  | 6.2  | 6.1  | 6.1  | 6.3  | 6.4  | 6.4  | 6.2  | 6.1  | 6.9  |
| 18   | 5.9 | 5.9 | 5.9 | 6   | 6    | 6.1  | 6.2  | 6.3  | 6.2  | 6.1  | 6.1  | 6.3  | 6.4  | 6.4  | 6.3  | 6.1  | 6    |
| 19   | 6   | 6   | 6   | 6   | 6.1  | 6.1  | 6.3  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.3  | 6.1  | 6    |
| 20   | 6.1 | 6.1 | 6.1 | 6.1 | 6.1  | 6.2  | 6.3  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.2  | 6    |
| 21   | 6.1 | 6.2 | 6.1 | 6.1 | 6.1  | 6.2  | 6.3  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.2  | 6    |
| 22   | 6.2 | 6.2 | 6.2 | 6.2 | 6.1  | 6.2  | 6.3  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  | 6.1  |
| 23   | 6.3 | 6.3 | 6.2 | 6.2 | 6.1  | 6.2  | 6.4  | 6.3  | 6.2  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  | 6.1  |
| 24   | 6.4 | 6.3 | 6.3 | 6.2 | 6.2  | 6.3  | 6.4  | 6.3  | 6.2  | 6.1  | 6.2  | 6.3  | 6.3  | 6.4  | 6.4  | 6.4  | 6.1  |
| 25   | 6.5 | 6.4 | 6.3 | 6.2 | 6.2  | 6.3  | 6.4  | 6.3  | 6.2  | 6.1  | 6.2  | 6.3  | 6.3  | 6.4  | 6.4  | 6.4  | 6.2  |
| 26   | 6.5 | 6.4 | 6.3 | 6.3 | 6.3  | 6.4  | 6.5  | 6.4  | 6.2  | 6.1  | 6.2  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.2  |
| 27   | 6.5 | 6.4 | 6.4 | 6.3 | 6.3  | 6.4  | 6.5  | 6.4  | 6.2  | 6.1  | 6.2  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  |
| 28   | 6.6 | 6.5 | 6.4 | 6.4 | 6.4  | 6.5  | 6.5  | 6.4  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  |
| 29   | 6.6 | 6.5 | 6.4 | 6.4 | 6.4  | 6.5  | 6.5  | 6.4  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  |
| 30   | 6.6 | 6.6 | 6.5 | 6.5 | 6.5  | 6.5  | 6.5  | 6.4  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  |
| 31   | 6.7 | 6.6 | 6.5 | 6.5 | 6.5  | 6.5  | 6.5  | 6.4  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  |
| 32   | 6.7 | 6.7 | 6.6 | 6.6 | 6.5  | 6.6  | 6.4  | 6.4  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.4  |
| 33   | 6.7 | 6.7 | 6.6 | 6.6 | 6.6  | 6.5  | 6.4  | 6.4  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.4  |
| 34   | 6.7 | 6.7 | 6.7 | 6.7 | 6.6  | 6.5  | 6.4  | 6.4  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.4  | 6.3  |
| 35   | 6.6 | 6.7 | 6.7 | 6.7 | 6.7  | 6.5  | 6.4  | 6.3  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.3  | 6.3  |
| 36   | 6.6 | 6.7 | 6.7 | 6.7 | 6.7  | 6.5  | 6.4  | 6.3  | 6.3  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.3  | 6.3  |
| 37   | 6.5 | 6.6 | 6.7 | 6.7 | 6.7  | 6.5  | 6.4  | 6.3  | 6.2  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.4  | 6.3  | 6.3  |
| 38   | 6.5 | 6.6 | 6.7 | 6.7 | 6.7  | 6.5  | 6.4  | 6.3  | 6.2  | 6.1  | 6.1  | 6.2  | 6.3  | 6.4  | 6.3  | 6.3  | 6.3  |
| 39   | 6.4 | 6.5 | 6.6 | 6.7 | 6.6  | 6.5  | 6.4  | 6.3  | 6.1  | 6    | 6.1  | 6.2  | 6.3  | 6.4  | 6.3  | 6.3  | 6.3  |
| 40   | 6.4 | 6.5 | 6.6 | 6.7 | 6.6  | 6.5  | 6.3  | 6.2  | 6.1  | 6    | 6.1  | 6.2  | 6.3  | 6.4  | 6.3  | 6.2  | 6.3  |
| 41   | 6.5 | 6.5 | 6.5 | 6.6 | 6.6  | 6.4  | 6.3  | 6.2  | 6    | 6    | 6.1  | 6.2  | 6.3  | 6.3  | 6.3  | 6.2  | 6.3  |
| 42   | 6.5 | 6.5 | 6.5 | 6.6 | 6.6  | 6.4  | 6.3  | 6.2  | 6    | 6    | 6.1  | 6.2  | 6.3  | 6.3  | 6.3  | 6.2  | 6.3  |
| 43   | 6.5 | 6.5 | 6.5 | 6.6 | 6.6  | 6.4  | 6.3  | 6.1  | 6    | 6    | 6.1  | 6.2  | 6.3  | 6.3  | 6.3  | 6.2  | 6.3  |
| 44   | 6.6 | 6.6 | 6.5 | 6.6 | 6.6  | 6.4  | 6.3  | 6.1  | 6.1  | 6    | 6.1  | 6.2  | 6.3  | 6.3  | 6.3  | 6.2  | 6.2  |
| 45   | 6.7 | 6.7 | 6.6 | 6.6 | 6.6  | 6.4  | 6.2  | 6.1  | 6.1  | 6    | 6.1  | 6.2  | 6.3  | 6.3  | 6.3  | 6.2  | 6.2  |
| 46   | 6.8 | 6.7 | 6.7 | 6.7 | 6.6  | 6.4  | 6.2  | 6.1  | 6.1  | 6    | 6.1  | 6.2  | 6.3  | 6.3  | 6.3  | 6.2  | 6.2  |
| 47   | 6.9 | 6.8 | 6.7 | 6.7 | 6.6  | 6.4  | 6.2  | 6.1  | 6.1  | 6    | 6.1  | 6.2  | 6.3  | 6.3  | 6.3  | 6.2  | 6.2  |
| 48   | 6.9 | 6.8 | 6.8 | 6.7 | 6.6  | 6.5  | 6.2  | 6.1  | 6.1  | 6    | 6.1  | 6.2  | 6.2  | 6.3  | 6.3  | 6.2  | 6.2  |
| 49   | 6.8 | 6.8 | 6.8 | 6.8 | 6.7  | 6.5  | 6.2  | 6.2  | 6.1  | 6.1  | 6.1  | 6.2  | 6.2  | 6.3  | 6.3  | 6.2  | 6.2  |
| 50   | 6.7 | 6.8 | 6.8 | 6.8 | 6.8  | 6.5  | 6.3  | 6.2  | 6.1  | 6.1  | 6.1  | 6.1  | 6.2  | 6.3  | 6.3  | 6.1  | 6.1  |

| dist | h0  | h25 | h50 | h75 | h100 | h150 | h200 | h250 | h300 | h350 | h400 | h450 | h500 | h550 | h600 | h650 | h700 |
|------|-----|-----|-----|-----|------|------|------|------|------|------|------|------|------|------|------|------|------|
| 51   | 6.7 | 6.7 | 6.8 | 6.8 | 6.8  | 6.5  | 6.3  | 6.2  | 6.2  | 6.1  | 6.1  | 6.1  | 6.2  | 6.2  | 6.2  | 6.1  | 6.1  |
| 52   | 6.7 | 6.7 | 6.8 | 6.8 | 6.8  | 6.5  | 6.4  | 6.2  | 6.2  | 6.1  | 6.1  | 6.1  | 6.1  | 6.2  | 6.2  | 6.1  | 6.1  |
| 53   | 6.7 | 6.7 | 6.8 | 6.8 | 6.8  | 6.6  | 6.4  | 6.2  | 6.2  | 6.1  | 6.1  | 6.1  | 6.1  | 6.1  | 6.2  | 6.1  | 6.1  |
| 54   | 6.8 | 6.8 | 6.8 | 6.8 | 6.8  | 6.6  | 6.4  | 6.3  | 6.2  | 6.1  | 6.1  | 6.1  | 6.1  | 6.1  | 6.1  | 6.1  | 6    |
| 55   | 6.8 | 6.8 | 6.8 | 6.8 | 6.8  | 6.6  | 6.5  | 6.3  | 6.2  | 6.2  | 6.1  | 6.1  | 6.1  | 6.1  | 6.1  | 6    | 6    |
| 56   | 6.8 | 6.8 | 6.8 | 6.8 | 6.8  | 6.7  | 6.5  | 6.3  | 6.2  | 6.2  | 6.1  | 6.1  | 6.1  | 6.1  | 6.1  | 6    | 6    |
| 57   | 6.8 | 6.8 | 6.8 | 6.9 | 6.8  | 6.7  | 6.5  | 6.4  | 6.2  | 6.2  | 6.2  | 6.2  | 6.1  | 6.1  | 6    | 6    | 6    |
| 58   | 6.8 | 6.8 | 6.9 | 6.9 | 6.8  | 6.7  | 6.5  | 6.4  | 6.3  | 6.2  | 6.2  | 6.2  | 6.1  | 6.1  | 6    | 6    | 6    |
| 59   | 6.9 | 6.9 | 6.9 | 6.9 | 6.9  | 6.7  | 6.5  | 6.4  | 6.3  | 6.2  | 6.2  | 6.2  | 6.2  | 6.1  | 6    | 6    | 6    |
| 60   | 6.9 | 6.9 | 6.9 | 6.9 | 6.9  | 6.7  | 6.5  | 6.4  | 6.3  | 6.3  | 6.2  | 6.2  | 6.2  | 6.1  | 6    | 6    | 6    |
| 61   | 6.9 | 6.9 | 6.9 | 6.9 | 6.8  | 6.7  | 6.5  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.2  | 6.2  | 6.1  | 6    | 6    |
| 62   | 7   | 6.9 | 6.9 | 6.9 | 6.8  | 6.7  | 6.6  | 6.4  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.2  | 6.1  | 6.1  | 6    |
| 63   | 7   | 6.9 | 6.9 | 6.8 | 6.7  | 6.7  | 6.6  | 6.5  | 6.4  | 6.4  | 6.4  | 6.3  | 6.3  | 6.2  | 6.2  | 6.1  | 6    |
| 64   | 7   | 6.9 | 6.8 | 6.7 | 6.7  | 6.7  | 6.6  | 6.5  | 6.5  | 6.4  | 6.4  | 6.4  | 6.4  | 6.3  | 6.2  | 6.1  | 6.1  |
| 65   | 7   | 6.9 | 6.8 | 6.7 | 6.7  | 6.7  | 6.6  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.4  | 6.3  | 6.2  | 6.1  | 6.1  |
| 66   | 7   | 6.9 | 6.8 | 6.7 | 6.7  | 6.7  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.2  | 6.2  | 6.1  |
| 67   | 7   | 6.9 | 6.8 | 6.7 | 6.7  | 6.8  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.3  | 6.2  | 6.1  |
| 68   | 7   | 6.9 | 6.8 | 6.7 | 6.7  | 6.8  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.3  | 6.2  | 6.2  |
| 69   | 7   | 6.9 | 6.7 | 6.7 | 6.6  | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.4  | 6.3  | 6.3  | 6.2  | 6.2  |
| 70   | 6.9 | 6.9 | 6.7 | 6.7 | 6.6  | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.3  | 6.3  | 6.2  | 6.2  |
| 71   | 6.9 | 6.9 | 6.7 | 6.7 | 6.6  | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.2  |
| 72   | 6.9 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.2  |
| 73   | 6.9 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.3  |
| 74   | 6.8 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.3  |
| 75   | 6.8 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.3  | 6.2  | 6.3  | 6.3  | 6.3  |
| 76   | 6.9 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  | 6.3  | 6.2  | 6.3  | 6.3  | 6.3  |
| 77   | 6.9 | 6.8 | 6.8 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.6  | 6.5  | 6.4  | 6.2  | 6.2  | 6.2  | 6.3  | 6.3  |
| 78   | 6.9 | 6.8 | 6.8 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.6  | 6.5  | 6.4  | 6.2  | 6.2  | 6.2  | 6.3  | 6.3  |
| 79   | 6.8 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.6  | 6.6  | 6.5  | 6.4  | 6.2  | 6.2  | 6.2  | 6.3  | 6.3  |
| 80   | 6.7 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.6  | 6.6  | 6.5  | 6.4  | 6.2  | 6.2  | 6.2  | 6.3  | 6.3  |
| 81   | 6.8 | 6.8 | 6.7 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.6  | 6.6  | 6.5  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.3  |
| 82   | 6.9 | 6.8 | 6.8 | 6.7 | 6.6  | 6.5  | 6.5  | 6.5  | 6.6  | 6.6  | 6.5  | 6.4  | 6.3  | 6.3  | 6.3  | 6.3  | 6.3  |
| 83   | 7   | 6.9 | 6.8 | 6.7 | 6.7  | 6.6  | 6.5  | 6.5  | 6.6  | 6.6  | 6.5  | 6.5  | 6.3  | 6.3  | 6.3  | 6.4  | 6.3  |
| 84   | 7   | 7   | 6.8 | 6.8 | 6.7  | 6.6  | 6.5  | 6.6  | 6.6  | 6.6  | 6.5  | 6.5  | 6.4  | 6.4  | 6.4  | 6.4  | 6.3  |
| 85   | 7   | 7   | 6.9 | 6.8 | 6.7  | 6.6  | 6.5  | 6.6  | 6.6  | 6.6  | 6.6  | 6.5  | 6.4  | 6.4  | 6.4  | 6.4  | 6.4  |
| 86   | 6.9 | 7   | 7   | 6.8 | 6.8  | 6.6  | 6.6  | 6.6  | 6.6  | 6.7  | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  |
| 87   | 7   | 7   | 7   | 6.9 | 6.8  | 6.7  | 6.6  | 6.6  | 6.7  | 6.7  | 6.6  | 6.5  | 6.5  | 6.5  | 6.5  | 6.5  | 6.4  |
| 88   | 7.1 | 7.1 | 7   | 6.9 | 6.8  | 6.8  | 6.6  | 6.6  | 6.7  | 6.7  | 6.6  | 6.6  | 6.6  | 6.6  | 6.6  | 6.5  | 6.4  |
| 89   | 7   | 7.1 | 7.1 | 7   | 6.9  | 6.8  | 6.7  | 6.7  | 6.7  | 6.7  | 6.6  | 6.6  | 6.6  | 6.7  | 6.7  | 6.6  | 6.5  |
| 90   | 7   | 7   | 7.1 | 7   | 6.9  | 6.8  | 6.7  | 6.7  | 6.7  | 6.7  | 6.6  | 6.7  | 6.7  | 6.7  | 6.7  | 6.7  | 6.5  |
| 91   | 7.1 | 7.1 | 7.2 | 7.1 | 7    | 6.9  | 6.8  | 6.7  | 6.7  | 6.7  | 6.7  | 6.7  | 6.7  | 6.8  | 6.8  | 6.7  | 6.6  |
| 92   | 7.1 | 7.2 | 7.2 | 7.2 | 7.1  | 6.9  | 6.8  | 6.8  | 6.7  | 6.8  | 6.7  | 6.8  | 6.8  | 6.8  | 6.8  | 6.8  | 6.7  |
| 93   | 7.2 | 7.2 | 7.2 | 7.2 | 7.1  | 7    | 6.9  | 6.8  | 6.8  | 6.8  | 6.8  | 6.8  | 6.8  | 6.9  | 6.8  | 6.9  | 6.7  |
| 94   | 7.1 | 7.2 | 7.2 | 7.2 | 7.2  | 7    | 6.9  | 6.9  | 6.9  | 6.9  | 6.9  | 6.9  | 6.9  | 6.9  | 7    | 6.9  | 6.8  |
| 95   | 7.2 | 7.2 | 7.2 | 7.2 | 7.2  | 7.1  | 7    | 7    | 6.9  | 6.9  | 6.9  | 6.9  | 6.9  | 7    | 7    | 7    | 6.9  |
| 96   | 7.3 | 7.2 | 7.3 | 7.3 | 7.3  | 7.2  | 7.1  | 7    | 7    | 7    | 6.9  | 7    | 7    | 7    | 7    | 7    | 6.9  |
| 97   | 7.4 | 7.3 | 7.3 | 7.3 | 7.3  | 7.2  | 7.1  | 7.1  | 7    | 7    | 7    | 7    | 7.1  | 7.1  | 7.1  | 7    | 7    |
| 98   | 7.5 | 7.3 | 7.3 | 7.3 | 7.3  | 7.3  | 7.2  | 7.1  | 7.1  | 7.1  | 7.1  | 7.1  | 7.1  | 7.1  | 7.1  | 7.1  | 7    |
| 99   | 7.5 | 7.3 | 7.3 | 7.3 | 7.4  | 7.3  | 7.2  | 7.2  | 7.2  | 7.1  | 7.1  | 7.2  | 7.2  | 7.2  | 7.2  | 7.1  | 7    |
| 100  | 7.3 | 7.3 | 7.3 | 7.4 | 7.4  | 7.3  | 7.2  | 7.2  | 7.2  | 7.2  | 7.2  | 7.2  | 7.2  | 7.2  | 7.2  | 7.2  | 7.1  |

| dist | h0  | h25 | h50 | h75 | h100 | h150 | h200 | h250 | h300 | h350 | h400 | h450 | h500 | h550 | h600 | h650 | h700 |
|------|-----|-----|-----|-----|------|------|------|------|------|------|------|------|------|------|------|------|------|
| 101  | 7.4 | 7.3 | 7.4 | 7.4 | 7.4  | 7.4  | 7.3  | 7.3  | 7.3  | 7.3  | 7.3  | 7.3  | 7.3  | 7.3  | 7.3  | 7.2  | 7.1  |
| 102  | 7.4 | 7.4 | 7.4 | 7.5 | 7.5  | 7.5  | 7.4  | 7.3  | 7.3  | 7.3  | 7.4  | 7.4  | 7.4  | 7.4  | 7.3  | 7.3  | 7.2  |
| 103  | 7.5 | 7.5 | 7.5 | 7.5 | 7.6  | 7.6  | 7.5  | 7.4  | 7.4  | 7.4  | 7.5  | 7.5  | 7.5  | 7.5  | 7.4  | 7.3  | 7.2  |
| 104  | 7.6 | 7.6 | 7.6 | 7.7 | 7.7  | 7.7  | 7.7  | 7.6  | 7.6  | 7.6  | 7.6  | 7.6  | 7.6  | 7.6  | 7.5  | 7.4  | 7.3  |
| 105  | 7.7 | 7.7 | 7.7 | 7.8 | 7.8  | 7.8  | 7.8  | 7.8  | 7.8  | 7.8  | 7.7  | 7.7  | 7.7  | 7.6  | 7.6  | 7.5  | 7.4  |
| 106  | 7.8 | 7.8 | 7.8 | 7.8 | 7.8  | 7.9  | 7.9  | 7.9  | 7.9  | 7.9  | 7.8  | 7.8  | 7.7  | 7.7  | 7.6  | 7.5  | 7.4  |
| 107  | 7.8 | 7.8 | 7.9 | 7.9 | 7.9  | 7.9  | 7.9  | 8    | 8    | 8    | 7.9  | 7.8  | 7.7  | 7.7  | 7.6  | 7.5  | 7.4  |
| 108  | 7.9 | 7.9 | 7.9 | 7.9 | 7.9  | 8    | 8    | 8    | 8    | 8    | 8    | 7.8  | 7.8  | 7.7  | 7.7  | 7.6  | 7.5  |
| 109  | 8   | 8   | 8   | 8   | 8    | 8    | 8    | 8    | 8    | 8    | 8    | 7.9  | 7.8  | 7.7  | 7.7  | 7.6  | 7.5  |

# Chapter 4

## Seismic Signals and Noise

(Version June 2013. DOI: 10.2312/GFZ.NMSOP-2\_ch4)

Peter Bormann<sup>1)</sup> and Erhard Wielandt<sup>2)</sup>

<sup>1)</sup> Formerly GFZ German Research Centre for Geosciences, Department 2: Physics of the Earth, Telegrafenberg, 14473 Potsdam, Germany; Now: E-mail: [pb65@gmx.net](mailto:pb65@gmx.net)

<sup>2)</sup> Formerly Institute of Geophysics, University of Stuttgart; Now: Cranachweg 14/1, D-73230 Kirchheim unter Teck, Germany; E-mail: [e.wielandt@t-online.de](mailto:e.wielandt@t-online.de)

|                                                                                                  | <b>page</b> |
|--------------------------------------------------------------------------------------------------|-------------|
| <b>4.1 Introduction (P. Bormann and E. Wielandt)</b>                                             | <b>2</b>    |
| 4.1.1 About this chapter                                                                         | 2           |
| 4.1.2 Nature and mathematical representation of signals and noise                                | 3           |
| <b>4.2 Different mathematical formulations of spectral analysis (E. Wielandt and P. Bormann)</b> | <b>4</b>    |
| 4.2.1 Fourier transformation of continuous transient signals                                     | 4           |
| 4.2.2 Fourier transformation of sampled signals                                                  | 6           |
| 4.2.3 Spectral analysis of stationary signals (noise)                                            | 6           |
| 4.2.3.1 Energy and power densities of continuous signals                                         | 7           |
| 4.2.3.2 Energy and power densities of sampled signals                                            | 8           |
| 4.2.3.3 Obtaining the ESD or PSD by spectral analysis of the autocorrelation                     | 9           |
| 4.2.3.4 Testing PSD and ESD computation                                                          | 9           |
| 4.2.3.5 Changing between displacement, velocity, and acceleration                                | 9           |
| 4.2.3.6 Representing PSDs in decibels                                                            | 10          |
| 4.2.4 Cross-correlation and coherence                                                            | 10          |
| 4.2.5 The uncertainty principle of the Fourier Transformation                                    | 11          |
| <b>4.3 A first look at seismic and instrumental noise (P. Bormann and E. Wielandt)</b>           | <b>11</b>   |
| 4.3.1 Global models for the PSD of seismic noise                                                 | 11          |
| 4.3.2 Instrumental self-noise                                                                    | 14          |
| <b>4.4 Comparing spectra of transient and stationary signals (E. Wielandt and P. Bormann)</b>    | <b>16</b>   |
| 4.4.1 Bandwidth and amplitude                                                                    | 16          |
| 4.4.2 The dynamic range of a seismograph                                                         | 17          |
| 4.4.3 Representing clip levels, signals, and noise in the same diagram                           | 18          |
| 4.4.4 Approximate conversion of power densities into recording amplitudes                        | 19          |
| <b>4.5 Seismograph response and waveform (P. Bormann)</b>                                        | <b>21</b>   |
| 4.5.1 Empirical case studies                                                                     | 21          |
| 4.5.2 Theoretical considerations on signal distortion in seismic records                         | 25          |
| <b>4.6 Characteristics and sources of seismic noise (P. Bormann)</b>                             | <b>30</b>   |
| 4.6.1 Ocean microseisms                                                                          | 30          |
| 4.6.2 Noise at the seabottom                                                                     | 33          |
| 4.6.2.1 Broadband noise                                                                          | 33          |
| 4.6.2.2 Short-period noise (0.5-50 Hz)                                                           | 35          |



|            |                                                                            |           |
|------------|----------------------------------------------------------------------------|-----------|
| 4.6.3      | Seismic noise on land                                                      | 36        |
| 4.6.3.1    | Long-period noise                                                          | 36        |
| 4.6.3.2    | Short-period noise (0.5 to 50 Hz)                                          | 38        |
| 4.6.3.3    | Seismic noise in urban environments                                        | 44        |
| 4.6.4      | Signal and SNR variations due to local site conditions                     | 45        |
| 4.6.5      | Installations in subsurface mines, tunnels and boreholes                   | 48        |
| <b>4.7</b> | <b>Improving the Signal-to-Noise Ratio by data processing (P. Bormann)</b> | <b>50</b> |
| 4.7.1      | Frequency filtering                                                        | 50        |
| 4.7.2      | Velocity filtering and beamforming                                         | 52        |
| 4.7.3      | Noise prediction-error filtering                                           | 53        |
| 4.7.4      | Noise polarization filtering                                               | 53        |
| <b>4.8</b> | <b>Global detection thresholds (P. Bormann)</b>                            | <b>54</b> |
| 4.8.1      | Detection thresholds for surface waves and long-period body waves          | 54        |
| 4.8.2      | Detection thresholds for short-period body waves                           | 55        |
|            | Acknowledgments                                                            | 57        |
|            | Recommended overview readings                                              | 58        |
|            | References                                                                 | 58 - 62   |

## 4.1 Introduction

### 4.1.1 About this chapter

Seismic signals are usually transient waveforms radiated from a localized natural or man-made seismic source. They can be used to locate the source, to analyze source processes, and to study the structure of the medium of propagation. In contrast, the term “seismic noise” designates undesired components of ground motion that do not fit in our conceptual model of the signal under investigation. What we identify and treat as seismic noise depends on the available data, on the aim of our study and on the method of analysis. Accordingly, data treated as noise in one context may be considered as useful signals in other applications. For example, short-period seismic noise can be used for microzonation studies in urban areas (see Chapter 14), and long-period noise for surface-wave tomography (Yanovskaya, 2012).

Although seismic noise proper relates only to the first four bullet point below disturbing noise in seismic records may in a wider sense comprise the following:

- Ambient vibrations due to natural sources (like ocean microseisms, wind, etc);
- Man-made vibrations (from industry, traffic, etc);
- Secondary signals resulting from wave propagation in an inhomogeneous medium (scattering);
- Effects of gravity (like Newtonian attraction of atmosphere, horizontal accelerations due to surface tilt);
- Signals resulting from the sensitivity of seismometers to ambient conditions (like temperature, air pressure, magnetic field, etc);
- Signals due to technical imperfections or deterioration of the sensor (corrosion, leakage currents, defective semiconductors, etc);
- Intrinsic self-noise of the seismograph (like Brownian noise, electronic and quantization noise);
- Artifacts from data processing.

The study of earthquakes or the imaging of the Earth with seismic wave arrivals requires their detection above background noise in seismic records. Levels of natural ambient noise may vary by 60 dB (a factor of 1000 in amplitude) depending on location, season, time of day, and weather conditions (see Fig. 4.30). This corresponds to differences in detection thresholds for seismic arrivals by about three magnitude units. Therefore, prior investigation of noise levels at potential recording sites is of utmost importance.

In the following we will look into

- different mathematical representations of seismic noise and signals;
- origins of natural and man-made ambient seismic noise;
- basic features of signals and noise in the wide frequency range of interest, from local micro-earthquake to normal mode recordings ( $0.2 \text{ mHz} < f < 100 \text{ Hz}$ );
- influence of sensor-recorder-filter responses and of the installation environment on the appearance of signals and noise in seismic recordings;
- task-dependent procedures for SNR improvement of earthquake and explosion records;
- global detectability thresholds for short-period and long-period signals.

For an overview of procedures that emphasize on the causes and treatment of noise in seismic exploration we refer to Kumar and Ahmed (2011).

#### 4.1.2 Nature and mathematical representation of signals and noise

As illustrated in Fig. 1.1, one of the key problems in seismology is to solve the inverse problem, i.e., to derive from the analysis of seismic records information on the structure and physical properties of the Earth through which the seismic waves propagate (see Chapter 2), as well as on the geometry, kinematics and dynamics of the seismic source process (see Chapter 3). The study of source processes is complicated by the fact that the seismic signals are weakened and distorted by geometric spreading and attenuation and due to reflection, diffraction, mode conversion and interference during their travel through the Earth. They are also distorted by the seismograph. While the Earth acts as a low-pass filter by attenuating higher frequencies most effectively, a mechanical seismograph is a second-order high-pass filter with a roll-off of  $-12 \text{ dB}$  (a factor of 4) per octave for periods larger than its eigenperiod (see Chapter 5).

Additionally, seismograms contain noise and the desired signals are sometimes completely masked by it. Therefore, one of the main issues in applied seismology is to ensure a good signal-to-noise ratio (SNR) both by selecting recording sites with low ambient noise and by data processing. The success of the latter largely depends on our understanding of the ways in which seismic signals and noise differ. The basic mathematical tool for this purpose is the harmonic or Fourier analysis, that is, the decomposition of a signal into sinewaves. Depending on the type of signal (transient or stationary, continuous or sampled) different mathematical formulations must be used.

The Fourier integral transformation (section 4.2.1) can only be applied to transient signals (signals that disappear or decay after some time so that they have a finite energy). For other signals, the integral may diverge. The discrete transformation (section 4.2.2) requires a sam-

pled signal of finite duration. Signals whose (mean) amplitude does not vary much within the interval of time over which we typically analyze earthquake signals are mathematically treated as stationary signals, that is, signals whose statistical properties do not change with time. Such signals require a different mathematical treatment. Of course, this is a mathematical idealization whose appropriateness depends on the time scale. Over longer intervals of time, almost all signals will change their statistical properties. For example, marine microseisms can normally be treated as stationary within the time frame of a seismogram but often change their amplitude and predominant period within a few days according to meteorological conditions over the oceans.

A Fourier transformation in the strict sense does not exist for stationary signals. This is not a practical problem because we can Fourier analyze such signals over finite time intervals of any desired length. A practically more important difference is that we are normally not interested in the detailed waveform of such signals because we want to remove them from the record, not to interpret them. (There are, however, interesting exceptions as mentioned in the introduction.) In most cases we only want to know how strong they are in different parts of the frequency spectrum. This information is contained in a quantity named **Power Spectral Density** that is explained in section 4.2.3.

Transient signals such as earthquake seismograms may have a complicated waveform, but this waveform is determined by the earthquake source, the structure of the Earth, and the properties of the seismograph. The amplitudes and phases of the harmonic components of the signal are not random numbers but follow certain mathematical relationships. They are, in principle, smooth functions of frequency, often smoother and simpler than the signal appears in time domain. For example, the waveform of a train of surface waves is essentially determined by a much simpler dispersion curve. Such waveforms or spectra are sometimes called deterministic, in contrast to the “stochastic” or “random” waveform of seismic noise. Strictly speaking, however, seismic noise is also deterministic. The difference is that we don’t know its sources and propagation paths well enough to predict or interpret the waveform, don’t want to care about them, and therefore assume that the spectral amplitudes and phases are random numbers. Again, this is a mathematical simplification that may or may not be appropriate.

## 4.2 Different mathematical formulations of spectral analysis

### 4.2.1 Fourier transformation of continuous transient signals

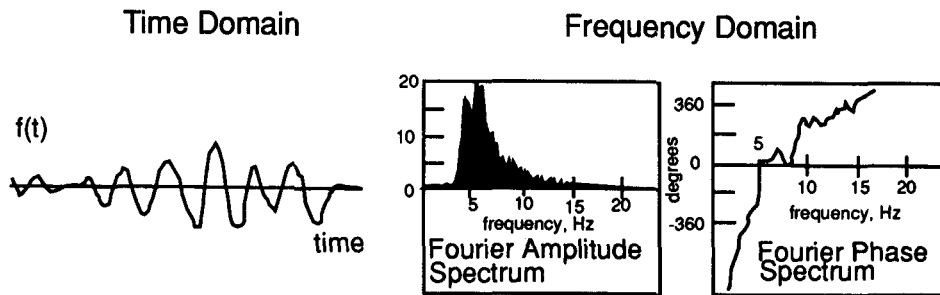
The signal radiated from a seismic source, be it an explosion or a shear rupture, is usually a more or less complicated displacement step function or velocity impulse of finite duration from milliseconds up to a few minutes at the most (see Chapter 2, Fig. 2.5 and Chapter 3, Figs. 3.4 and 3.9). According to the **Fourier theorem** any arbitrary **transient function**  $f(t)$  in the *time domain* can be represented by an equivalent function  $F(\omega)$  in the **frequency domain**. The mathematical relationship between the two domains is defined by the **Fourier integral transformation**:

$$(4.1) \quad f(t) = (2\pi)^{-1} \int_{-\infty}^{\infty} F(\omega) \exp(i\omega t) d\omega$$

$$(4.2) \quad F(\omega) = |F(\omega)| \exp(i\phi(\omega)) = \int_{-\infty}^{\infty} f(t) \exp(-i\omega t) dt$$

Other sign conventions may be used, e.g.  $\exp(-i\omega t)$  in Eq. (4.1) and  $\exp(i\omega t)$  in Eq. (4.2) in wave propagation studies in order to assure that the wave-number vector is positive in the direction of wave propagation. The normalization factor  $(2\pi)^{-1}$  may be applied in the backward or in the forward transformation, or its square root in both.  $f(t)$  needs not be a continuous function in a mathematical sense but it must be defined at every instant of time, with the possible exception of isolated points where a step may occur.

$F(\omega)$  is the **complex Fourier transform** of the continuous signal  $f(t)$ , usually written as a function of the angular frequency  $\omega$  which is  $2\pi$  times the common frequency (the number of cycles per second). The absolute value of the Fourier transform,  $|F(\omega)|$ , is called the **amplitude spectrum**, and its phase  $\phi(\omega)$  the **phase spectrum**. Fig. 4.1 gives an example. The phase may practically be measured in radians, cycles ( $2\pi$  radians), or degrees, although only radians are used in mathematical formulas. Since the integral in Eq. (4.1) is equivalent to a sum, the Fourier transformation implies that an arbitrary transient signal, even an impulsive one, can be represented as a sum of time-harmonic functions (sinewaves). Fig. 4.2 illustrates how a sum of harmonic terms can equal an arbitrary function.



**Fig. 4.1** A signal recorded as a function of time (left) can be represented equivalently in the frequency domain by its complex Fourier spectrum. The amplitude (middle) and phase spectrum (right) are both needed to calculate the complete time series (reproduced from Lay and Wallace, 1995, Figure 5.B1.1, p. 176; with permission of Elsevier Science, USA).

The concept of frequency filtering (for example, high-, low- or band-pass filtration) has its roots in analog signal processing and is therefore intimately connected to the Fourier Integral Transformation while practical procedures on a computer do not use integrals and are often only approximations to the corresponding analog procedures. Frequency filtering reduces the amplitudes of harmonic components of a signal in part of the frequency spectrum, usually in order to suppress noise. Examples will be given in paragraph 4.7.1 below. Filtering may also modify the phases. It will, in general, change the amplitude of the desired signal in time domain. Therefore, magnitudes of seismic events determined from amplitude and period readings of seismic phases are comparable only when determined from analog seismic records with identical standard frequency responses or digital records filtered accordingly (see IASPEI standards for magnitude measurement in IS 3.3). Systematic differences between various magnitude scales as well as saturation effects are the consequence of filtering the input signals with different responses in

different and often band-limited frequency ranges (see Chapter 3, Tab. 3.1 and Figs. 3.5, 3.20 and 3.47).

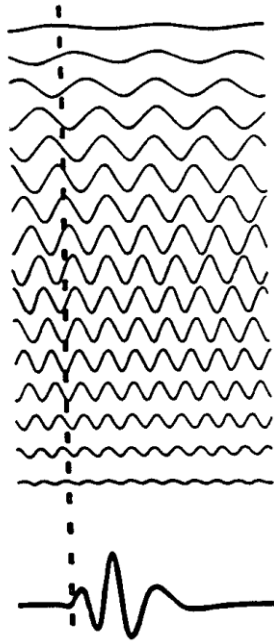
#### 4.2.2 Fourier transformation of sampled signals

A mathematically simpler and more computer-adequate formulation of the harmonic analysis is the *Discrete Fourier Transformation*:

$$(4.3) \quad a_k = \frac{1}{M} \sum_{l=-M/2}^{M/2-1} c_l e^{2\pi j k l / M}$$

$$(4.4) \quad c_l = \sum_{k=0}^{M-1} a_k e^{-2\pi j k l / M}$$

where  $a_k$  is a *time series* (a series of equidistant samples of a continuous signal  $f(t)$ ) and  $c_l$  are complex *Fourier coefficients*, each of which can again be decomposed into an amplitude and a phase factor. The index  $l$  represents the frequency. To obtain the frequency in physical units we replace  $k/M$  in the exponent by  $k\Delta t/M\Delta t$  where  $\Delta t$  is the sampling interval. Then  $k\Delta t$  is the time of sample  $a_k$  and  $l/M\Delta t$  is the frequency of the phase factor after  $c_l$ .  $T = M\Delta t$  is the length of the signal in time. Both series,  $a_k$  and  $c_l$ , are periodic outside their original interval of definition. So other conventions with respect to the index ranges of the summation are possible; all sums over  $M$  consecutive samples are equivalent. Supplementary information on different aspects of the Fourier transformation is found in Chapter 5, section 5.2.3 ff.



**Fig. 4.2** The transient signal is formed by summing up the harmonic terms of Eq. (4.3), a discretized version of Eq.(4.1). The amplitudes of each harmonic term vary, being prescribed by the amplitude spectrum  $|c_l|$ . The shift of the phase of each harmonic term is given by the phase spectrum (reproduced from Lay and Wallace, 1995, Figure 5.B1.2, p. 177; with permission of Elsevier Science, USA).

#### 4.2.3 Spectral analysis of stationary signals (noise)

As we have already said above, in the analysis of stationary noise we are normally not interested in the exact waveform or other details that do not convey useful information. In most cases, we only want to know “how strong” the noise is, or whether it will prevent us from detecting a specific earthquake signal. Since with frequency filters we may be able to sup-

press noise in one part of the spectrum and still preserve signals in another part, it is necessary to consider the spectral distribution rather than the total strength of both signals. There lies, however, a mathematical difficulty in this seemingly simple task: the formal measures of “strength” for transient and stationary signals are incompatible and cannot easily be compared. Mathematically speaking, transient signals have a finite energy and zero power (in the average over all times) while stationary signals have an infinite energy but a finite power. The same holds for the spectral densities. For a quantification of stationary noise, the concept of **Power Spectral Density (PSD)** is therefore appropriate while transient signals are more adequately described by their **Energy Spectral Density (ESD)**. The relationship between the two densities is trivial; we must however go into some mathematical detail to explain how they are calculated and used. In some situations (such as for the comparison of earthquake signals with instrumental clip levels) neither the PSD nor the ESD are suitable measures, and we must use other criteria which will be the subject of section 4.4.

Generally, the spectral density  $P(\omega)$  of some quantity  $Q$  is a function of frequency whose integral over a given frequency band (such as the passband of a specific seismograph) gives the total amount of  $Q$  in that band. When  $Q$  is the power of a signal, the physical dimension of  $P$  is power per bandwidth (per Hertz). However, not every quantity that has the physical dimension X per Hertz is a spectral density. For example, the Fourier transform  $F(\omega)$  in eq. (4.2) has the dimension  $\mu\text{m}$  per Hz when the seismic signal  $f(t)$  is measured in  $\mu\text{m}$ , but its absolute value  $|F(\omega)|$  is not an “amplitude spectral density” even if this name has been used by some authors (including Aki and Richards in their well-known textbook, 1980). Integrating  $|F(\omega)|$  over a bandwidth does, in general, NOT yield the amplitude of  $f(t)$  in that bandwidth, as demonstrated in Fig. 4.6.

#### 4.2.3.1 Energy and power densities of continuous signals

We can now explain how the Spectral Densities of Energy and Power are actually determined for a given signal. Mathematically we must distinguish between transient and stationary signals, and between continuous and sampled signals. Practically, of course, we always analyze signals of finite length that are, mathematically speaking, transient, and for digital processing the signals must be sampled. Still, most concepts behind data processing (such as using a sine-wave to represent a single frequency) come from the realm of continuous signals, and we must therefore start from the Fourier Integral Transformation even if it is not a practical tool for digital processing.

The Fourier Integral Transformation as formulated in Eqs. (4.1) and (4.2) has the mathematical property (known as Rayleigh’s or Parseval’s theorem)

$$(4.5) \quad \int_{-\infty}^{\infty} f(t)^2 dt = \int_{-\infty}^{\infty} \frac{|F(\omega)|^2}{2\pi} d\omega = \int_{-\infty}^{\infty} |F(2\pi\nu)|^2 d\nu$$

On the left side, we have what in data processing is called the total energy of signal  $f(t)$  (this is not a physical energy but proportional to it). We have assumed that the energy is finite even if the signal may mathematically have an infinite length (for example, it may decay exponentially). The right side then represents the energy as an integral of  $E = |F(2\pi\nu)|^2$  over all fre-

quencies  $\nu$ . Thus, according to the above definition, the integrand  $E$  is the Energy Spectral Density of the signal.

Note that negative frequencies must be included in the integration. For real signals, the ESD is a symmetric function,  $E(-2\pi\nu) = E(2\pi\nu)$ , so we can as well double it and integrate only over positive angular frequencies. This convention is normally used in engineering and other practical applications; the ESD is then given as twice the “mathematical” value. The same applies to the PSD in the next paragraph. It is important to clarify which convention is being used.

Power is energy per time. Normally we are not interested in the instantaneous power but in the average power over at least one cycle of the signal (if it is approximately periodic). The concept of power is mainly applied to stationary signals such as sinewaves or seismic noise whose average power is constant or varies slowly. A Fourier transform in a strict sense is not defined for such signals, but we can Fourier analyze the signal over any finite time interval of length  $T$ . When both sides of Eq. (4.5) are divided by  $T$ , the left side represents the average power, and  $P = E/T = F(2\pi\nu)|^2/T$  on the right side is the Power Spectral Density for that interval.

#### 4.2.3.2 Energy and power densities of sampled signals

We now consider a signal represented by  $M$  samples taken at equidistant intervals  $\Delta t = T/M$  where  $T$  is the total duration of the signal. The power  $W$  of such a signal is defined as the average square of the samples:

$$(4.6) \quad W = M^{-1} \sum a_k^2$$

Again, this is not a physical power, but the latter is proportional to  $W$ . The square root of  $W$  is the *effective* or *root-mean-square (rms) amplitude*  $a_{rms}$  of the signal. Now it is a property of the discrete Fourier transformation Eqs. (4.3) and (4.4) that

$$(4.7) \quad \sum a_k^2 = M^{-1} \sum |c_l|^2$$

Like in the case of Eq. (4.5), the total power  $W = M^{-1} \sum a_k^2 = M^{-2} \sum |c_l|^2$  is now expressed as a sum of contributions from different frequencies. Only discrete frequencies  $l/T$  are used in the transformation, but if we go back to the domain of continuous signals and consider the Fourier coefficients  $c_l$  as samples of a continuous function, then each coefficient stands for a frequency interval of width  $1/T$  (this is sometimes called a “frequency bin”). The ratio of power over bandwidth at frequency  $l/T$  is therefore

$$(4.8) \quad P_l = M^{-2} T |c_l|^2 = M^{-1} \Delta t |c_l|^2$$

The series  $P_l$  is the discrete equivalent of the PSD function. It offers a practical way to compute the PSD for stationary signals. For a graphical representation, the  $P_l$  series is often smoothed and

decimated because the original coefficients are normally positive random numbers. The suitably smoothed PSD approaches, at each frequency, a definite limit for  $M \rightarrow \infty$  when the signal is stationary (this is essentially the definition of stationarity). For (supposedly) stationary signals like seismic background noise, calculating the PSD of a long but finite section is the only possible form of a spectral analysis. The PSD is not equivalent to a Fourier transform; the signal cannot be reconstructed from it with an inverse transformation because the phase information is lost. For finite sections of a stationary signal we can of course use the discrete Fourier transformation if we later need a backward transformation. The smoothed and decimated PSD is however normally more useful because it contains all essential information in a concentrated form, and eliminates arbitrary normalization factors. The discrete equivalent of the Energy Spectral Density is obtained as  $E_l = T \cdot P_l$ .

#### 4.2.3.3 Obtaining the ESD or PSD by spectral analysis of the autocorrelation

An alternative way of determining the spectral density of energy or power is to calculate the Fourier transformation of its autocorrelation. An autocorrelation can be defined for stationary signals that have no Fourier transform; the signal is then essentially defined via its autocorrelation, and needs not be explicitly known. The autocorrelation of a stationary signal  $f(t)$  is the expectation value of the product  $f(t) f(t + \tau)$ , often written as  $\langle f(t) f(t + \tau) \rangle$ . It is a function of the delay (or time lag)  $\tau$  and normally decays rapidly for large time lags. So one can truncate it without committing a large error and then Fourier transform it. The method can as well be applied to finite sections of a signal. The properly normalized Fourier transform is the PSD of the signal. For details, see Havskov and Alguacil, 2002. Truncating the autocorrelation also has the desirable effect of smoothing the PSD. Nevertheless the method is now obsolete as a practical tool; it dates back to the time when the Fast Fourier Transformation was not available, and relatively short truncated series were more convenient to transform.

#### 4.2.3.4 Testing PSD and ESD computation

Although the PSD and ESD of a given signal are in principle independent of the method of computation and of the normalization of the Fourier transform, there often remains an uncertainty whether a given formula or computer program delivers properly normalized values. The question is easy to settle. All that must be done is to analyze an arbitrary signal, calculate its power or energy in time domain, and check if the integral of the PSD or ESD over all (or over positive) frequencies reproduces it.

#### 4.2.3.5 Changing between displacement, velocity, and acceleration

By differentiating under the integral in 4.1, we see that the time-derivative  $\dot{f}(t)$  of a signal  $f(t)$  has the Fourier transform  $j\omega F(\omega)$ . So if the signal was originally measured as a displacement and we need the Fourier transform of the velocity, we obtain it by multiplying the original Fourier transform by  $j\omega = j \cdot 2\pi\nu$  where  $\nu$  is the common frequency. The Fourier transform of the acceleration is  $-\omega^2 F(\omega)$ . We can of course also go back from acceleration or velocity to displacement by dividing by the appropriate factor. The ESD and PSD are positive quadratic functions of the Fourier amplitudes, so differentiating the signal brings a factor  $\omega^2$ , integration a factor  $\omega^{-2}$ , etc. A PSD of velocity can thus easily be converted into that of displacement or acceleration. It has become a standard to measure seismic noise as a PSD of acceleration, that is,



in  $(\text{m/s}^2)^2$  per Hz or  $\text{m}^2\text{s}^{-3}$ . We note again that “power” in this context is not the physical power in watts but simply the mean square amplitude of ground acceleration.

#### 4.2.3.6 Representing PSDs in decibels

As in acoustics, the power ratio  $r = (a_2/a_1)^2$  between two signals of amplitude  $a_1$  and  $a_2$  is often expressed in Decibels (dB). This is a logarithmic measure that expresses a ratio as a difference of logarithms. The difference in dB is  $10 \log_{10}[(a_2/a_1)^2] = 20 \log_{10}[(a_2/a_1)]$ . When expressing the power spectral density  $P_a$  of acceleration relative to the metric unit  $1 (\text{m/s}^2)^2/\text{Hz}$ , we get

$$(4.9) \quad \mathbf{P_a[dB]} = 10 \log_{10} [P_a / 1 (\text{m/s}^2)^2/\text{Hz}]$$

For the rest of this paragraph, we must caution the reader that **numerical values of physically incommensurable quantities will be compared**. Equations (4.10) and (4.11) and the following two text lines do not imply that the decibel units on both sides of the equal sign mean the same. They only look the same because the incommensurable reference levels were omitted, not, however, in Figure 4.3, where the three different power units for  $P_a$ ,  $P_v$  and  $P_d$  have been given on the ordinate. This clarification notwithstanding one gets by substituting the period  $T = 1/f$  (in s) for the frequency  $f$  in (4.6) and (4.7) :

$$(4.10) \quad \mathbf{P_v[dB]} = \mathbf{P_a[dB]} + 20 \log (T/2\pi)$$

and

$$(4.11) \quad \mathbf{P_d[dB]} = \mathbf{P_a[dB]} + 40 \log_{10} (T/2\pi) = \mathbf{P_v[dB]} + 20 \log (T/2\pi)$$

Consequently, for the period  $T = 2\pi = 6.28$  s  $\mathbf{P_a} = \mathbf{P_v} = \mathbf{P_d}$  (only the numbers are equal!) Also,  $(\mathbf{P_d} - \mathbf{P_a}) = 2 \times (\mathbf{P_v} - \mathbf{P_a}) = \text{constant}$  for any given period, negative for  $T < 2\pi$  s and positive for  $T > 2\pi$  s (Fig. 4.3).

#### 4.2.4 Cross-correlation and coherence

The **cross-correlation** of two stationary signals  $f(t)$  and  $g(t)$  at the time lag  $\tau$  is defined as the expectation value (practically, mean value) of the product  $f(t) g(t + \tau)$ . Properly normalized, it is a measure of the similarity between the two signals and is called their **coherence** (consult a textbook on data processing for details). Like the Power Spectral Density, it can be calculated via the Fourier transformation and specified as a function of frequency, in which case the name **coherence spectrum** is more adequate (but not always used). Each signal is fully coherent with itself (the coherence equals 1); independent random signals are incoherent (their coherence equals zero) although strictly this is only true in the limit of an infinite duration and after removing the mean. More interesting is the case that signals are partially coherent. One can then conclude that with some (frequency-dependent) fraction of the power, a common signal is hidden in them, for example a seismic signal masked by instrumental noise of the sensors.

Seismic signals depend not only on time but also on the location where they are recorded. It is therefore meaningful to ask whether the signal recorded at one location is coherent with the same signal recorded at another location. This aspect of coherence is usually expressed as a “coherence

length” over which the recorded signals have a specified coherence. The coherence length is only a fraction of a wavelength for omnidirectional noise (noise propagating in all directions) but can be much longer when noise originates in a limited source area and propagates essentially in one direction (Fig. 4.22). In a plane wave, the coherence length is theoretically infinite.

## 4.2.5 The uncertainty principle of the Fourier Transformation

The time signal  $f(t)$  and its Fourier transform  $F(\omega)$  according to eq. (4.2) are related to each other through many rules that can be found in the textbooks (a few textbooks are named at the beginning of section 5.2). One such relationship, the uncertainty principle, is important for the following discussions. It states that an impulsive signal (one whose energy is concentrated in a short interval of time) must have a large bandwidth; a signal with a small bandwidth cannot be concentrated in time, or have a sharp onset. For any signal, the product of its width (uncertainty, variance) in time and its bandwidth (in terms of angular frequency) cannot be smaller than  $\frac{1}{2}$ . Only a Gaussian pulse (one with the envelope  $\exp(-\alpha t^2)$ ) gives the minimum value. Many figures in this chapter, starting from Fig. 4.6, are illustrations of the uncertainty principle. In quantum mechanics, the same mathematical relationship has the consequence that the position and momentum of a particle cannot be known at the same time; if the particle is sharply localized, then its momentum is uncertain and vice versa.

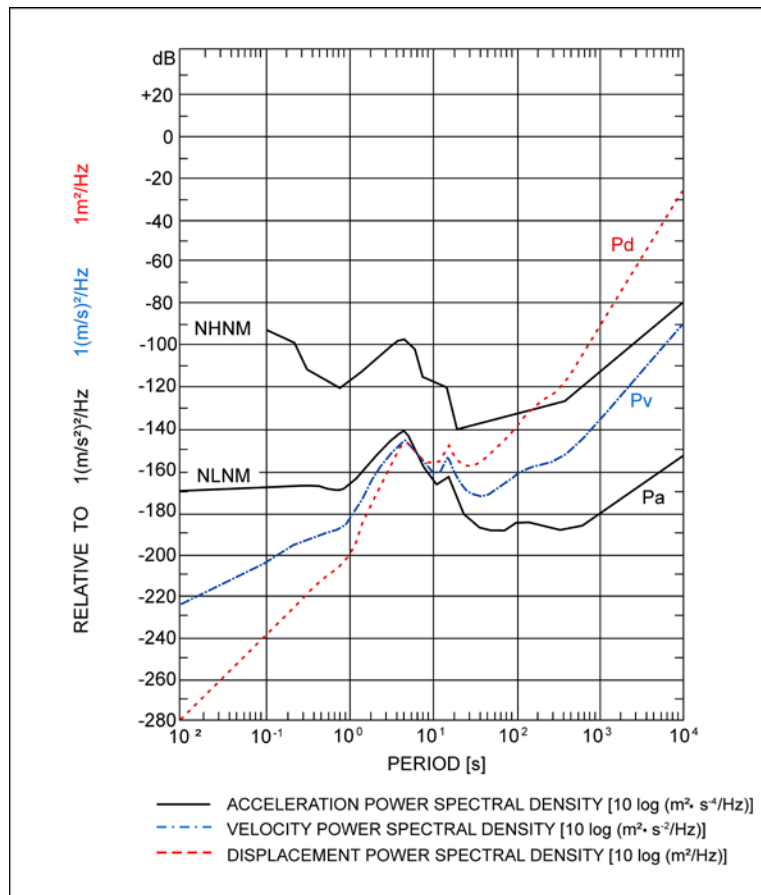
## 4.3 A first look at seismic and instrumental noise

### 4.3.1 Global models for the PSD of seismic noise

Fig. 4.3 is a summary of noise levels observed in a worldwide network of seismic stations (Peterson 1993). The graph shows the upper and lower envelopes (solid lines) of a large collection of curves representing the PSD of ground acceleration. The “engineering” convention is used, that is, the PSD has to be integrated over positive frequencies only. The envelopes are commonly referred to as the New High Noise Model (NHNM) and New Low Noise Model (NLNM). For periods which define the corners of the NLNM and NHNM polygons, Tables 4.1 and 4.2 give the displacement, velocity and acceleration power densities both in their respective kinematic units and in dB. A slightly different model was worked out by Berger et al. (2004) and is known as the GSN low-noise model. Fig. 4.3 also shows the lower envelopes for the PSDs of displacement and velocity. Displacements vary enormously between the short-period and the long-period end of the seismic spectrum. For the NLNM alone the variation is 260 dB (equivalent to 13 orders of magnitude in the signal amplitude) in the period range of seismological interest. The variation is reduced to about 130 dB for velocity and to 50 dB for acceleration. Understandably the PSD of acceleration is preferred for graphic representations. The noise peak of ocean microseisms around 5 s period (see 4.3.1) stands out most clearly here, as does the minimum near 0.04 Hz, termed “a window for earthquakes” by Savino et al. (1972). The height of the microseismic peak exhibits a strong seasonal variation. Another important minimum in the vertical-component seismic noise exists around 3 mHz which is important for free oscillation observations. The physical reasons for its existence have been discussed by Zürn and Wielandt (2007).

The ocean microseisms separate the two traditional domains of short-period ( $T < 2\text{--}3$  s) and long-period seismology ( $T > 10$  s). Broadband recordings spanning over this peak were rare after WWII, especially in the western world after the introduction of the US World-Wide Standard Seismograph Network (WWSSN). But even nowadays, with digital broadband-velocity

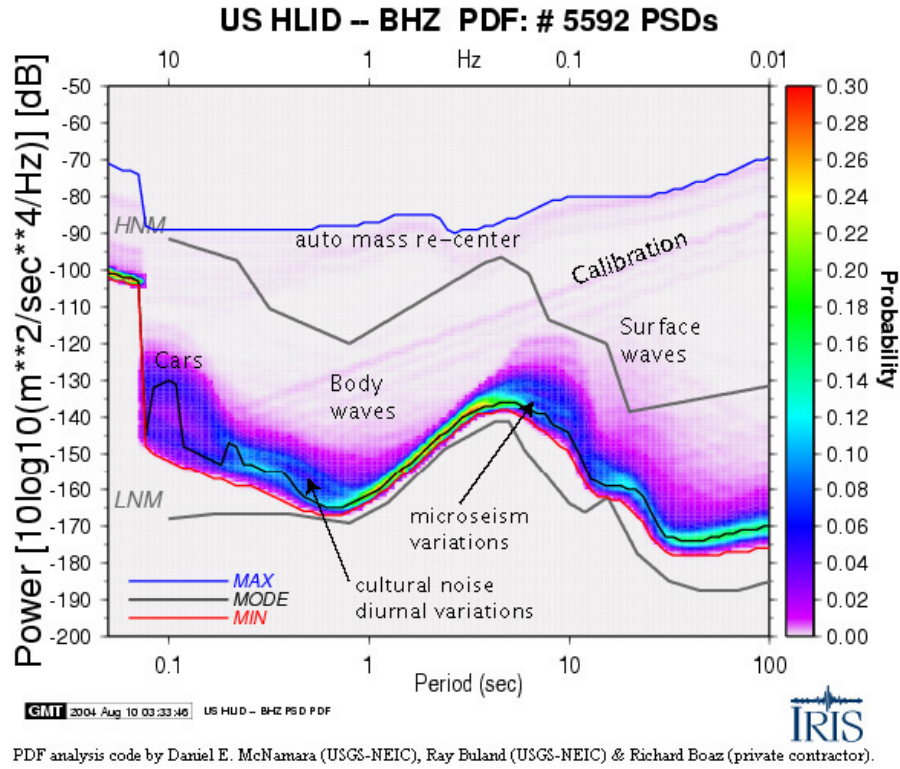
recording being the standard almost everywhere, it is usually necessary to filter data into the two bands above or below this peak to clearly detect seismic phase arrivals above the noise.



**Fig. 4.3** Envelope curves of acceleration noise power spectral density  $P_a$  (in units of dB related to  $1 \text{ (m/s}^2\text{)}^2/\text{Hz}$ ) as a function of noise period (according to Peterson, 1993). They define the new global high (NHNM) and low noise models (NLNM) which are currently the accepted standard curves for generally expected limits of seismic noise. Exceptional noise may exceed these limits. For the NLNM the related curves calculated for the displacement and velocity power spectral density  $P_d$  and  $P_v$  in units of dB with respect to  $1 \text{ (m/s}^2\text{)}^2/\text{Hz}$  and  $1 \text{ m}^2/\text{Hz}$  are given as well. Modified version of Figure 2 from P. Bormann, 1998 © Springer; with kind permission from Springer Science+Business Media.

It is often desirable to plot many PSD curves into one graph – for example, PSDs from all stations of a large network at the same time, or from many hourly or daily records of a single station. In that case, certain areas of a graph like Fig. 4.3 may become completely black, and it may be impossible to distinguish or label individual curves. A better way is then not to plot individual PSD curves but to represent the **probability density function or PDF** of the ensemble of PSD curves. (The PDF is unrelated to the .pdf format for electronic documents.) The PDF measures the relative density of PSD values (how many of the observed values lie in a small area of the graph) and is encoded with a color scale (Figure 4.4). PDF software is part of software packages like PQLX (<http://earthquake.usgs.gov/research/software/pqlx.php>; USGS Open-File Report 10-1292) or SEISCOMP3 (<http://www.seiscomp3.org/>). In interactive versions, the user can click into the plot area and obtain a listing of those seismic records

that have PSDs with the selected combination of frequency and PSD level. This can help to identify the cause of elevated noise levels, such as traffic, weather, or technical problems.



**Fig. 4.4** An example for color-coding the density (in the drawing plane) of curves of Power Spectral Density vs. period: a PDF diagram. IRIS DMC, Seattle.

**Table 4.1** Noise power spectral densities at selected periods and in different units which define the new global low-noise model (NLNM) as given by Peterson (1993). Peterson published values for  $P_a$  [dB] only. The respective numbers for ground acceleration ( $P_a$ ), velocity ( $P_v$  and  $P_v$ ) and displacement ( $P_d$  and  $P_d$ ) have been calculated using Eqs. (4.4) to (4.8). Between the listed periods the values are to be linearly interpolated in a PSD-logT diagram.

| T [s] | $P_a$ [ $m^2s^{-4}/Hz$ ] | $P_a$ [dB] | $P_v$ [ $m^2s^{-2}/Hz$ ] | $P_v$ [dB] | $P_d$ [ $m^2/Hz$ ]    | $P_d$ [dB] |
|-------|--------------------------|------------|--------------------------|------------|-----------------------|------------|
| 0.10  | $1.6 \times 10^{-17}$    | -168.0     | $4.1 \times 10^{-21}$    | -203.9     | $1.0 \times 10^{-24}$ | -239.9     |
| 0.17  | $2.1 \times 10^{-17}$    | -166.7     | $1.6 \times 10^{-20}$    | -198.1     | $1.1 \times 10^{-23}$ | -229.4     |
| 0.40  | $2.1 \times 10^{-17}$    | -166.7     | $8.7 \times 10^{-20}$    | -190.6     | $3.5 \times 10^{-22}$ | -214.6     |
| 0.80  | $1.2 \times 10^{-17}$    | -169.2     | $1.9 \times 10^{-19}$    | -187.1     | $3.2 \times 10^{-21}$ | -214.5     |
| 1.24  | $4.3 \times 10^{-17}$    | -163.7     | $1.7 \times 10^{-18}$    | -177.8     | $6.5 \times 10^{-20}$ | -191.9     |
| 2.40  | $1.4 \times 10^{-15}$    | -148.6     | $2.0 \times 10^{-16}$    | -157.0     | $3.0 \times 10^{-17}$ | -165.3     |
| 4.30  | $7.8 \times 10^{-15}$    | -141.1     | $3.6 \times 10^{-15}$    | -144.4     | $1.7 \times 10^{-15}$ | -147.7     |
| 5.00  | $7.8 \times 10^{-15}$    | -141.1     | $4.9 \times 10^{-15}$    | -143.1     | $3.1 \times 10^{-15}$ | -145.1     |
| 6.00  | $1.3 \times 10^{-15}$    | -149.0     | $1.1 \times 10^{-15}$    | -149.4     | $1.0 \times 10^{-15}$ | -149.8     |
| 10.00 | $4.2 \times 10^{-17}$    | -163.8     | $1.0 \times 10^{-16}$    | -159.7     | $2.7 \times 10^{-16}$ | -155.7     |
| 12.00 | $2.4 \times 10^{-17}$    | -166.2     | $8.7 \times 10^{-17}$    | -160.6     | $3.2 \times 10^{-16}$ | -155.0     |
| 15.60 | $6.2 \times 10^{-17}$    | -162.1     | $3.8 \times 10^{-16}$    | -154.2     | $2.3 \times 10^{-15}$ | -146.3     |
| 21.90 | $1.8 \times 10^{-18}$    | -177.5     | $2.2 \times 10^{-17}$    | -166.7     | $2.6 \times 10^{-16}$ | -155.8     |
| 31.60 | $3.2 \times 10^{-19}$    | -185.0     | $7.9 \times 10^{-18}$    | -171.0     | $2.0 \times 10^{-16}$ | -156.9     |

|        |                       |         |                       |         |                       |         |
|--------|-----------------------|---------|-----------------------|---------|-----------------------|---------|
| 45.00  | $1.8 \times 10^{-19}$ | - 187.5 | $9.1 \times 10^{-18}$ | - 170.4 | $4.7 \times 10^{-16}$ | - 153.3 |
| 70.00  | $1.8 \times 10^{-19}$ | - 187.5 | $2.2 \times 10^{-17}$ | - 166.6 | $2.8 \times 10^{-15}$ | - 145.6 |
| 101.00 | $3.2 \times 10^{-19}$ | - 185.0 | $9.7 \times 10^{-17}$ | - 160.9 | $2.1 \times 10^{-14}$ | - 136.8 |
| 154.00 | $3.2 \times 10^{-19}$ | - 185.0 | $1.8 \times 10^{-16}$ | - 157.2 | $1.1 \times 10^{-13}$ | - 129.4 |
| 328.00 | $1.8 \times 10^{-19}$ | - 187.5 | $4.9 \times 10^{-16}$ | - 153.1 | $1.3 \times 10^{-12}$ | - 118.7 |
| 600.00 | $3.5 \times 10^{-19}$ | - 184.4 | $3.2 \times 10^{-15}$ | - 144.8 | $3.0 \times 10^{-11}$ | - 105.2 |
| $10^4$ | $6.5 \times 10^{-16}$ | - 151.9 | $3.5 \times 10^{-14}$ | - 87.9  | $4.1 \times 10^{-3}$  | - 23.8  |
| $10^5$ | $4.9 \times 10^{-11}$ | - 103.1 | $1.2 \times 10^{-2}$  | - 19.1  | $2.6 \times 10^6$     | + 65.0  |

**Table 4.2** Noise power spectral densities at selected periods and in different units which define the new global high-noise model (NHNM) as given by Peterson (1993). Peterson published values for  $\mathbf{P_a}$  [dB] only. The respective numbers for ground acceleration ( $P_a$ ), velocity ( $\mathbf{P_v}$  and  $P_v$ ) and displacement ( $\mathbf{P_d}$  and  $P_d$ ) have been calculated using Eqs. (4.4) to (4.8). Between the listed periods the values are to be linearly interpolated in a PSD-logT diagram.

| T [s]  | $P_a$ [ $\text{m}^2\text{s}^{-4}/\text{Hz}$ ] | $\mathbf{P_a}$ [dB] | $P_v$ [ $\text{m}^2\text{s}^{-2}/\text{Hz}$ ] | $\mathbf{P_v}$ [dB] | $P_d$ [ $\text{m}^2/\text{Hz}$ ] | $\mathbf{P_d}$ [dB] |
|--------|-----------------------------------------------|---------------------|-----------------------------------------------|---------------------|----------------------------------|---------------------|
| 0.10   | $7.1 \times 10^{-10}$                         | - 91.5              | $1.8 \times 10^{-13}$                         | - 127.5             | $4.5 \times 10^{-17}$            | - 163.4             |
| 0.22   | $1.8 \times 10^{-10}$                         | - 97.4              | $2.2 \times 10^{-13}$                         | - 126.5             | $2.7 \times 10^{-16}$            | - 155.6             |
| 0.32   | $8.9 \times 10^{-12}$                         | - 110.5             | $2.3 \times 10^{-14}$                         | - 136.4             | $6.6 \times 10^{-17}$            | - 162.2             |
| 0.80   | $1.0 \times 10^{-12}$                         | - 120.0             | $1.6 \times 10^{-14}$                         | - 137.9             | $2.6 \times 10^{-16}$            | - 155.8             |
| 3.80   | $1.6 \times 10^{-10}$                         | - 98.0              | $5.8 \times 10^{-11}$                         | - 102.4             | $2.1 \times 10^{-11}$            | - 106.7             |
| 4.60   | $2.2 \times 10^{-10}$                         | - 96.5              | $1.2 \times 10^{-10}$                         | - 99.2              | $6.4 \times 10^{-11}$            | - 101.9             |
| 6.30   | $7.9 \times 10^{-11}$                         | - 101.0             | $8.0 \times 10^{-11}$                         | - 101.0             | $7.9 \times 10^{-11}$            | - 101.0             |
| 7.90   | $4.5 \times 10^{-12}$                         | - 113.5             | $7.1 \times 10^{-12}$                         | - 111.5             | $1.4 \times 10^{-11}$            | - 109.5             |
| 15.40  | $1.0 \times 10^{-12}$                         | - 120.0             | $6.0 \times 10^{-12}$                         | - 112.2             | $3.6 \times 10^{-11}$            | - 104.4             |
| 20.00  | $1.4 \times 10^{-14}$                         | - 138.5             | $1.4 \times 10^{-13}$                         | - 128.4             | $1.4 \times 10^{-12}$            | - 118.4             |
| 354.80 | $2.5 \times 10^{-13}$                         | - 126.0             | $8.0 \times 10^{-10}$                         | - 91.0              | $2.6 \times 10^{-6}$             | - 55.9              |
| $10^4$ | $9.7 \times 10^{-9}$                          | - 80.1              | $2.5 \times 10^{-2}$                          | - 16.1              | $6.2 \times 10^4$                | + 47.9              |
| $10^5$ | $1.4 \times 10^{-5}$                          | - 48.5              | $3.6 \times 10^3$                             | + 35.5              | $9.0 \times 10^{11}$             | + 119.6             |

### 4.3.2 Instrumental self-noise

Instrumental self-noise was no problem with classical seismographs of relatively low magnification and with large masses. According to Wielandt (see Chapter 5, section 5.5) the Brownian (thermal) motion of the seismic mass is a limiting factor only when the mass is very small (less than a few grams). Presently manufactured observatory-grade seismometers have sufficient mass to make the Brownian noise negligible against noise from other sources. However, all modern seismographs use semiconductor amplifiers. They produce, like other active (power-dissipating) electronic components, continuous electronic noise. Its origin is manifold but ultimately related to the quantization of the electric charge. Electromagnetic transducers, such as those used in geophones, also produce thermal electronic noise (resistor noise, Johnson noise). The contributions from semiconductor noise and resistor noise are often comparable, and together limit the sensitivity, i.e., the achievable resolution, of the system. In sections 5.5.6 and 5.5.7 Wielandt discusses self-noise of short-period electromagnetic seismographs and of modern broadband force-balance seismometers.

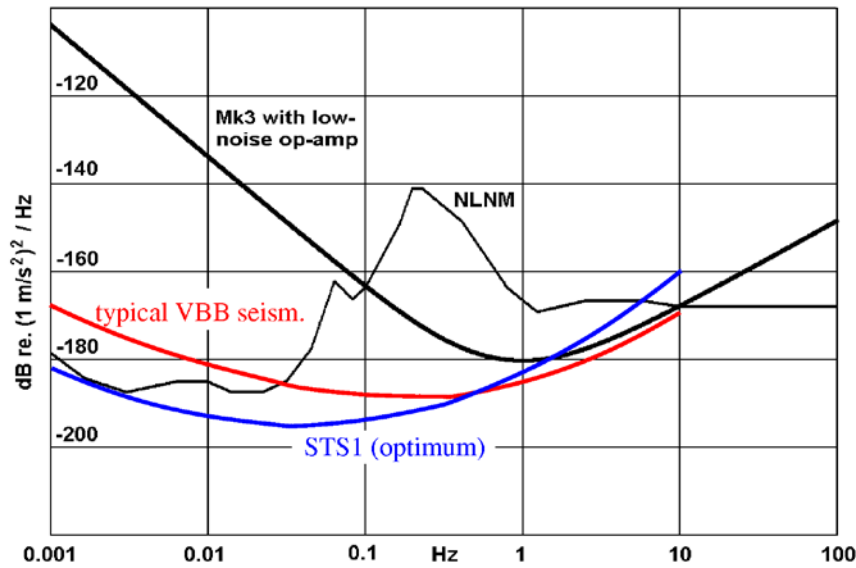
It is not possible to distinguish between ground noise and instrumental self-noise in the signal recorded from a single instrument. To identify the presence of instrumental noise one can run two or more instruments close to each other on the same pier (so-called **Huddle test**). After

correction for the transfer function of these instruments the ground noise will appear in their records as being coherent. Consequently, the difference signal between the various instruments will represent the incoherent part produced by the instruments themselves. However, it remains the question which of the instruments contributes at what degree to the difference signal? In order to distinguish between the contributions from these seismometers Sleeman et al. (2006) proposed a Huddle test with three seismometers. The power spectral density of self-noise of seismometer 1 then is

$$(4.12) \quad N_1(f) = P_1(f) [1 - (C_{12}(f) \cdot C_{13}(f))/C_{32}(f)]$$

where  $f$  = frequency,  $P_1(f)$  = power spectral density of the output of seismometer 1, and  $C_{kl}(f)$  = coherence spectrum for seismometers  $k$  and  $l$ .

Besides their more or less stationary self-noise, seismometers occasionally produce transient disturbances. They may originate in stressed mechanical parts, in slightly defective semiconductors, or in outgassing porous materials. The level of instrumental noise also depends on the installation. Only part of the instrumental noise is self-noise in a strict sense. Another part can be eliminated by careful shielding against environmental disturbances. For identifying and measuring instrumental noise see Chapter 5, sections 5.5.6 and 5.5.7. Figure 4.5 shows typical levels of self-noise for short-period and broadband seismometers.



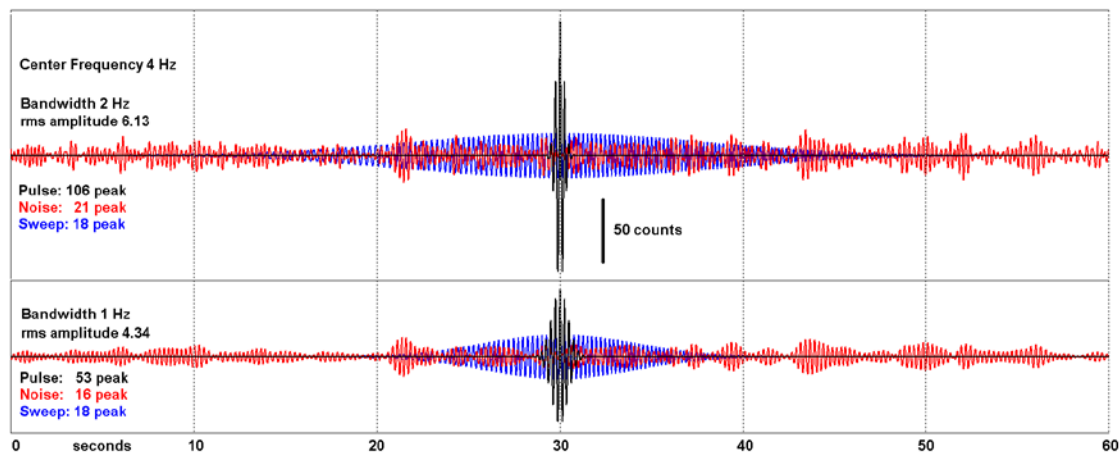
**Fig. 4.5** Typical self-noise of different seismometer types in comparison with the NLNM (thin black curve). Black: a short-period, passive Sensonics Mk3 seismometer with a low-noise preamplifier (predicted; compare the theoretical curves for different circuits in section 5.5.6 of Chapter 5). The resolution can be improved by using a larger seismic mass as in the Geotech GS-13 or some historical electromagnetic seismometers. Red: typical performance of well-installed VBB seismometers (such as STS2, Trillium 240, Guralp CMG-3T, or Reftek 151-120). Blue: at very low seismic frequencies, the STS1 (originally designed in 1975) is still the most sensitive broadband seismometer when properly installed. The STS1 is the primary sensor of the Global Seismic Network (GSN).

## 4.4 Comparing spectra of transient and stationary signals

### 4.4.1 Bandwidth and amplitude

For the purpose of the following discussion, we must subdivide the signal categories “transient” and “stationary” further: the transient signals into impulsive and oscillatory ones, the stationary signals into random signals (noise) and monochromatic sinewaves. A typical seismogram, especially a dispersed train of surface waves, are in the “oscillatory” category. Impulsive seismic signals are rare in nature but at least theoretically, the seismic signal near the focus of an earthquake should be impulsive. Ambient seismic noise is often a superposition of all types, the “random” components being due to a large number of independent sources (traffic, wind, waves in the ocean), “transient” components being caused by identifiable sources such as passing cars, and nearly monochromatic components originating from machinery such as sawmills and pumps.

When the PSD of a stationary signal is constant over the bandwidth of a seismograph or a filter and we reduce the filter bandwidth, then the signal power is proportional to the bandwidth because it is the integral of a constant power density over the bandwidth. The rms amplitude of the signal, which is the square root of the power, must therefore be proportional to the square root of the bandwidth. This explains why a “spectral amplitude density” cannot be generally defined; such a definition would imply that the amplitude is generally proportional to the bandwidth, which it is not. It may nevertheless be so in special cases. The peak amplitude of an impulsive signal can be proportional to the bandwidth when the width of the pulse is inverse to its bandwidth so that the total energy does not increase as the square of the peak amplitude. On the other hand, the peak amplitude of a sweep signal (a sinewave with decreasing or increasing



**Fig. 4.6** Each trace shows three signals with identical amplitude spectra but different phase spectra. All traces were obtained by band-pass filtration of a signal with a constant Power Spectral Density with a Gaussian filter whose center frequency was 4 Hz and total 1/e-bandwidth 2 Hz (top), resp. 1 Hz (bottom). The phase was chosen so that the complex spectrum represented an impulsive signal (a *wavelet*, black), random noise (red), and a sweep (blue). The Gaussian wavelet is the shortest pulse that can be obtained in the given bandwidth. As discussed above, its peak amplitude is proportional to the bandwidth, that of the sweep is (almost) constant, and that of the noise is roughly proportional to the square root of the bandwidth. When the bandwidth is reduced, the pulse becomes broader, the noise smoother, and the sweep shorter.

frequency) may be nearly independent of the filter bandwidth. In this case the width of the signal in time decreases with the bandwidth, and again the total energy is proportional to the bandwidth. Fig. 4.6 illustrates the effect of bandwidth on the peak amplitude of different signal types. Consequences for the evaluation of seismograms are discussed in section 4.5.

#### 4.4.2 The dynamic range of a seismograph

The **usable dynamic range** of a seismograph is in principle defined as the range of amplitudes between the smallest and the largest recordable signal. The ability to record small seismic signals is limited by ambient seismic noise, self-noise of the sensor and the digitizer, and non-seismic environmental noise to which the sensor responds. With this definition the smallest recordable signal does not only depend on technical properties of the instrument but also on seismic noise and other local conditions. For large signals a limit is reached when signals are clipped or otherwise unacceptably distorted. This limit is called the operating range of the instrument.

This definition of the dynamic range may seem clear and obvious, but it is not. The problem is that both noise and clip levels are frequency-dependent, and we must compare them in different frequency bands. However, as we have just seen, narrowing the bandwidth can change the amplitude ratios when the signals are not of the same type. The dynamic range not only depends on frequency but also on the bandwidth. We cannot define it as a function of frequency alone. This becomes clear when we investigate the detectability of a pure sinewave in the presence of random noise. An early discussion of this topic with a view to quantization noise was published by Zörn (1974).

Theoretically, no matter how small the amplitude of a sinewave is, it can always be extracted from random noise provided that we can observe the signals over an unlimited time. We can reduce the amplitude of the noise below any given level by reducing the bandwidth while the amplitude of the sinewave remains the same – we must however wait for the response of the filter to the sinewave to reach a steady state. So the dynamic range of a sensor for sinusoidal signals is theoretically infinite. A meaningful definition of the dynamic range must refer to a seismologically useful finite bandwidth in which all signal amplitudes are compared, for example, a bandwidth of one octave or one-half octave around each frequency. The bandwidth should be small enough so that distinguishable signals (such as primary and secondary microseisms, see paragraph 4.6.1) can be resolved on the frequency scale, but large enough to preserve their amplitude. (Some remarks on this topic are also found in paragraph 4.5.2.) Equivalently, considering the uncertainty principle of the Fourier transformation, one could as well specify a realistic time of observation, which cannot be smaller than a few cycles of the signal. This involves some arbitrariness and therefore it is necessary to explain what “dynamic range” exactly means in a specific case. Evans et al. (2010) suggest a particular choice that could be used as a standard.

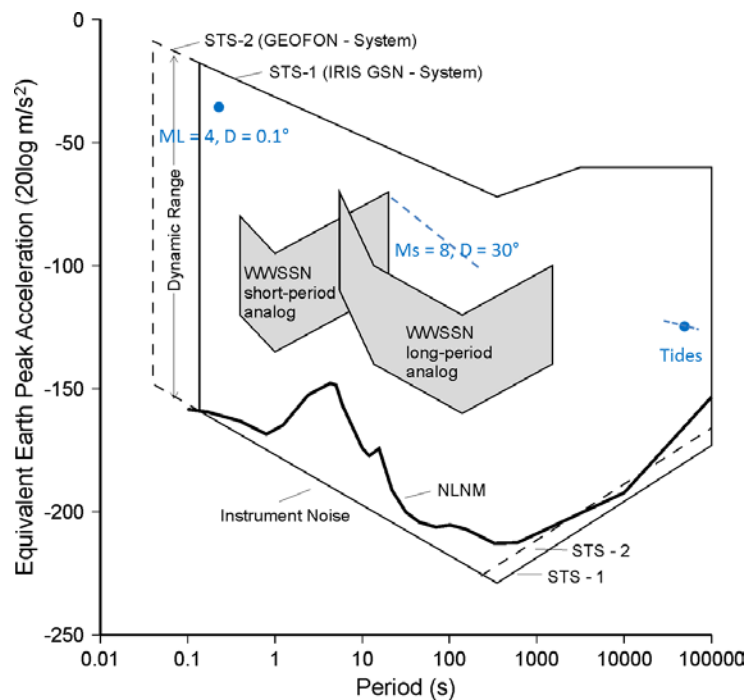
In contrast to classical analog recordings on paper, which had a dynamic range of some 40-50 dB only, modern digital data loggers with a 24 bit ADC (Analog-to-Digital Converter; see Chapter 6) allow digital recordings with a dynamic range of  $6 \times 24 = 144$  dB. This allows to cover a rather wide range of signal amplitudes and still to resolve seismic noise at the NLNM level (see Fig. 4.7). Covering a range of 144 dB in analog form would correspond to an amplitude range between 1 mm and 16 km!



### 4.4.3 Representing clip levels, signals, and noise in the same diagram

Once we have decided to represent seismic or instrumental noise by its amplitude in a small but seismologically useful bandwidth, we can at least approximately represent clip levels, typical amplitudes of seismic signals, and noise levels in the same diagram. Figure 4.7 is an example. In this case the author of the figure has chosen to annotate the vertical axis with “equivalent earth peak acceleration”, which is perfect to describe clip levels, adequate to describe expected amplitudes of earthquake signals, and a bit awkward for noise because no simple relationship exists between its peak and rms amplitudes. Under certain assumptions, however, an “average peak amplitude” of the noise can be defined, which has been done here. The assumed bandwidth is 1/3 octave.

Alternatively, one might express all signals as rms amplitudes. This would be perfect for stationary noise, a bit inconvenient for clip levels (because the rms amplitude must be multiplied with  $\sqrt{2}$  to obtain the peak amplitude of a sinewave), and awkward for seismic signals because their rms amplitude depends on the averaging time. It makes no sense to average the energy present in, say, one-half octave over the whole length of the seismogram since that energy is usually concentrated in a much shorter interval. One would rather determine the rms amplitude from the largest few cycles at each frequency. Despite the inherent arbitrariness, this kind of diagram has proved very useful for a graphical representation of seismograph performance.



**Fig. 4.7** A representation of the bandwidth and dynamic range of conventional analog (WWSSN short- and long-period) and digital broadband seismographs (STS1/VBB and STS2 with good shielding; see DS 5.1 and IS 5.4). The depicted lower bound is determined by the instrumental self-noise. The scale is in decibels (dB) relative to  $(1 \text{ m/s}^2)^2$ . Noise is measured in a constant relative bandwidth of 1/3 octave and represented by “average peak” amplitudes equal to 1.253 times the RMS amplitude. NLNM is the global New Low Noise Model according to Peterson (1993). Modified version of Fig. 7.48 in Chapter 7. Amplitudes of typical signals according to J. M. Steim (in blue).

#### 4.4.4 Approximate conversion of power densities into recording amplitudes

According to Aki and Richards (1980) the *maximum* amplitude of a wavelet  $f(t)$  near  $t = 0$  can be *roughly approximated* by the product of what they call “amplitude spectral density” and the bandwidth of the wavelet, i.e.,

$$(4.13) \quad f(t)_{t=0} = |F(\omega)| \sqrt{2(f_u - f_l)}$$

with  $f_u$  and  $f_l$  being the upper and lower corner frequencies of the band-passed signal. (Remember that such an approximation is only possible for signals whose energy is maximally concentrated in time.) Likewise, if the power spectral density of noise is defined according to Eq. (4.5) then we get for the *mean square amplitude* of noise in the time domain:

$$(4.14) \quad \langle f^2(t) \rangle = 2P(f_u - f_l)$$

This is of course a simplified version of the general rule that the power spectral density (PSD) must be integrated over the passband of a filter to obtain the power (or *mean square amplitude*) at the output of the filter. The square root of this power is then the *root mean square* (RMS) or effective amplitude

$$(4.15) \quad a_{\text{RMS}} = [2P \times (f_u - f_l)]^{1/2}$$

If the power spectral density is defined according to the engineering approach (see 4.2.3.1) as  $P_e = 2P$ , then

$$(4.16) \quad a_{\text{RMS}} = [P_e \times (f_u - f_l)]^{1/2}$$

We note again that stationary signals must be characterized by their PSD, and **specifying seismic noise by its RMS amplitudes is meaningless without definition of the bandwidth**. The values given by the NLNM and NHNM in Fig. 4.3 and Tabs. 4.1 and 4.2, respectively, follow the engineering convention. For consistency, we will use this convention for the rest of this chapter.

For the reasons discussed in section 4.4, it is often necessary to represent the amplitude of noise (whether rms or average-peak) in a **constant relative bandwidth RBW** over the whole frequency range. This means, the bandwidth  $B$  around each frequency  $f_0$  is a fixed fraction or multiple of  $f_0$ . The lower and upper band limits,  $f_l$  and  $f_u$ , are chosen so that  $f_0$  is the geometric mean of the two (so  $f_0$  appears in the middle between  $f_l$  and  $f_u$  on a logarithmic frequency scale). When the bandwidth is  $n$  octaves or  $m$  decades ( $n$  and  $m$  normally being fractions, not integers), then the following relationships hold:

$$(4.17) \quad f_l = 2^{-n/2} f_0, \quad f_u = 2^{n/2} f_0, \quad \text{RBW} = 2^{n/2} - 2^{-n/2} \text{ and } B = (2^{n/2} - 2^{-n/2}) f_0$$

and (4.16) can be written as

$$(4.18) \quad a_{\text{RMS}} = [P_l \times (f_u - f_l)]^{1/2} = [P_e \times f_0 \times (2^{n/2} - 2^{-n/2})]^{1/2} = (P_e \times f_0 \times \text{RBW})^{1/2}$$

Octaves can be converted easily into decades and vice versa by using the relation

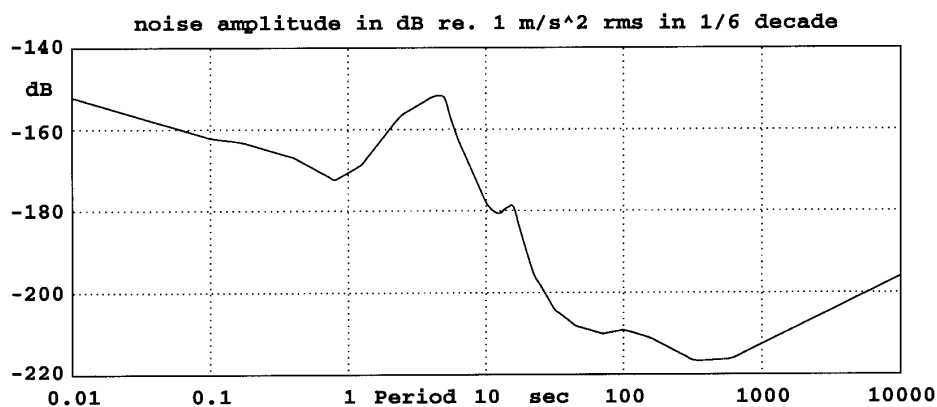
$$(4.19) \quad m = n \cdot \log_{10}(2) = n \cdot 0.3010$$

Aki and Richards (1980, vol.1, p. 498) converted PSD into ground motions by putting the bandwidth of the noise signal at half the considered (center) frequency, i.e., by assuming  $f_u - f_l = 0.5 f_0$ . This corresponds to a relative bandwidth of roughly 2/3 octave. Using the definition of power on which Eq. (4.14) is based they obtained  $a_{\text{RMS}} = (P \times f_0)^{1/2}$ .

**Table 4.3** Bandwidth  $B$  of typical octave and decade filters centered at  $f_0$ .

| Octaves | $B$         | Decades | $B$         |
|---------|-------------|---------|-------------|
| 1       | $0.707 f_0$ | 1       | $2.846 f_0$ |
| 2/3     | $0.466 f_0$ | 1/2     | $1.215 f_0$ |
| 1/2     | $0.348 f_0$ | 1/3     | $0.786 f_0$ |
| 1/3     | $0.232 f_0$ | 1/6     | $0.386 f_0$ |

Other authors (e.g., Fix, 1972; Melton, 1978) have used an integration bandwidth of 1/3 octave (a standard bandwidth in acoustics) for computing RMS amplitudes from PSD. Melton reasoned that this is nearly  $\pm 10\%$  about the center period in width and thus close to the tolerance with which an analyst can measure the period on an analog seismogram. Therefore, using a 1/3 octave bandwidth seemed to him a reasonable convention for calculating RMS noise amplitudes from PSD. The differences, as compared to RMS values based on 1/4 or 1/2 octave bandwidths, are less than 20%. But 1/3 octave amplitudes will be only about 70% or 50% of the respective RMS amplitudes calculated for 2/3 or 4/3 octave bandwidth, respectively. Typical response curves of short-period narrowband analog seismographs for recording transient teleseismic body-wave onsets have bandwidths between about 1 and 2 octaves. For deriving spectral magnitudes, Duda and Kaiser (1989) filtered broadband records with one octave wide bandpasses in intervals of one octave distance with center periods of 0.25 s, 0.5 s, ... 32 s, and 64 s; see Fig. 3.55 in Chapter 3). Fig. 4.8 represents the NLNM (compare Fig. 4.3 ) as rms amplitudes in a constant relative bandwidth of 1/6 decade.



**Fig. 4.8** The USGS New Low Noise Model, here expressed as RMS amplitudes of ground acceleration in a constant relative bandwidth of one-sixth decade, corresponding approximately to 1/2 octave (see Tab. 4.1). A bandwidth of 1/6 decade extends from 82.5% to 121% of the central frequency  $f_0$ . The corresponding values for 1/2 octave are 84% and 119%.

There is a 95% probability that the instantaneous amplitude of random noise with a Gaussian amplitude distribution will lie within a range of  $\pm 2a_{\text{RMS}}$ . Peterson (1993) showed that both broadband and long-period noise amplitudes follow closely a Gaussian probability distribution. This is true for largely natural ambient noise, often not, however, for short-period and broadband noise in urban and industrialized areas. There the noise is often dominated by transient or periodic signals (see, e.g., Figs. 7.17, 7.19, and 7.21 in Chapter 7; also Groos and Ritter, 2009 and section 4.6.3). Yet, in case of a Gaussian probability distribution, the absolute peak amplitudes of the narrowband filtered signal envelopes should follow a Rayleigh distribution. In the case of an ideal Rayleigh distribution the theoretical *average peak amplitudes* (APA) are  $1.253 a_{\text{RMS}}$ . From test samples of narrowband filtered VBB and LP noise records Peterson (1993) measured APA values between 1.194 and 1.275. Therefore, RMS amplitudes in 1/6-decade bandwidth correspond approximately to average peak amplitudes in 1/3 octave bandwidth. An example: According to Fig. 4.8 the minimum vertical ground noise between 10 and 20 s is at -180 dB relative to  $1\text{m/s}^2$ . This corresponds to average peak amplitudes of  $10^{-180/20} \text{m/s}^2 = 1 \text{ nm/s}^2$  in 1/3 octave bandwidth. Accordingly, the total average peak amplitude in this one octave band between 10 and 20 s is approximately  $\sqrt{3} \text{ nm/s}^2$ .

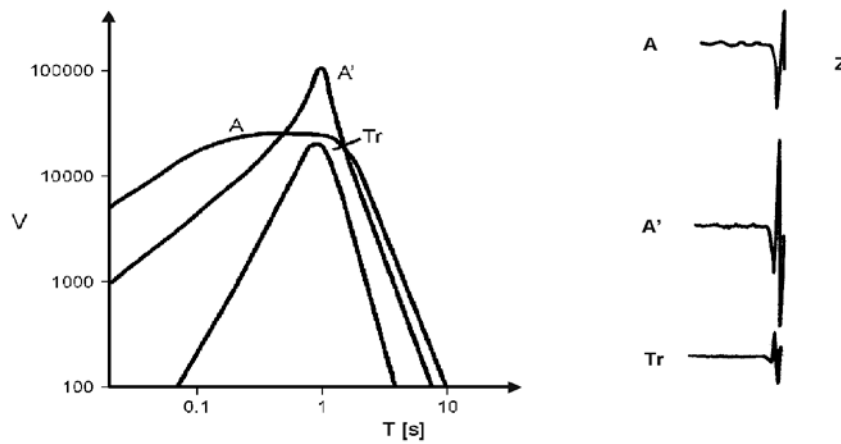
E. Wielandt offers an interactive program NOISECON which can be downloaded via the link from the NMSOP-2 front page to the listing **Download programs and files**. It converts noise specifications into all kinds of standard and non-standard units and compares them to the USGS NLNM. EX 4.1 provides exercises by hand for calculating RBWs and transforming PSDs into  $a_{\text{RMS}}$  for various kinematic units and bandwidths. It is complemented also by several exercises combining eye-estimates and NOISECON applications for interpreting and converting noise spectra.

## 4.5 Seismograph response and waveform

### 4.5.1 Empirical case studies

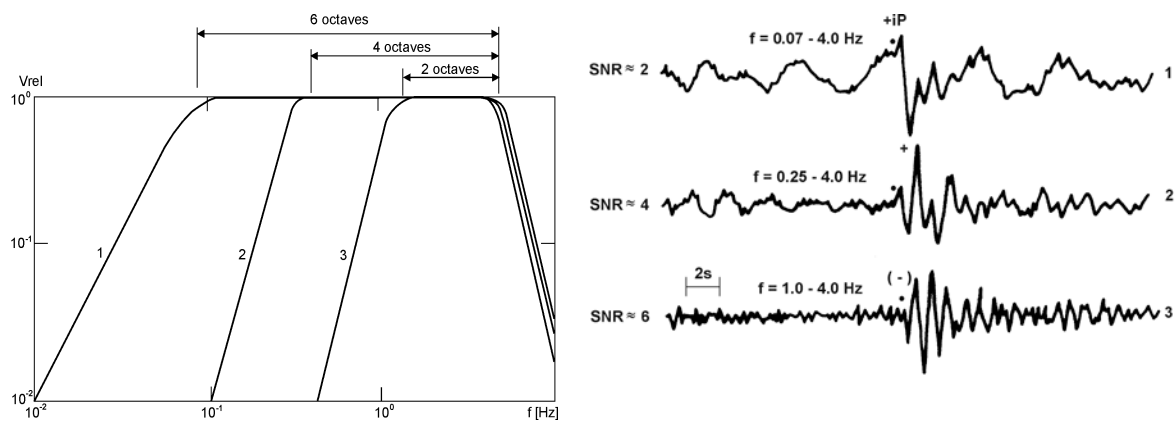
While in recordings of type A with four-octave bandwidth, the first half cycle contains the largest amplitude, the maximum amplitude in records with 1/2 octave (type A') and one octave bandwidth (type Tr) is reached only at the third half-swing. Also, the first motion amplitude in the one octave record Tr is strongly reduced as compared to that in record A with four octave bandwidth, despite having nearly the same peak magnification. Accordingly, we have to consider that in narrowband records of high magnification (as with WWSSN short-period seismographs; bandwidth about 1.5 octaves) the reduced first motion amplitudes might get lost in the presence of noise. Since reliable first motion polarity readings are crucial for the determination of fault plane solutions (see section 3.4 in Chapter 3) and discriminating earthquakes from explosions, narrowband recordings might result in an unacceptable loss of primary information.

Fig. 4.9 shows earthquake P-wave onsets recorded by short-period seismographs with 1-Hz seismometers but amplitude response curves of different bandwidth. The maximum amplitude of the P-wave onset in the – as compared to the filter in Fig. 4.11 – more narrowband Tr record is only about 1/2 of that in record A although seismographs have nearly the same peak magnification at 1 Hz for steady-state harmonic oscillations. And in record A' the maximum amplitude is only twice as large, even though the A' instrument has four times larger peak amplification than instrument A.

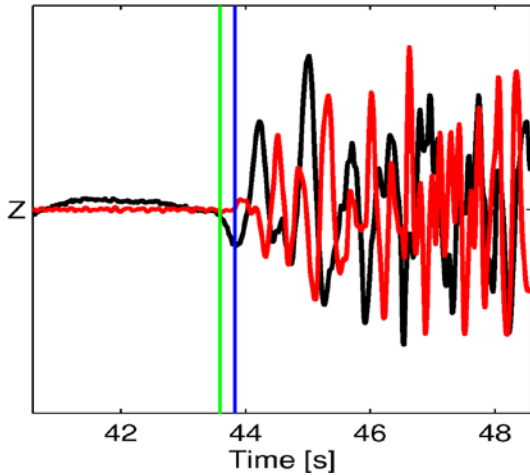


**Fig. 4.9** Left: Displacement amplitude magnification curves of three traditional analog types of short-period seismographs used at seismic station MOX with 1/2 octave (type A'), one octave (Tr - trigger seismograph) and four octave bandwidth (type A), respectively; right: records with these seismographs of a P-wave onset of a deep earthquake at an epicentral distance of  $72.3^\circ$  and hypocentral depth of 544 km (copied from Bormann, 1973).

The following figures, 4.10 to 4.14, illustrate the difficulty of interpreting narrowband seismograms correctly with respect to polarities, amplitudes, and the presence and onset times of secondary arrivals.



**Fig. 4.10** A medium-period velocity-proportional digital broadband record (bandwidth almost 6 octaves between 0.07 – 4 Hz;) at station MOX of an underground nuclear explosion at the Nevada test site (upper record trace), and the same signal filtered with a 4-octave and 2-octave bandpass filter (middle and lower record trace). The positive first motion (to be expected from an explosion) is clearly seen in the BB record despite of the low SNR, but it is buried in the noise of the 2 octave record despite the general SNR improvement due to narrowband filtering. The different absolute amplitude levels in the three records have been normalized to the same peak amplitude.

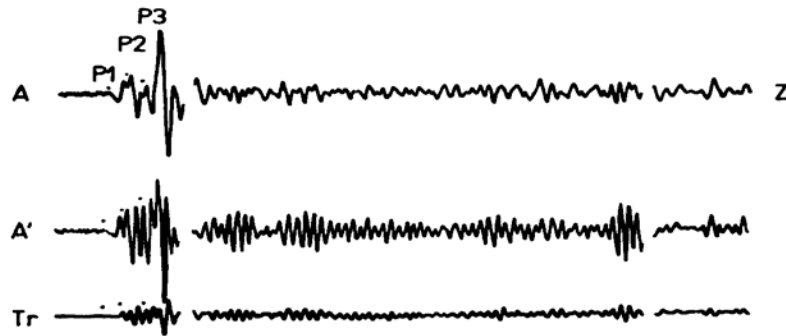


**Fig. 4.11** Unfiltered (black) and filtered (red) local event waveform of a broadband record (Guralp CMG, 60s–50 Hz) using a 3rd order Butterworth bandpass filter (2–10 Hz). In the more narrow-band high-frequency filtered record the amplitude of the first oscillations is strongly reduced and the phase significantly shifted. The negative first motion, still distinct in the broadband record, is no longer recognizable (copy of Fig. 16.25 in Chapter 16 with permission of the authors Küperkoch et al., 2012)

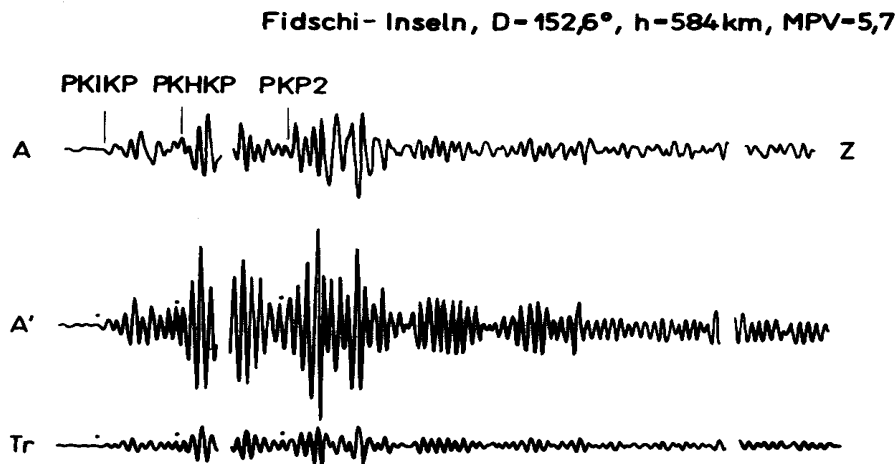
This systematic underestimation of amplitudes of transient body-wave onsets of short duration in narrowband records - and thus of related magnitude estimates within the first few cycles as for many mb data (Chapter 3, section 3.2.5.2 and Fig. 3.46) - has been a matter of considerable debate between the American and Russian delegations in the early Geneva talks to achieve a Comprehensive Nuclear-Test-Ban Treaty (CTBT). In the Soviet Union standard seismographs with amplitude characteristics of type A (2 to 4 octaves) and broadband characteristics of type Kirnos with about 7 octaves bandwidth (Fig. 3.20 in Chapter 3 and Figure 1 in IS 3.7) were used to determine body-wave magnitudes, whereas American-designed WWSSN stations had a bandwidth of only about 1.5 octaves, and short-period P-wave amplitudes were measured within the first few cycles only for mb magnitude determinations (Engdahl and Gunst, 1966). A consequence of these differences in magnitude determination was that the American delegation reported a much larger number of weak, unidentified seismic events per year than the Soviet delegation and therefore felt that they required hundreds of U.S. unmanned stations on Soviet territory as well as the possibility for on-site inspections. This blocked, amongst other reasons, the agreement on a comprehensive test-ban treaty for two decades. Today these problems are more of historical interest. Present days analysts may filter digital broadband data as desired. But it remains a problem to decide what filter is appropriate for solving best the given task. Therefore, analysts should be aware of both the reasons and possible size of filter effects. For assuring globally compatible magnitude data, however, the International Association of Seismology and Physics of the Earth's Interior (IASPEI, 2005 and 2013 and IS 3.3) has now agreed on some classical filter standards (such as Wood-Anderson, WWSSN-SP and WWSSN-LP) to be applied before measuring the amplitudes and periods for the most widely used types of magnitudes while other responses, such as the Kirnos-D displacement broadband, have proven to be best suited for teleseismic secondary phase recognition and onset time picking.

A narrower bandwidth causes a longer and more oscillating record of an impulsive onset. This makes it difficult to recognize, in narrowband records, secondary onsets following closely behind the first one, e.g., onset sequences due to a multiple earthquake rupture (Fig. 4.12), depth phases in the case of shallow earthquakes or branching/crossing of travel-time curves (see Chapter 2 and Fig. 4.13). But the identification and proper time picking of such closely spaced secondary arrivals may be crucial for a better understanding of complex rupture processes, for improved estimates of hypocenter depth or for studies of the fine structure of the Earth.

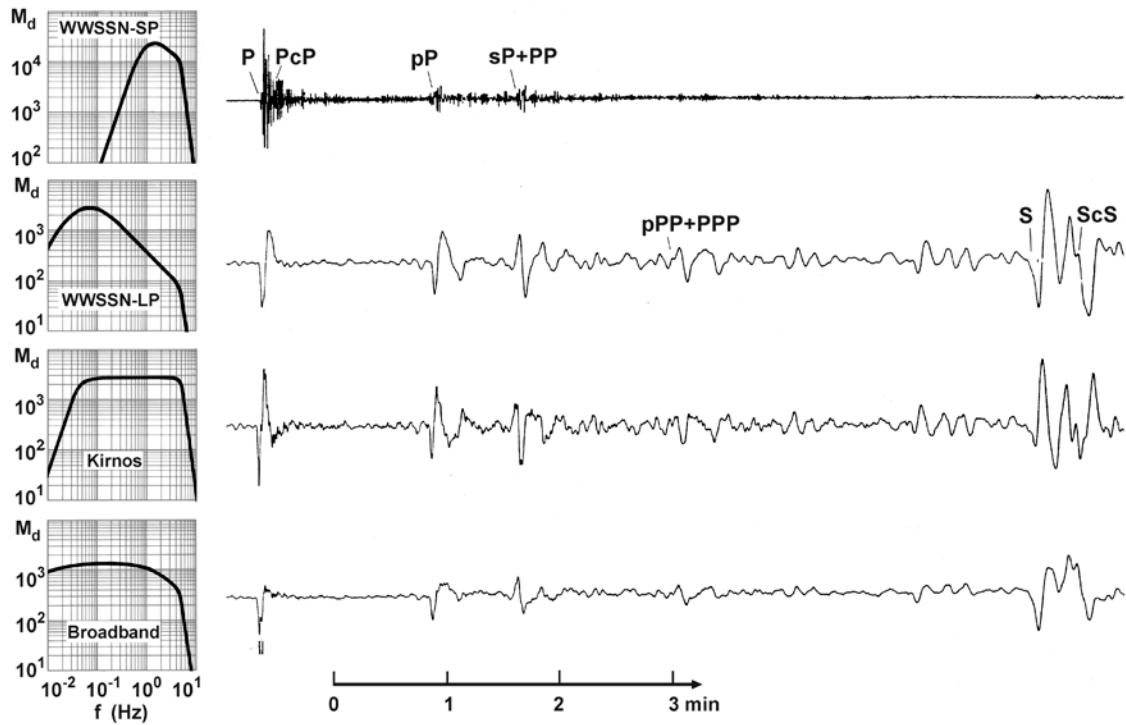
It is crucial, therefore, to record seismic signals with as large a bandwidth and with as high a linearity, resolution and dynamic range as possible, thus preserving the primary information with least distortion. Filtering should only be applied afterwards, as required for special purposes. With feedback-controlled broadband seismometers and digital data loggers with 24 bit resolution being common nowadays, this is no longer a problem (see Chapters 5 and 6). In Fig. 4.14 it is clearly recognizable that in the displacement-proportional broadband record of about 10 octaves bandwidth the P-wave onset looks rather simple (negative impulse with only slight positive over-swing of the second half-cycle). Its appearance resembles the expected source displacement pulse in the far field (see Fig. 2.4).



**Fig. 4.12** Short-period records of station MOX according to Bormann (1973) of a multiple rupture event at Honshu ( $D = 88.0^\circ$ ) with different amplitude response characteristics according to Fig. 4.9 left.



**Fig. 4.13** Short-period records of station MOX according to Bormann (1973) of a sequence of core phases corresponding to the travel-time branches PKPdf (PKIKP), PKPbc (old PKHKP) and PKPab (PKP2) (see Chapter 11) with different amplitude response characteristics according to Fig. 4.9 left.



**Fig. 4.14** Records of a deep earthquake ( $h = 570$  km,  $D = 75^\circ$ ) at the Gräfenberg Observatory, Germany. They have been derived by filtering a velocity-proportional digital broadband record (passband between 0.05 and 5 Hz) according to the response curves of some traditional standard characteristics (WWSSN\_SP and LP, Kirnos) while the bottom trace shows the result of computational restitution of the (nearly real) true ground displacement by extending the lower corner period  $T_0$  well beyond 100s (see text) (from Buttkus, 1986; with permission of Schweizerbart [www.schweizerbart.de](http://www.schweizerbart.de)).

#### 4.5.2 Theoretical considerations on signal distortion in seismic records

The basic theory of seismometry is outlined in Chapter 5. For a more comprehensive introduction to general filter theory and its applications in digital seismology (with exercises) see “Of Poles and Zeros: Fundamentals of Digital Seismology” by Scherbaum (3<sup>rd</sup> revised edition, 2007). The book is accompanied by a CD-ROM “Digital Seismology Tutor” by Schmidtke and Scherbaum, which is a very versatile tutorial tool for demonstrating signal analysis and synthesis. E. Wielandt also offers the program **fildemo**. It can be downloaded via the link to *Download Programs & Files* on the NMSOP-2 front page or via Internet (<http://www.geophys.uni-stuttgart.de/~erhard/downloads/> or <http://www.software-for-seismometry.de/>). Therefore, we will not dwell on it further, however, we will illustrate by way of example some of the essential effects of signal distortion by the transfer function of the seismograph. Signal distortion due to wave propagation effects in the Earth and ways how to eliminate at least some of them are discussed in Chapter 2.

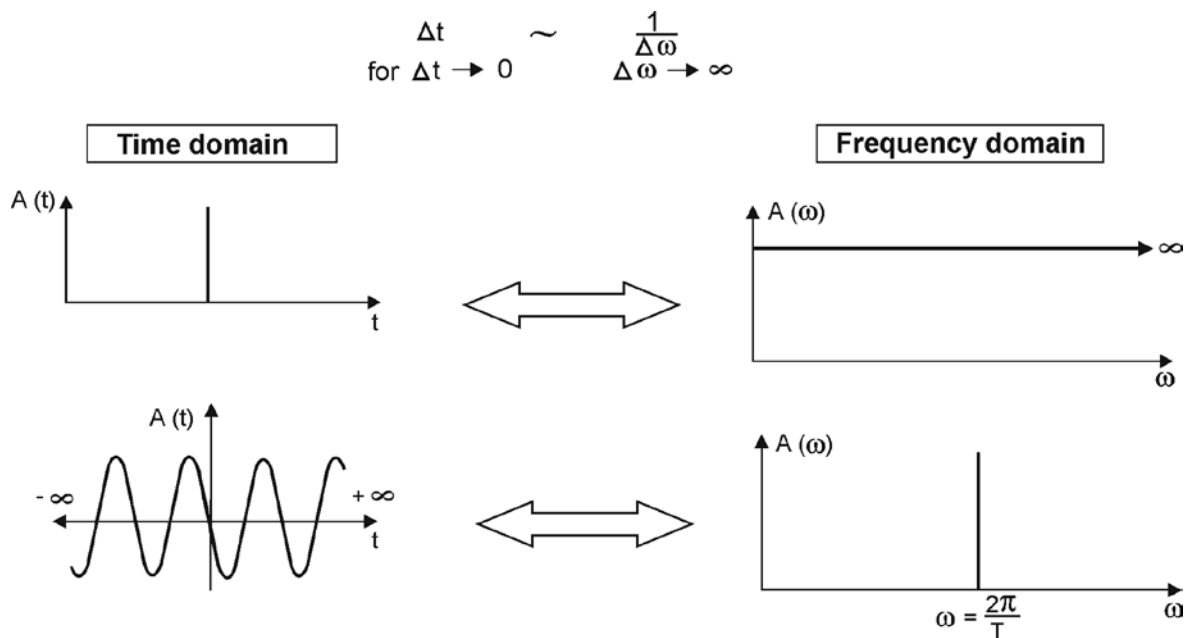
As implied by Eq. (4.2), a Dirac (or needle) impulse (see Chapter 5, section 5.2.4) in the time domain is equivalent to a constant (“white”) amplitude spectrum in the frequency domain. Thus, if the far-field seismic source pulse comes close to a needle impulse of very short duration (e.g., an explosion) we would need in fact a seismograph with (nearly) an infinite bandwidth in order to be able to reproduce this impulse-like transient signal. On the other hand, a



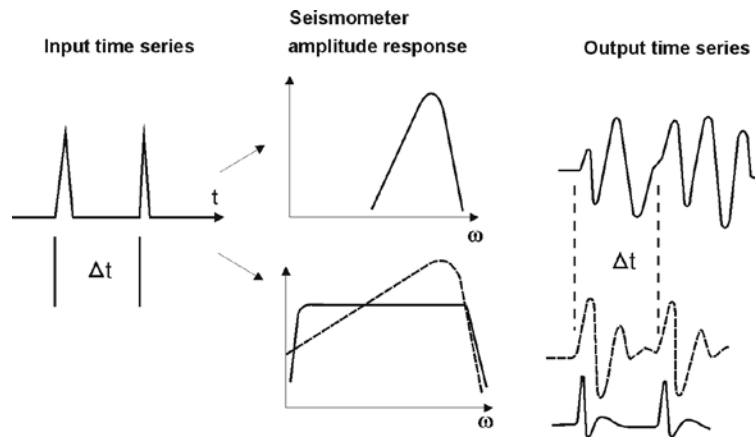
monochromatic harmonic signal corresponds to just one spectral line in the frequency spectrum, or, the other way around, if the input signal is a needle impulse with an infinite spectrum but the bandwidth of the seismograph is extremely narrow, then the output signal would not be a needle impulse but rather a slowly growing and decaying sinusoid. Fig. 4.15 depicts these extreme cases and Fig. 4.16 sketches seismographic recordings of an impulse sequence with qualitatively different response characteristics.

According to theoretical considerations by Seidl and Hellweg (1988), the seismometer period  $T_0$  has to be about 100 times larger than the duration  $\tau_s$  of the source-time input function when the source signal shape and its “signal moment” (area under the impulse time curve) is to be reproduced with a relative error  $< 8\%$ . As a rule of thumb, these authors state that the relative error is less than about 10% if  $T_0 > 20 \pi \tau_s$ . This means that extreme long-period seismographs would be required to reproduce with sufficient accuracy the displacement impulse of strong seismic events.

By “signal restitution” (i.e., instrument response correction or “deconvolution”) procedures (Seidl, 1980; Seidl and Stammer, 1984; Seidl and Hellweg, 1988; Ferber, 1989) the eigenperiod of long-period feedback seismometers such as STS1 ( $T_0 = 360$  s) and STS2 ( $T_0 = 125$  s) can be computationally extended - in the case of high signal-to-noise ratio - by a factor of about 3 to 10, and thus the very low-frequency content of the signals can be retrieved. Simulations of standard frequency responses from BB records are available in some of the software packages for signal pre-processing (e.g., PREPROC, Plešinger et al. 1996) or seismogram analysis (e.g., SEISAN, Havskov and Ottemöller, 1999 b, and Seismic Handler by K. Stammer; see <http://ftp.geo.uib.no/pub/seismo/SOFTWARE/SEISAN/> and <http://www.szgrf.bgr.de/sh-doc/index.html>, respectively).

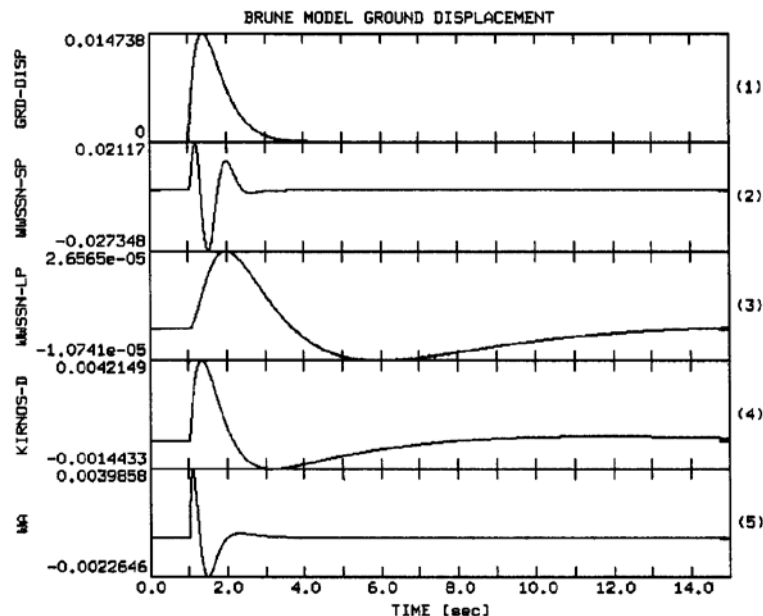


**Fig. 4.15** Sketch of the equivalent representation of a needle impulse (above) and a stationary infinite monochromatic harmonic signal (below) in the time and frequency domain.



**Fig. 4.16** Schematic illustration of the appearance of a sequence of seismic input impulses in record outputs of seismographs with narrow-band displacement response (uppermost trace) and broadband responses (lower traces: broken line – velocity response, full line – displacement response).

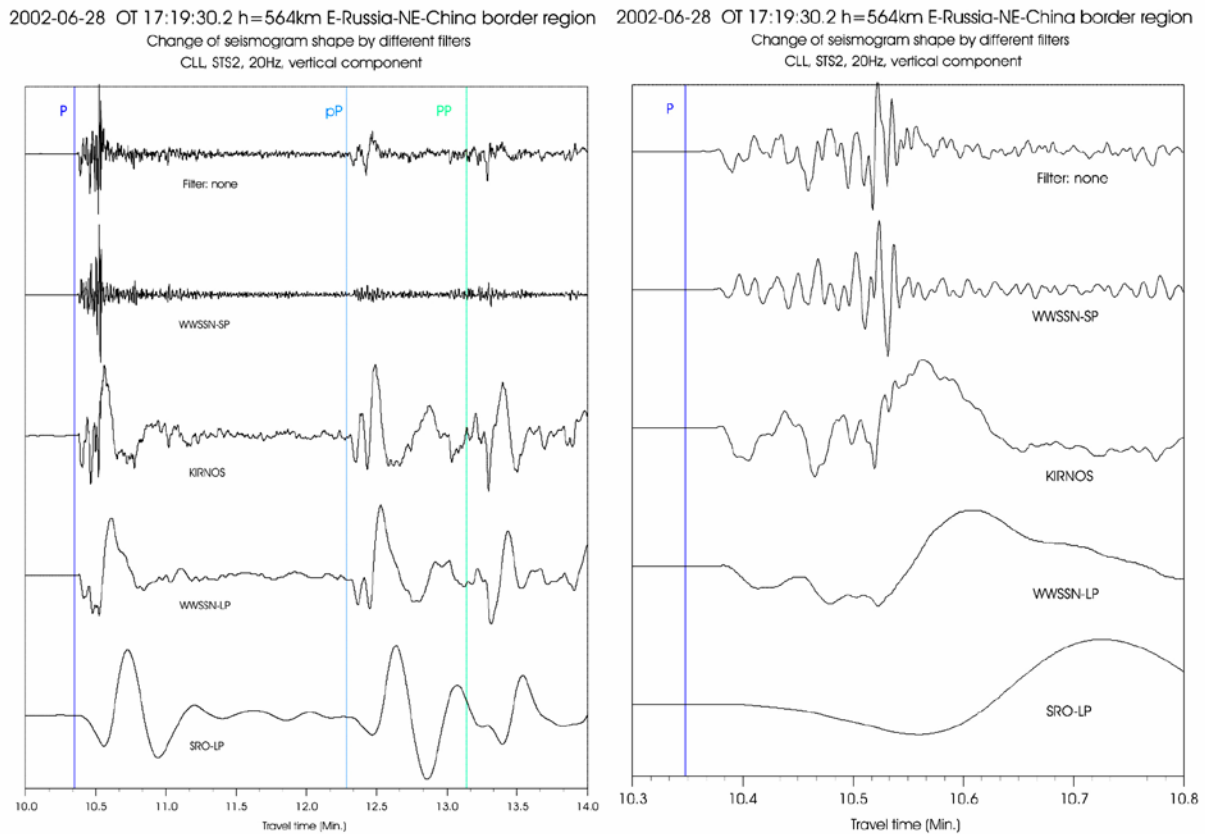
Fig. 4.17 shows the very different response of three standard seismograph systems of different damping and bandwidth to a synthetic ground displacement input according to the Brune model of earthquake shear dislocation. The response has been simulated by using the PITSA seismological analysis software of Scherbaum and Johnson (1992). For the amplitude response of the seismographs of type Wood-Anderson (WA), Kirnos, WWSSN long-period (LP) and WWSSN short-period (SP) see Fig. 3.20 in Chapter 3.



**Fig. 4.17** Distortion of a synthetic ground displacement signal according to the Brune model of earthquake shear dislocation (top trace) by standard seismograph systems (for their response curves see Fig. 3.20 in Chapter 3) (from Scherbaum 2001, “Of Poles and Zeros”, Fig. 10.2, p. 167 © Springer; with kind permission from Springer Science+Business Media).

The strong distortion in narrowband recordings after transient onsets is due to their pronounced *transient response* (TR). It is due to the time required by the seismograph system to achieve the values of frequency-dependent magnification and phase shift determined by its amplitude- and phase-frequency characteristics for steady-state harmonic oscillations (see Chapter 5, Fig. 5.3 and Figure 1a and b in IS 5.2). In Fig. 4.17, the effect of phase-shift adaptation during the time of transient response is clearly seen, especially in the records of the WWSSN and Wood-Anderson short-period instruments. Accordingly, the period of the first half cycle appears to be much shorter than that of the second and third half cycle.

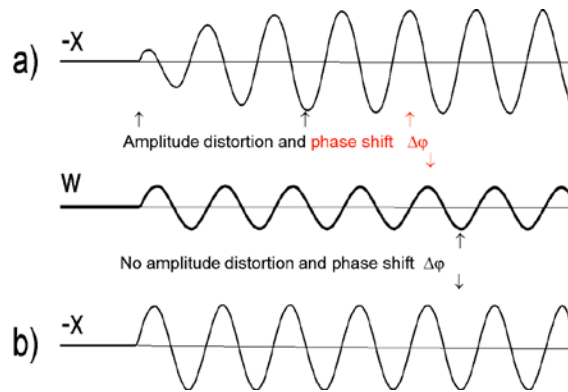
Fig. 4.18 compares the response of the same seismographs and of the SRO-LP seismograph with the unfiltered velocity broadband record of the STS2 (see DS 5.1) from an earthquake in the Russia-China border region. The differences in record appearance depending on the response characteristic of the seismograph and the time resolution of the record are striking.



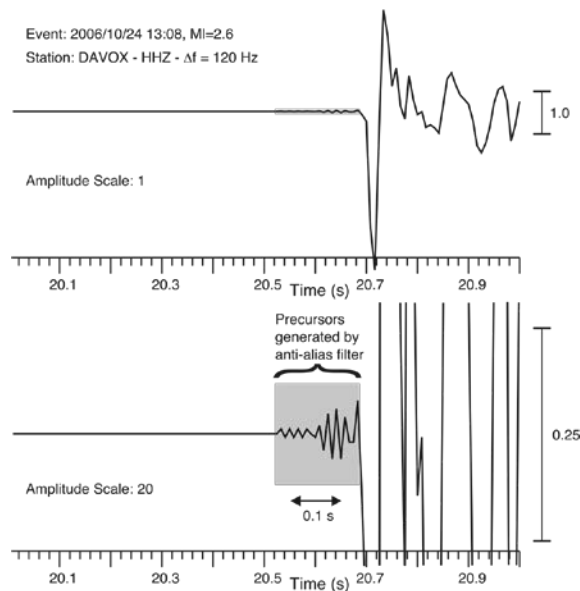
**Fig. 4.18** Record segments from an earthquake at the Russia-China border of 4 min (left) and 30 s duration (right). Uppermost traces: unfiltered STS2 velocity broadband seismogram; other traces: filtered records which simulate the seismograms of standard recordings of type WWSSN-SP, Kirnos, WWSSN-LP and SRO-LP (courtesy of S. Wendt, 2002).

Fig. 4.19 gives an example from a simulation of seismometer response to a monochromatic harmonic ground motion  $w(t)$  of frequency 1 Hz as input. It has been compiled from two different snapshots of an interactive web site demonstration of the Technical University of Clausthal, Germany, which is no longer available. Trace a) has been recorded with a seismometer of eigenperiod  $T_s = 1$  Hz and damping  $D_s = 0.2$  (i.e., resonance at 1 Hz!) while for trace

b)  $T_s = 20$  s and  $D_s = 0.707$ . In the first record the transient response takes about 3 s before the steady-state level of constant amplitudes corresponding to the amplitude response of the seismometer and the constant phase shift of about  $110^\circ$  have been reached (after the 6<sup>th</sup> half cycle). In record b) the transient response takes less than half a second and the seismometer mass follows the ground motion with practically no phase shift. Although the transient response of the 20 s long-period seismometer is much longer than that of the 1 s short-period seismometer, it is not excited here because the signal has very little energy at the eigenperiod.



**Fig. 4.19** Simulation of displacement signal output  $x(t)$  (= relative displacement of the seismometer mass) of a spring-mass pendulum seismometer responding to a monochromatic harmonic ground motion  $w(t)$  of period  $T = 1$  s (thick line in the middle). a) Displacement output of a seismometer with low damping ( $D_s = 0.2$ ) and eigenperiod  $T_s = 1$  s (i.e., resonance); b) Displacement output of a long-period and normally damped seismometer ( $T_s = 20$  s;  $D_s = 0.707$ ). For discussion see text.



**Fig. 4.20** Precursors caused by acausal anti-alias filtering observed on the broadband registration of a local earthquake near Davos, Switzerland (2006/10/24 13:08,  $M_l = 2.6$ ). Sampling frequency is 120 Hz. The precursors become obvious when enlarging the amplitude scale (lower trace). (Copy of Figure 7 in IS 11.4; courtesy of Diehl et al., 2012)

Seismographs are real physical, i.e., causal filters. They respond to analog input signals with unavoidable phase shifts during the transient response time. Digital filters, however, can be designed in a way that they do not produce phase shifts (Scherbaum, 2007; NMSOP Chapter 6). Such systems implement acausal digital anti-alias filters with symmetrical zero-phase shift impulse responses. As a consequence, the onset of impulsive signals (also of signals with frequency close to the Nyquist frequency) may be obscured by artifact precursory oscillations and their true onset becomes difficult, if not impossible, to determine. Fig. 4.20 shows an example. As outlined in IS 11.4, this effect can in principle be minimized by an inverse filtering process, described e.g. in Scherbaum (2001 and 2007). However, this is a nontrivial procedure that requires knowledge of the original anti-alias filter coefficients. Especially for large and inhomogeneous data sets, compiling the correct information can be rather difficult or even impossible.

## 4.6 Characteristics and sources of seismic noise

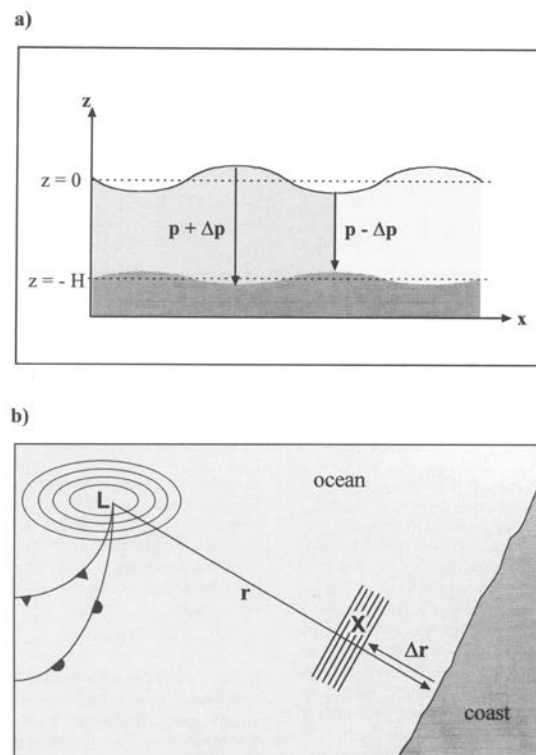
### 4.6.1 Ocean microseisms

Most of the early 20<sup>th</sup> century seismographs by Wiechert, Mainka, Galizyn, Bosch-Omori, Milne-Shaw and others were medium-period broadband systems. The more sensitive ones with 100 to 500 times magnification were already able to record the ever-present ground motions with peak amplitudes around 6 s which constitute the background noise for any seismic measurement. They were termed microseisms. Such recordings were first reported by Algue in 1900. Wiechert (1904) proposed at the second international seismological conference that these microseisms are caused by ocean waves on coasts. Later it was found that one must discriminate between: a) the smaller primary ocean microseisms with periods between 10 and 20 s, typically around  $14 \pm 2$  s and b) the secondary or *double frequency microseism* which is related to the main noise peak around  $6 \pm 2$  s (see Fig. 4.3, Fig. 4.4, and Fig. 4.5).

Primary ocean microseisms are generated only in shallow waters in coastal regions. Here the wave energy can be converted directly into seismic energy either through vertical pressure variations, or by the smashing surf on the shores, which have the same period as the water waves ( $T \approx 10$  to 16 s). Haubrich et al. (1963) compared the spectra of microseisms and of swell at the beaches and could demonstrate a close relationship between the two data sets. Energy is transferred between ocean waves and seismic waves through non-linear interaction of ocean waves and bathymetry (Hasselmann, 1963). Measurements from instruments in coastal regions show spectral peaks that track ocean wave frequencies incident at the coast, although this energy decays rapidly inland (Haubrich and McCamy, 1969). The peak is also called the single-frequency peak because it mimics swell frequencies. In the center of continents or of an ocean basin, the single-frequency peak is related to storm waves on remote coastlines (Cessaro, 1994). The relative stability in amplitude for continental sites around the globe suggests persistent, stable sources. Holcomb (1998) also reports on a persistent peak of unknown origin in the noise spectrum at 26 s period. The signals are largest during the southern winters and the amplitude of the peak is correlated between sites worldwide.

Contrary to the primary ocean microseisms, the secondary ocean microseisms could be explained by Longuet-Higgins (1950) and Hasselmann (1963) as being generated by the superposition of ocean waves of equal period traveling in opposite directions, thus generating standing gravity waves of half the period. These standing waves cause non-linear pressure

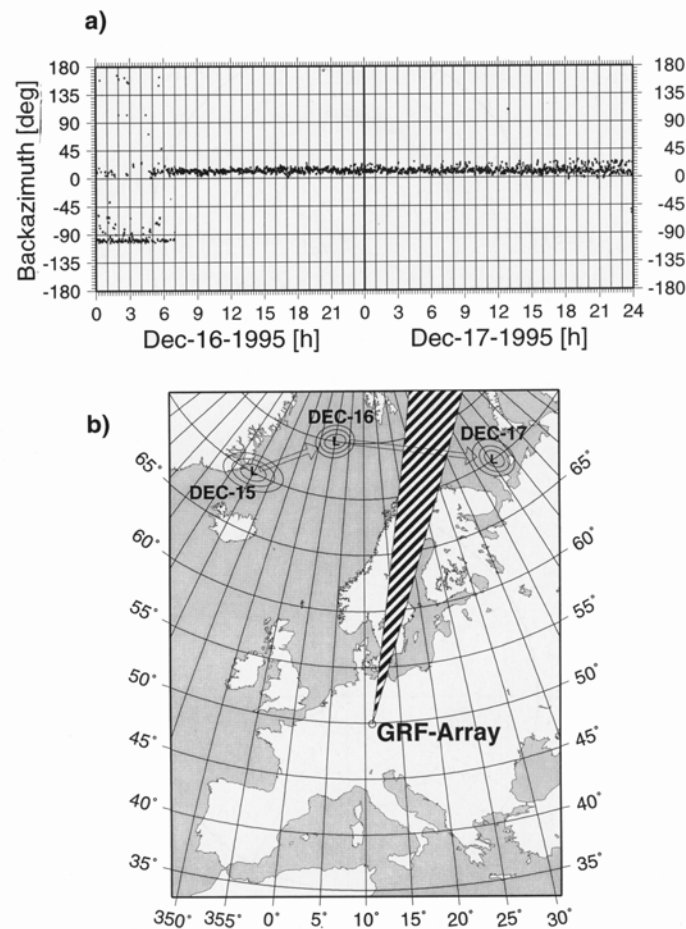
perturbations that propagate without attenuation to the ocean bottom and couple there into an elastic wave of much larger wavelength and higher phase velocity if the propagation directions are nearly opposite. The area of interference X may be off-shore where the forward propagating waves, generated by a low-pressure area L, superpose with the waves traveling back after being reflected from the coast (Fig. 4.21b). But it may also be in the deep ocean when the waves, excited earlier on the front side of the low-pressure zone, interfere later with the waves generated on the back-side of the propagating cyclone. Microseisms are mostly recorded with periods between 5 and 9 sec and are the result of energetic ocean waves between 10 and 18 sec period. Horizontal and vertical noise amplitudes of marine microseisms are similar. The particle motion is dominantly of Rayleigh-wave type, i.e., elliptical polarization of the particle motion in the vertical propagation plane. But some energy is present also as body waves, and heterogeneous Earth structure couples some energy into Love waves.



**Fig. 4.21** Schemes for the generation of a) primary and b) secondary microseisms (for explanations see text). L – cyclone low-pressure area, X – area of interference where standing waves with half the period of ocean waves develop. Copy of Figure 12 from Friedrich et al. 1998 © Springer; with kindpermission from Springer Science + Business Media.

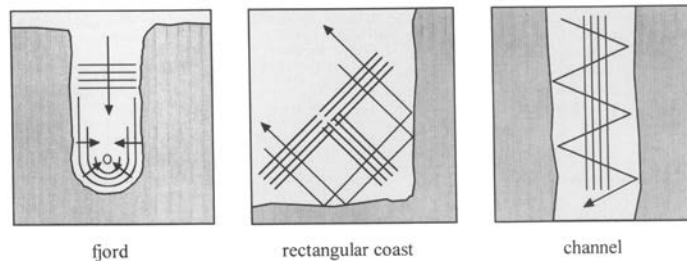
Medium-period ocean/sea microseisms experience low attenuation. They may therefore propagate hundreds of km inland. Since they are generated in relatively localized source areas, they have, when looked at from afar, despite the inherent randomness of the source process, a rather well developed coherent part, at least in the most energetic and prominent component. This allows one to locate the source areas and track their movement by means of seismic arrays (e.g., Cessaro, 1994; Friedrich et al., 1998; Fig. 4.22). This possibility has already been used decades ago by some countries, e.g., in India, for tracking approaching monsoons with seismic networks under the auspices of the Indian Meteorological Survey, although Cessaro (1994) showed that the primary and secondary microseism source locations do not follow the

storm trajectories directly. While near-shore areas may be the source of both primary and secondary microseisms, the pelagic sources of secondary microseisms meander within the synoptic region of peak storm wave activity.



**Fig. 4.22** An example of good spatial coherence of medium-period secondary ocean microseisms at a longer distance from the source area which, in this case, allows rather reliable determinations of the backazimuth of the source area by f-k analysis with seismic arrays (see Chapter 9). Figure a) above shows how the backazimuth determination changed from one day to the next, while b) shows the location of the two storm areas and the seismic array. Observations by at least two arrays permit localization and tracking of the noise-generating low-pressure areas. Copy of Figure 7 from Friedrich et al., 1998 © Springer; with kind permission of from Springer Science+Business Media.

Note that the noise peak of secondary microseisms has a shorter period when generated in shallower inland seas or lakes ( $T \approx 2$  to  $4$  s) instead of in deep oceans. Also, off-shore interference patterns largely depend on coastal geometries and the latter may allow the development of internal resonance phenomena in bays, fjords or channels (see Fig. 4.23) which affect the fine spectrum of microseisms. In fact, certain coastlines may be distinguished by unique “spectral fingerprints” of microseisms.



**Fig. 4.23** Examples for coastline geometries that provide suitable interference conditions for the generation of secondary microseisms. Copy of Figure 13 from Friedrich et al., 1998 © Springer; with kind permission of from Springer Science + Business Media.

## 4.6.2 Noise at the seabottom

### 4.6.2.1 Broadband noise

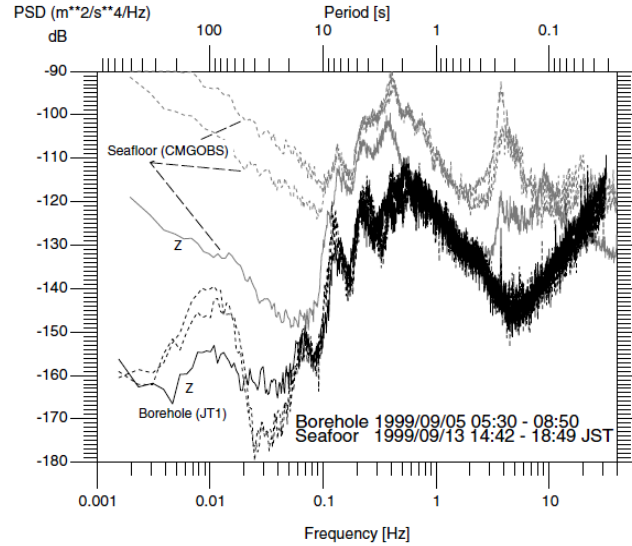
The microseism mechanism dominates sea floor noise up to frequencies of a few hertz. At slightly higher frequencies (5-10Hz), short-period noise associated with breaking waves may become important (McCreery et al., 1993). While there is little difference in the microseism band between land and the sea floor at longer periods the major difference occurs at frequencies near 1 Hz. Almost always there is much more noise energy around 1 Hz on the ocean floor than on land, often by more than 40 dB (see subsection 4.6.2.2 on short-period ocean noise below). This huge difference has a profound effect on the relative detection thresholds for short-period teleseismic arrivals in these two different recording environments, which is usually much inferior on the ocean floor. For more details on regional/local differences see Webb (2002).

In recent years more and more ocean-bottom seismographs (OBS; see main section 7.5 in Chapter 7 and Havskov and Alguacil) have been deployed in order to overcome the inhomogeneous distribution of land-based seismic stations. But permanent OBS installations are still rare. Generally, the noise level at the ocean bottom, even in deep seas, is higher than that on land (by about 10 to 30 dB) and increases with higher frequencies (e.g., Bradner and Dodds, 1964; Webb, 2002, and Fig. 7.85 in Chapter 7). It can be reduced by installing the OBS in boreholes (see Fig. 4.24). The analogues of wind and air pressure generated noise on the Earth surface are on the ocean floor the ocean currents. According to Webb (1988) they also set up a turbulent boundary layer with advecting pressure fluctuations. However, on the ocean floor the deformation signal decays already completely within a few meters depth below the sea floor because much lower velocities and hence much shorter wavelengths are associated with ocean currents than with wind.

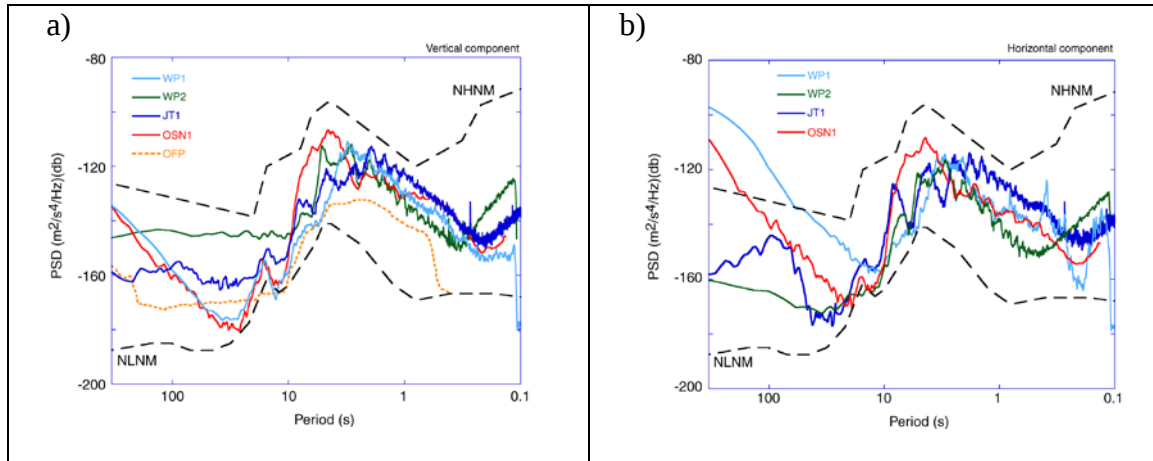
On the ocean bottom, as on land, the secondary microseism noise peak between 0.1 and 1 Hz dominates. Background noise levels in this frequency range tend to be higher in the Pacific than in the Atlantic because of its larger size and its general weather conditions (see Fig. 4.25 below and Figure 2 in Webb, 2002). While short-period body-wave arrivals around 1 Hz have been clearly recorded during calm-weather periods by OBSs in the North Atlantic, even at teleseismic distances, they are recognizable in OBS records in the Pacific only for very large events at distances of less than a few tens of degrees. On the other hand, long-period P, S and surface waves are consistently well recorded by OBSs in the noise minimum between about 0.03 and 0.08 Hz for magnitudes  $6 \pm 0.3$  even at distances  $D > 100^\circ$  (Blackman et al., 1995).



However, because of the generally inferior recording conditions broadband seismic observations from the sea floor remain rare. The primary motivation for installing stations on the sea floor has been to improve the global ray path coverage for tomographic studies of the Earth and for regional studies of seismicity and structure. Yet, significant improvements in global detection thresholds will only be obtained near the sea-floor stations itself. For some significant results see section 7.5.3 of Chapter 7.



**Fig. 4.24** Difference of typical noise power spectra in a borehole installation and a seafloor installation near the borehole of site JT1 (see Fig. 7.80 in Chapter 7). The seafloor observation is noisier than that in the borehole and the horizontal components are much noisier than the vertical (Z) components (after Araki et al., 2004; Figure copied from Fig. 7.71 in Chapter 7; with permission of M. Shinohara, 2013).



**Fig. 4.25** Comparison of ambient seismic noises from seafloor borehole installations: a) for vertical component; b) for horizontal component. WP-1, WP-2 and JT-1 denote observatories in the western Pacific, OSN-1 an observatory off Hawaii (Stephen et al., 1999), and OFP an observatory in the North Atlantic Ocean (Montagner et al., 1994). NHNM and NLNM indicate the Peterson (1993) New High Noise Model and New Low Noise Model, respectively. (Figure copied from Fig. 7.72 in Chapter 7; with kind permission of M. Shinohara, 2013).

According to Webb (1998), horizontal component noise levels on sea-floor sensors can be as much as 60 dB higher than vertical component noise levels at periods near 100 s due to tilt

(e.g., site WP1 in Fig. 4.25). Tilt noise will appear on vertical component records as well if the vertical component sensor is not very precisely oriented with the true plume line. Most vertical component records from the sea floor show evidence of tilt noise due to currents which can be rather strong and thus the records very noisy on shallow-water sites (Romanowicz et al., 1998). However, because of the short wavelengths associated with pressure fluctuations due to ocean-floor currents burial of the sensors just a few meters below the ground will already strongly reduce (up to about 60 dB) the long-period horizontal component noise levels (Webb, 1998). But incomplete burial allowing for convection of sea water in the borehole may deteriorate this positive effect (Beauduin and Montagner, 1996; Vernon et al., 1998). Webb and Crawford (1999) show that sea floor pressure data can be used to remove most of the infragravity wave deformation signal from vertical component OBS data in the band from 0.001 to 0.04 Hz. Improvements of 10-20 dB in SNR for long-period surface waves could thus be achieved at sea-floor stations in the Pacific. For more information about noise and signal recording conditions and installation requirements in an ocean bottom environment see main section 7.5 in Chapter 7 and Webb (2002).

#### **4.6.2.2 Short-period noise (0.5-50 Hz)**

Webb (2002) gives a rather detailed overview of the mechanisms which generate short-period noise on the sea floor and how this noise decays with depth below the sea floor. Here we summarize only some of the most interesting facts and give related references for further reading.

The sea floor in the band from 0.5 to 5 Hz is almost invariably noisier than the typical land site because of microseisms generated by small-amplitude, high-frequency ocean waves that can be produced already by light winds. Such locally generated microseisms maintain for most of the time sea-floor noise levels some 40 dB above land sites. These noise levels "saturate" with little increase above a frequency-dependent saturation wind velocity, which is near 1 Hz about 5 m/s (McCreery et al., 1993). During long intervals of calm winds, however, these noise levels may fall by 20 dB or more (Webb, 1998; Wilcock et al., 1999). Because wind speeds exceed 5 m/s over most sites in the oceans for most of the time, the average sea-floor sites are usually rather noisy near 1 Hz. A notable exception is the Arctic Ocean during winter when the ice cap prevents the excitation of the ocean waves on the surface. Parts of the North Atlantic and Indian oceans can also experience long intervals of near calm conditions during the summer months and may provide quiet sites for detection of short-period body waves (Webb, 1998). In contrast, most sites in the Pacific seem to be noisy most of the time (McCreery et al., 1993).

Installing seismometers within sea-floor boreholes will not necessarily reduce this microseism noise level because the energy is coupled from ocean waves into acoustic waves and then into Rayleigh-wave modes below the sea bed which penetrate down to depths of several kilometres. This is beyond the present range of ocean drilling. However, at shallow depths some SNR improvement is expected because microseisms are scattered by bathymetry and topography at the sediment-basement interface producing short-wavelengths Stoneley shear modes. Their amplitude decay below the sea floor rapidly with frequency and depth (Bradley et al., 1997). But signal levels at depth are also smaller (because of changing impedance). Therefore the resulting improvement in signal-to-noise ratio is near zero at 0.2 Hz, but it increases to more than 10 dB at 5 Hz. Correlation lengths for this type of ocean bottom noise at frequencies above 0.4 Hz, measured between elements of a small-aperture array of OBS, were between 100 and 200 m only (Schreiner and Dorman, 1990).

More high-frequency noise (5-50Hz) on the sea floor tends to be associated with man-made (cultural) sources. A good review of "low-frequency acoustic noise" ( $< 100\text{Hz}$ ), e.g., radiated by marine mammals, can be found in Richardson et al. (1995). While noise frequencies between 5-10 Hz are dominantly wind-generated and caused by wave breaking, most of the cultural ocean noise is more high-frequency and caused by ship traffic (cavitation of propellers and noise from vibration of machinery). Distinct lines associated with generators (50 or 60Hz) or with other machinery (motors, generators, thrusters and pumps, etc.) also make up much of the shipping noise, with noise levels peaking between 10 and 100 Hz. The lowest-frequency components are associated with the rotation of propellers. Large ships such as supertankers produce sharp spectral lines at frequencies of 7 Hz or lower but with many higher harmonics (Richardson et al., 1995). Areas near high-traffic ship routes can be 20-25 dB noisier than regions remote from shipping. Other sources of cultural noise observed on the ocean floor in the seismic band include active seismic sources (airguns) and offshore drilling (Richardson et al., 1995).

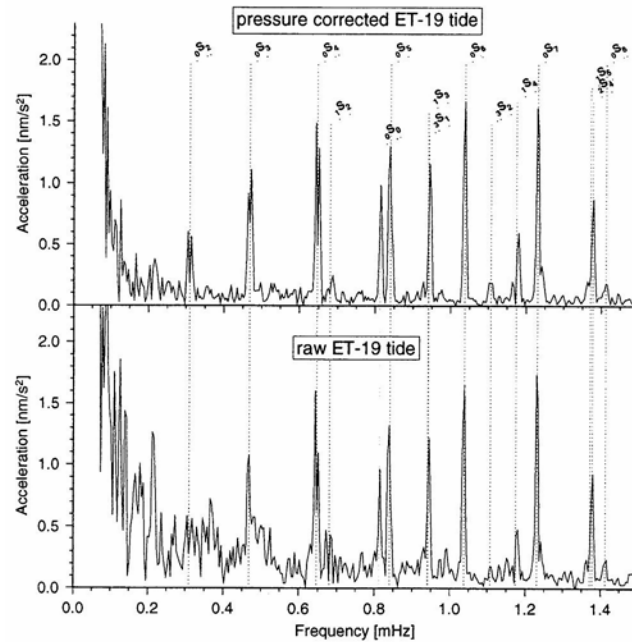
For some locations large whales are another natural noise source. Blue and fin whales produce very loud calls at frequencies as low as 16 Hz. They are easily seen locally on sea-floor seismometer records. In some special locations whale calls can be so frequent as to interfere with active source seismic experiments. Also the movement of ice in the polar regions may be a natural noise source in Arctic and Antarctic waters (Richardson et al., 1995).

Most favourable for high-frequency seismic observations is the transition band (5-10Hz), above the microseisms and below the typical shipping noise. It provides relatively quiet conditions that allow the detection of oceanic Pn and Sn phases (also known as Po and So) from moderate magnitude oceanic events to great distances (Butler et al., 1987).

### **4.6.3 Seismic noise on land**

#### **4.6.3.1 Long-period noise**

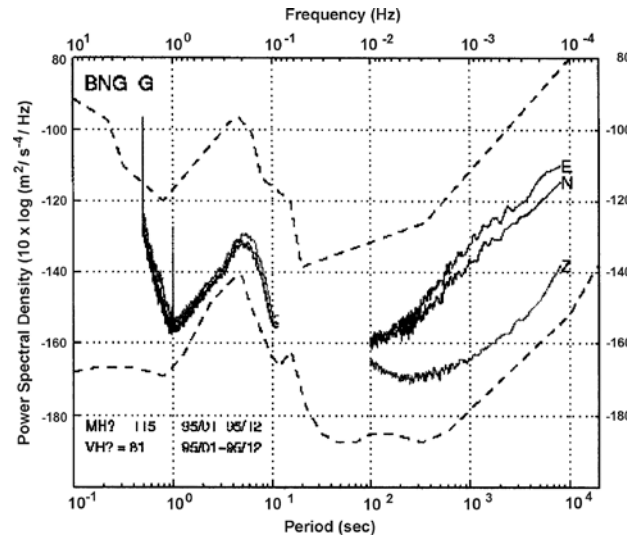
Ground noise on land at very long period (0.2 to 50 mHz) is usually associated with atmospheric pressure fluctuations. At these very low frequencies, vertical component seismometers react to changes in gravity as the mass of the atmosphere above a site changes with atmospheric pressure. At higher frequencies, the ground deforms under the fluctuating pressures associated with turbulence in the boundary layer of the atmosphere (Murphy and Savino, 1975). If seismometers are not fully sealed they may also experience apparent accelerations due to buoyancy effects on the seismometer mass, and pressure fluctuations may cause long-period horizontal component noise by distorting the case of the seismometer or the walls of the vault (Savino et al., 1972). Long-period seismometers are also sensitive to temperature changes. Therefore, careful installations aim at sealing the sensor from both pressure and temperature fluctuations (see IS 5.4). Zürn and Widmer (1995) showed that vertical component noise levels between 0.2 and 1.7 mHz can be reduced by more than 10 dB by subtracting the scaled, locally recorded pressure signal from the vertical acceleration record, thereby improving the SNR for normal mode observations (see also Beauduin et al. 1996 and Fig. 4.26).



**Fig. 4.26** Linear acceleration amplitude spectra of LaCoste-Romberg Earth tide gravimeter ET-19 records at the BFO observatory in SW Germany after the big Shikotan earthquake of October 4, 1994. Bottom: raw data; Top: barometric-pressure corrected data. Copy of Figure 1 in Zürn and Widmer (1995): On noise reduction in vertical seismic records below 2 mHz using barometric pressure. *Geophys. Res. Lett.*, 22(24), 3537-3540; © American Geophysical Union.

Turbulence in the atmospheric boundary layer produces pressure fluctuations at the Earth's surface that cause significant deformation to depths of a few tens of meters below the surface. This noise source is a primary reason why broadband seismometers at permanent stations are often installed beneath the Earth's surface in shallow boreholes (see section 7.4.5 and Fig. 7.59 in Chapter 7). For instruments on the surface, short-wavelength atmospheric pressure fluctuations produce a large tilt signal that makes the horizontal components particularly noisy. Horizontal noise power may then become significantly larger than vertical noise power. Tilt couples gravity into the horizontal components but much less into the vertical component (see Chapter 5, section 5.3.3 and Figs. 5.8 and 5.9). The ratio increases with the period and may reach a factor of up to 300 (about 50 dB). Therefore a site can be considered as still favorable when the horizontal noise at 100 to 300 s is still within 20 dB above the vertical component noise power, as in Fig. 4.27. Horizontal component noise is mainly due to local tilt, which may also be caused by traffic and wind pressure. At very long periods, deformation of the ground, the seismometer vault, or the building due to insolation gradients (uneven heating by the sunshine) may become noticeable. Generally, recording sites poorly coupled to the ground or on poorly consolidated ground will tend to be noisy, both at long and short periods. Other reasons for increased long-period noise may be air circulation in the seismometer vault or underneath the sensor cover. Special care in seismometer installation and shielding is therefore required in order to reduce drifts and long-period environmental noise (see Chapters 5 and 7 as well as IS 5.4).

For more details also on the causes and phenomena of long-period noise see Webb (2002).



**Fig. 4.27** Seismic noise at the station BNG (Bangui, Central Africa) as compared to the new global seismic noise model by Peterson (1963) (from the FDSN Station Book). The upper and lower broken curves denote the NHNM and NLNM noise models according to Peterson (1993).

#### 4.6.3.2 Short-period noise (0.5 to 50 Hz)

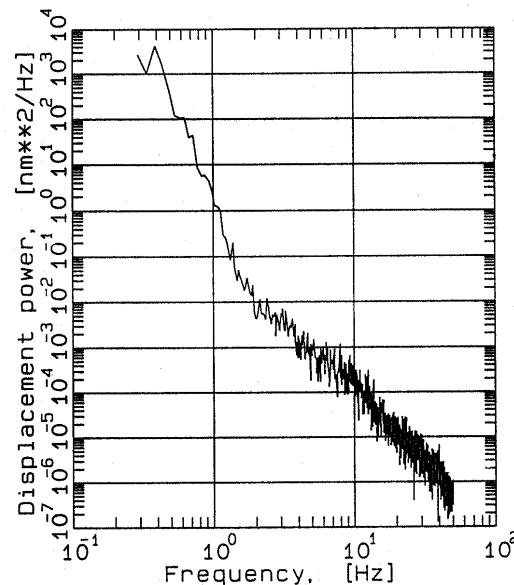
Short-period seismic noise may have natural causes such as wind (wind friction over rough terrain; trees and other vegetation or built-up objects swinging or vibrating in the wind), rushing waters (waterfalls or rapids in rivers and creeks) etc. Wind-generated noise is broadband, ranging from about 0.5 Hz up to about 15 to 60 Hz (Young et al., 1996). But the dominant sources of high-frequency noise are man-made (power plants, factories, rotating or hammering machinery, road and rail traffic, etc.; see sections 7.1 and 7.2 in Chapter 7.) and often referred to as *cultural noise*. Wind-generated noise couples mostly into surface-wave modes, cultural noise, however, couples at least partly into body waves that can propagate also to greater depth (Carter et al., 1991). Cultural noise is in principle avoidable, although it is often impractical to place seismic station sites at sufficient distances away from cities or highways. Some rules of thumb suggest that it may be necessary to site short-period recording stations as far as 25 km from power plants or rock-crushing machinery, 15 km from railways, 6 km from highways, and a kilometer or more from smaller roads. And with respect to moving waters it is recommended to place stations away from moving water between some 60 km for very large waterfalls and dams to 15 km for smaller rapids (see Willmore, 1979 and Tab. 7.5 in Chapter 7). However, noise travels further across competent rock than through alluvial filled valleys. Therefore, "safe" distances may be from 1/2 to 2/3 of the values above, depending on propagation path (Willmore, 1979).

Most of the short-period natural and man-made sources are distributed, stationary or moving. Their contributions, coming from various directions, superpose to a rather complex, more or less stationary random noise field. The particle motion of short-period noise is therefore more erratic than for long-period ocean noise. Nevertheless, polarization analysis, averaged over moving time-windows, sometimes reveals preferred azimuths of the main axis of horizontal particle motion hinting at localized noise sources. Also the vertical component is clearly developed and averaged particle motion in 3-component records indicates fundamental Rayleigh-wave type polarization. A rather popular and cost-effective microzonation method is based on this assumption. It derives information about the fundamental resonant frequency of

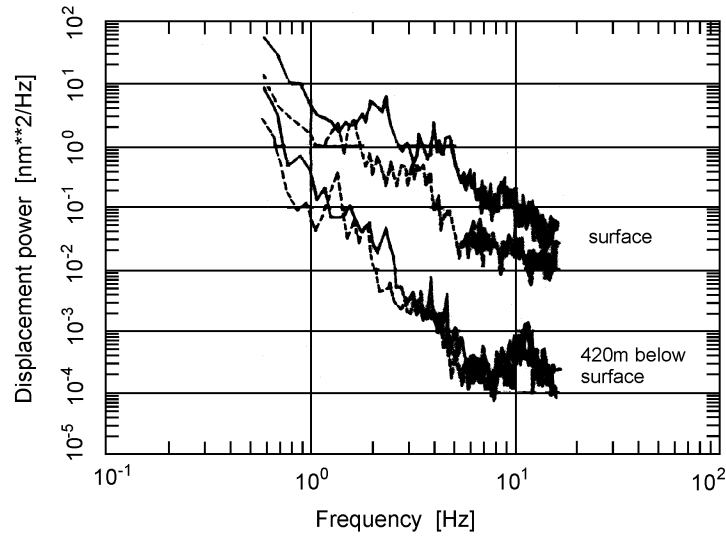
the soft-soil cover and estimates local site amplification of ground motion from the peak in the horizontal to vertical component spectral noise ratio (H/V *Nakamura method*; e.g., Nakamura, 1989; Bard, 1999; Parolai et al., 2001 and 2002; Chapter 14).

According to the above, every recording site has different noise characteristics depending on its distance from the ocean, industrial and settlement areas, major infrastructure facilities, the wind climate of the site and the depth of burial of the sensor. At some locations a seasonal cycle is evident due to variations in wind, or water flow in nearby rivers. E.g., according to Fyen (1990), spring runoff raises noise levels by up to 15 dB between 0.5 and 15 Hz at the NORESS array. In settlement and industrial areas a diurnal cycle may dominate due to variable human activity during the day.

High-frequency noise spectra on land during quiet intervals resemble each other at quieter sites. The rather featureless displacement power spectrum follows a power-law dependence in frequency, proportional to  $\approx f^{-2}$  (see Fig. 4.28 and Bungum et al., 1985). This spectral appearance changes drastically due to the strong impact of cultural noise, especially in the frequency band between 1 and 10 Hz (see Fig. 4.31). Wind noise depends on the strength of the wind and the character of the site. Recorded noise power near 1 Hz in a surface vault may increase by 10-20 dB, as compared to records on calm days, when wind speed reaches about 5 m/s. According to Withers et al. (1996) short-period wind noise becomes detectable in records at surface sites far from cultural sources at wind velocities above about 3 m/s and at 4 m/s for subsurface sensors as well (see Fig. 4.29). The primary mechanism by which wind couples into high-frequency noise is probably by the direct action of the wind on trees, bushes, and other structures, although at lower frequency the pressure fluctuations associated with wind can directly drive motions of the ground.



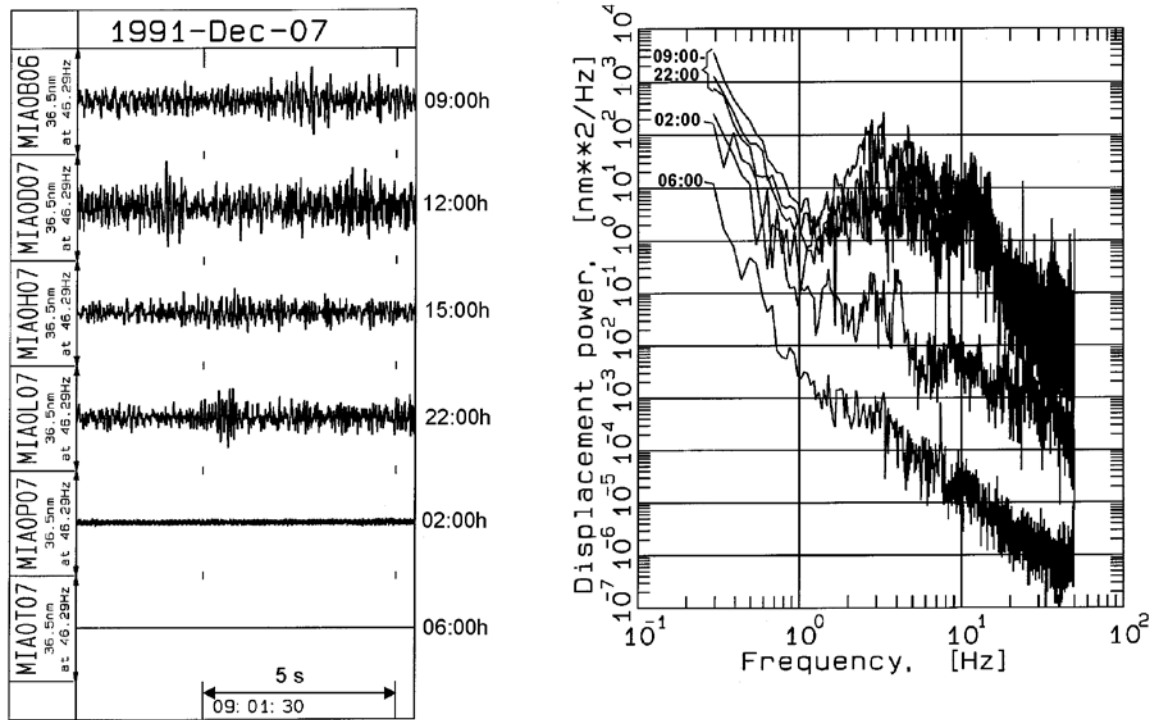
**Fig. 4.28** Spectrum of displacement power spectral density calculated from 6 moving, 50% overlapping intervals of short-period noise records, 4096 samples long each, i.e., from a total record length of about 80 s at a rather quiet site in NW Iran.



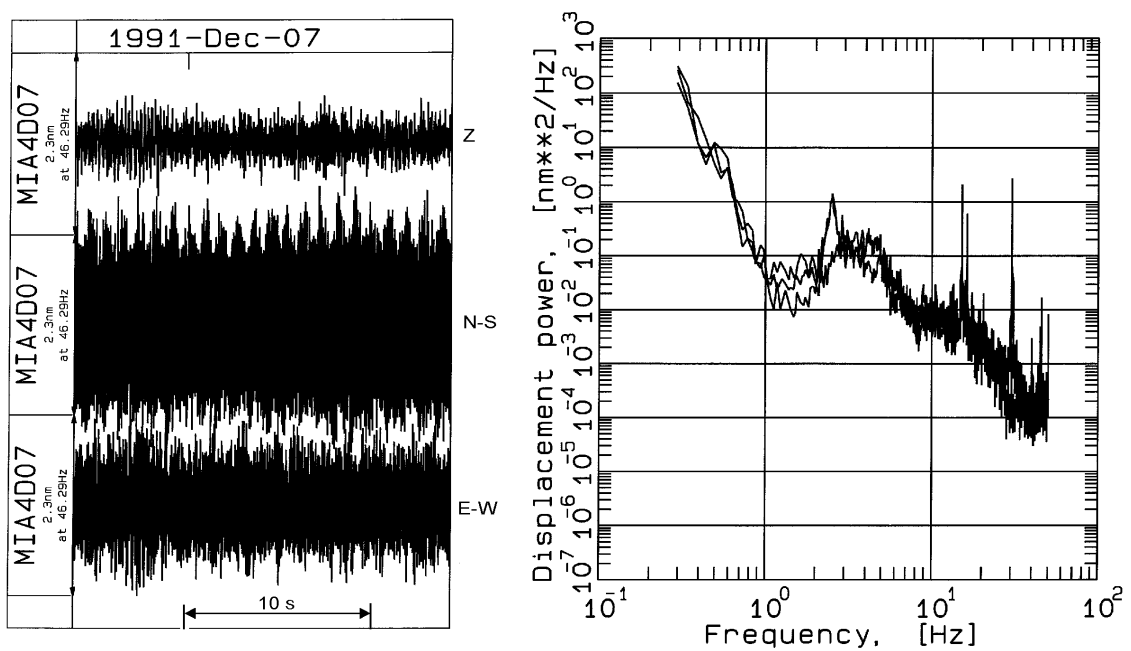
**Fig. 4.29** Displacement power noise spectra measured at the surface (upper curves) and at 420 m below the surface in a disused salt mine at Morsleben, Germany (lower curves) on a very quiet day (hatched lines) and on a day with light wind on the surface (wind speed about 4 m/s; full lines).

Cultural noise varies greatly between sites (Fig. 4.34). Noise levels may differ also by 30 dB or even more within a day (e.g., Fig. 4.30 and Fyen, 1990), as well as from day to day. Therefore, it is common to present ensembles of spectra by per-centile ranking to allow discrimination between infrequent very loud sources and more typical background levels (e.g., Bungum et al., 1985). Industrial and traffic noise levels are usually lowest on weekends but increase during holidays because of higher traffic. Sources such as power plants, transformer stations etc. generate narrowband noise, producing energetic spectral lines at 50 or 60 Hz as well as related harmonics and sub-harmonics (e.g., at 30 Hz, 25 Hz, 12.5 Hz) depending on location (Fig. 4.31). The NORESS array data included lines at 6, 12, and 17 Hz due to reciprocating saws in a nearby sawmill (Fyen, 1990), and records at Moxa (MOX) station in Germany have been spoiled over decades by a pronounced 2 Hz peak due to heavy steam engine driving a steel-press for rails at a distance of 16 km. It may be possible to remove some of such narrowband noise through application of digital or analog filters (see section 4.7). Other types of machinery, however, may also produce narrowband peaks in the noise spectra, but with time-variable frequencies as motor speed may vary. These noise sources are more difficult to remove.

Comparisons of noise spectra from surface vaults and subsurface installations in mines (Fig. 4.29) or boreholes (Fig. 4.32) have repeatedly shown significant improvements in SNR for high-frequency phases recorded at depths, sometimes even as shallow as a few meters (e.g., Aster and Shearer, 1991). The signal-to-noise ratio for high-frequency arrivals can be optimized by choosing sites with a lossy blanket of weathered near-surface material (Withers et al., 1996). Large reductions in noise level may also be seen in the first 5 m between the weathered rock of the surface and the more competent rock in the borehole. Wind noise, however, may couple into open boreholes. Therefore, covering open boreholes can reduce wind noise considerably.

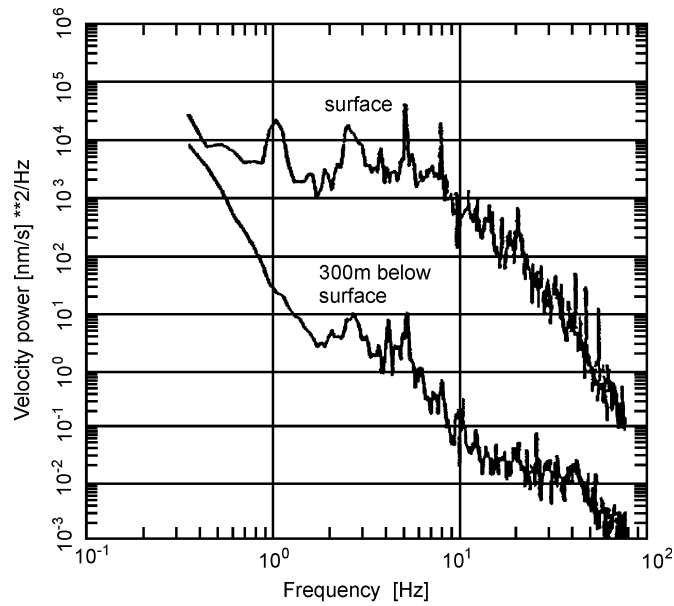


**Fig. 4.30** Comparison of relatively quiet sections of vertical component noise records (left; without strong transients) and related power spectra (right) at a reference site in the district capital town of Miene in NW Iran. The measurements were made at different times of the day.



**Fig. 4.31** Noise records and related power spectra near to a transformer house and power line. Note the monochromatic spectral lines around 13, 30 and 50-60 Hz, either induced by the AC current frequency and its lower harmonics and/or caused by the vibration of the transformer.

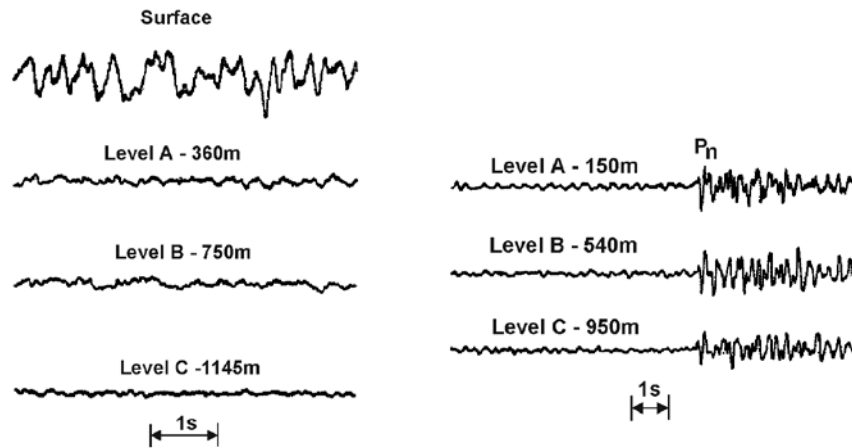




**Fig. 4.32** Velocity power density spectra as obtained for noise records at the surface (top) and at 300 m depth in a borehole (below) near Gorleben, Germany (courtesy of M. Henger).

Most comparisons of borehole and vault noise levels show 10-30 dB lower noise levels in the boreholes, with most of the reduction in noise level occurring in the first 100 m (Young et al., 1996). Broadband spectral levels fall also with increasing borehole depth but narrow spectral lines due to cultural sources and or waveguide trapping may show less depth dependence (e.g., Galperin et al., 1978). This is also demonstrated by the very different noise decay with depth in typical 1-Hz records made in two deep boreholes in the US (Fig. 4.40). While in a borehole in Texas the noise amplitudes decreased steadily (up to a factor of 30) down to 3000 m depth below surface, they decreased in another borehole in Oklahoma down to about 2000 m only and then increased again towards larger depth. At this greater depth a layer with 22% lower P-wave velocity was found by means of borehole seismic measurements (noise traveling in a low-velocity layer?).

The surface-wave nature of seismic noise (including ocean noise) is the reason for the exponential decay of noise amplitudes with depth, which is not the case for body waves (Fig. 4.33). Since the penetration depth of surface waves increases with wavelength, high frequency noise attenuates more rapidly with depth. In case of Fig. 4.33 the noise power at 300 m depth in a borehole was reduced, as compared to the surface, by about 10 dB, at  $f = 0.5$  Hz, 20 dB at 1 Hz and 35 dB at 10 Hz. Withers et al. (1996) found that for frequencies between 10 to 20 Hz, the SNR could be improved between 10 to 20 dB and for frequencies between 23 and 55 Hz as much as 20 to 40 dB by deploying a short-period sensor at only 43 m below the surface. But both noise reduction and signal behavior with depth depend also on local geological conditions (see 4.6.5). Because of the surface-wave character of short- and medium-period noise, the horizontal propagation velocity of seismic noise is frequency dependent. It is close to the shear-wave velocity in the uppermost crustal layers, which is about 2.5 to 3.5 km/s for outcropping hard rock and about 300 to 650 m/s for unconsolidated sedimentary cover. This is rather different from the apparent horizontal propagation velocity of P waves and all other steeply emerging teleseismic body-wave onsets.



**Fig. 4.33** Recording of short-period seismic noise (left) and signals (right) at the surface and at different depth levels of a borehole seismic array. Note the significantly larger Pn amplitude closer to the surface. It is due to the surface amplification effect which may reach a factor of two for vertical incident waves. (Figure copied from Bormann, 1966; modified from Broding et al., 1964).

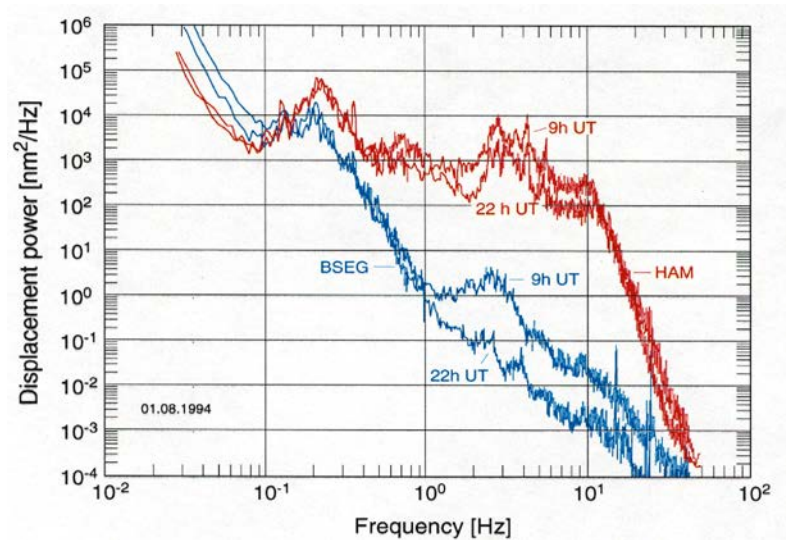
Signals which have a similar waveform and polarization, so that they can interfere constructively, are termed coherent. This is usually the case for seismic signals generated and radiated by a common source process. The degree of coherence is defined by the ratio between the auto- and the cross-correlation of the time series (the cross-correlation is squared and divided by both autocorrelations, to render the result independent of the signal amplitudes). It may vary between 0 and 1. For seismic noise it shows distinct frequency dependence. Spatial coherence (paragraph 4.2.4) may be rather high for long-period ocean microseisms. Accordingly, the correlation radius, i.e., the longest distance between two seismographs for which the noise recorded in certain spectral ranges is still correlated, increases with the noise period. It may be several km for  $f < 1$  Hz but drops to just a few tens of meters or even less for  $f > 50$  Hz. For seismic noise, it is usually not larger than a few wavelengths.

Generally, noise levels increase with higher wind speeds. However, there is apparently no linear relationship between wind speed and the amplitude of wind generated noise. Wind noise increases dramatically at wind speeds greater than 3 to 4 m/s and may reach down to several hundred meters depth below the surface at wind speeds  $> 8$  m/s (Young, 1996). But generally, the level and variability of wind noise is much higher at or near the surface and is reduced significantly with depth (Fig. 4.30). Moreover, while for wind speeds below 3 to 4 m/s, one may observe omnidirectional background noise with relatively large coherence length at frequencies below 15 Hz, the coherence length is strongly reduced at higher wind speeds with increased air turbulence and local wind pressure fluctuations (Withers et al., 1996).

Thus, differences in the frequency spectrum, horizontal wave-propagation velocity, degree of coherence and depth dependence between (short- and medium-period) seismic noise and seismic waves allows to improve the signal-to-noise-ratio (SNR) by installing seismic sensors either at reasonable depth below the surface or by way of data processing (see 4.7).

#### 4.6.3.3 Seismic noise in urban environments

In urban, densely populated and industrialized environment both natural and cultural noise components may play a major role in a wide range of frequencies. Measuring and analyzing urban seismic noise (USN) requires broadband recordings and data analysis. Fig. 4.34 compares the noise power density spectrum measured at the former seismological station in the city of Hamburg, Germany, with the spectrum measured on a hard rock site in a quiet Spa village some 70 km apart.

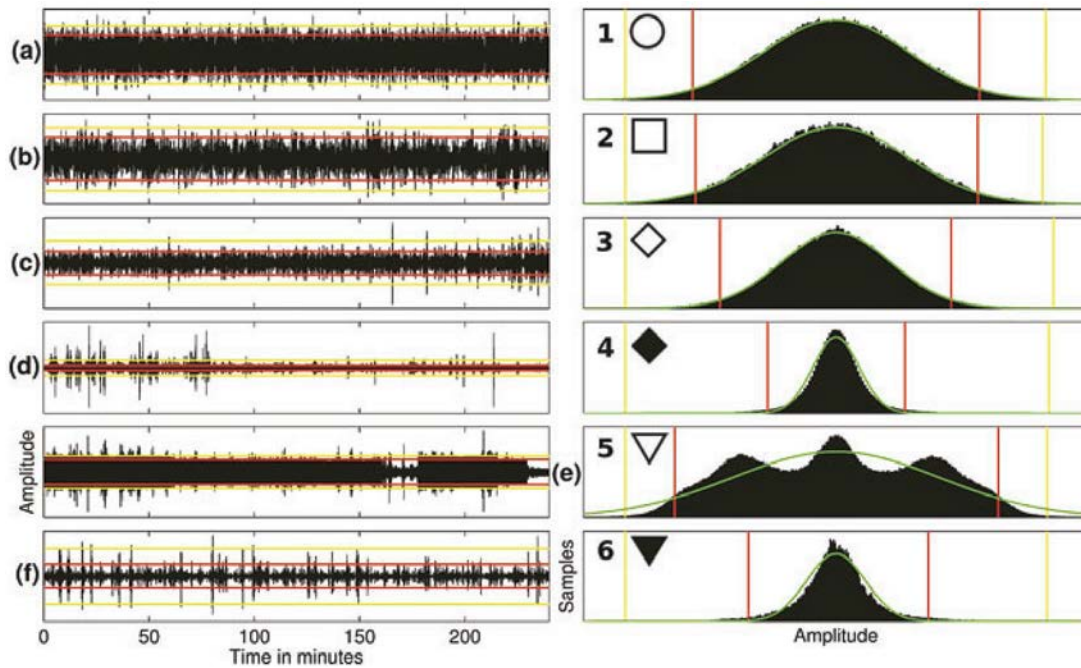


**Fig. 4.34** Noise power spectra at HAM (upper two curves, red) and BSEG (lower two curves, blue) determined from vertical-component records on August 1, 1994, at daytime (around 10 a.m. local time) and nighttime (around 11 p.m. local time), respectively. (Color version of figure 9 in Bormann et al., 1997 © Springer; with kind permission of Springer Science + Business Media).

From Fig. 4.34 the following conclusions can be drawn: While in a remote quiet urban setting the seismic noise spectrum may decay at night almost monotonously towards higher frequencies the cultural noise activity may increase the daytime PSD for frequencies above 1 Hz by some 10 dB. In contrast, in a very busy industrialized city such as Hamburg, the seismic noise level may be 20 dB to 40 dB higher for frequencies of about 0.5 – 30 Hz and may be reduced during night time by some 2 dB to 10 dB only between 1 Hz and 10 Hz. In the metropolitan city of Berlin the “cultural noise” difference at frequencies exceeding 1 Hz between midday and midnight is nearly constant at 10 dB for all frequencies (see Fig. 7.31 in Chapter 7), but it may be larger at other towns with more pronounced diurnal variability of human and industrial activities.

Groos and Ritter (2009) stated that the generally high variability of USN does not allow to characterize a sample time series of USN comprehensively by single measures such as the standard deviation of a seismic noise series or the PSD at a given frequency. For the metropolitan city of Bucharest, Romania, they calculated long-term spectrograms of up to 28 days duration from broadband seismic recordings. The aim was to identify the frequency-dependent behavior of the time-variable processes that contribute to USN. Based on the spectral analysis of the data in eight frequency ranges between 8 mHz and 45 Hz they proposed a time-domain classification which allows to identify both Gaussian distributed seismic noise time series as well as time series that are dominated by transient or periodic signals due to traffic,

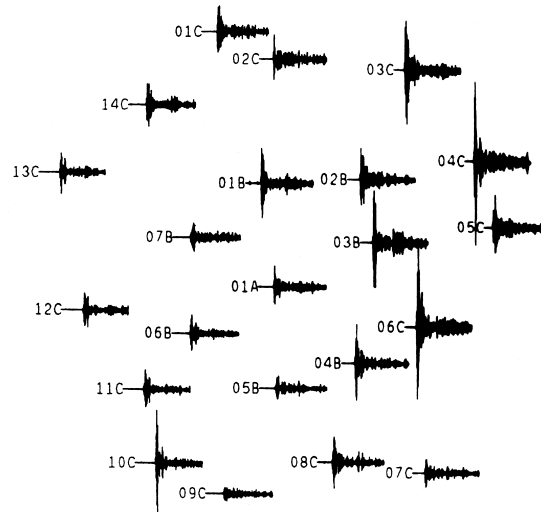
rotating machinery etc. Groos and Ritter (2009) found that only 40% of the analyzed time series are characterized by Gaussian distributed noise and that the most common deviations from a Gaussian distribution are due to large-amplitude transient signals (Fig. 4.35). And most astonishingly, they found significant variations of the statistical noise properties with daytime in the whole frequency range between 0.04 and 45 Hz, i.e., both below and above the microseismic noise peak. This result points to a broadband human influence on USN. The authors recommend to use such automatically derived information from broadband seismic data to select suitable time windows for H/V noise studies (see Chapter 14) or ambient noise tomography.



**Fig. 4.35** Time series (left-hand panels) and their histograms with the Gaussian distributions (right-hand panels) of vertical-component USN recorded in Bucharest according to Groos and Ritter (2009). Estimated from the mean and the upper boundary of the 68% interval of the corresponding time series. The red and yellow lines in the right-hand panels delimit the 2-sigma and 3-sigma range of the Gaussian distributed time series. Note that the more the time series are dominated by short transient noise signals the more the real amplitude frequency distributions deviate from the respective Gaussian ones. The non-bell shaped multimodal distribution under (e) is due to time series dominated by sinusoidal signals of different dominating frequencies and the asymmetric distribution under (f) is due to the dominance of asymmetric signals in the time series. (Copy of Figure 5 on p. 1220 in Groos and Ritter (2009); © Geophys. J. Int.)

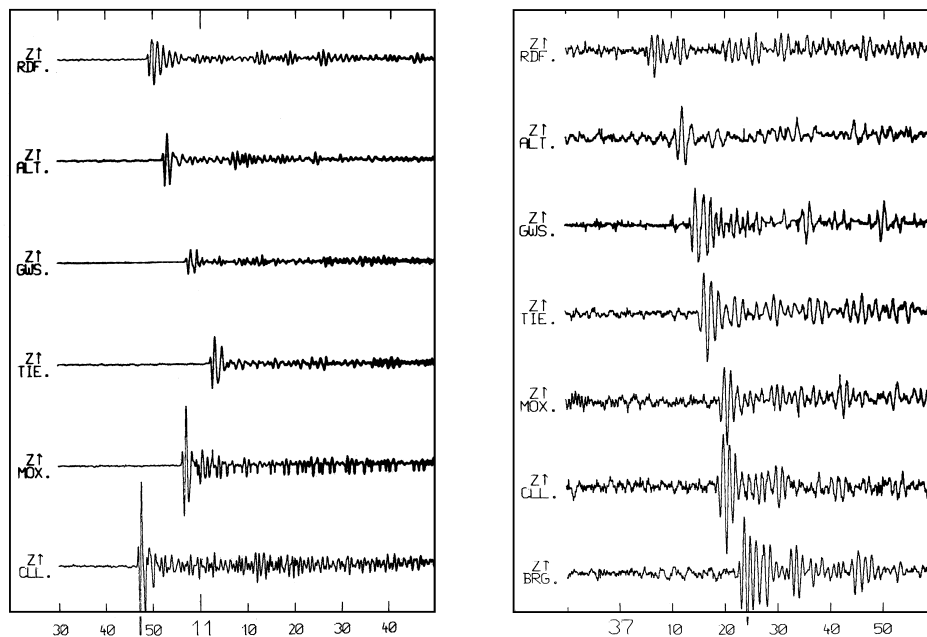
#### 4.6.4 Signal and SNR variations due to local site conditions

Compared to hard rock sites, both noise and signals may be amplified on soft soil cover. This signal amplification may partly or even fully outweigh the higher noise observed on such sites. Signal strength observed for a given event may vary strongly (up to a factor of about 10 to 30) within a given array or station network, even if its aperture is much smaller than the epicentral distance to the event ( $< 10\text{-}20\%$ ), so that differences in backazimuth and amplitude-distance relationship are negligible (Fig. 4.36).



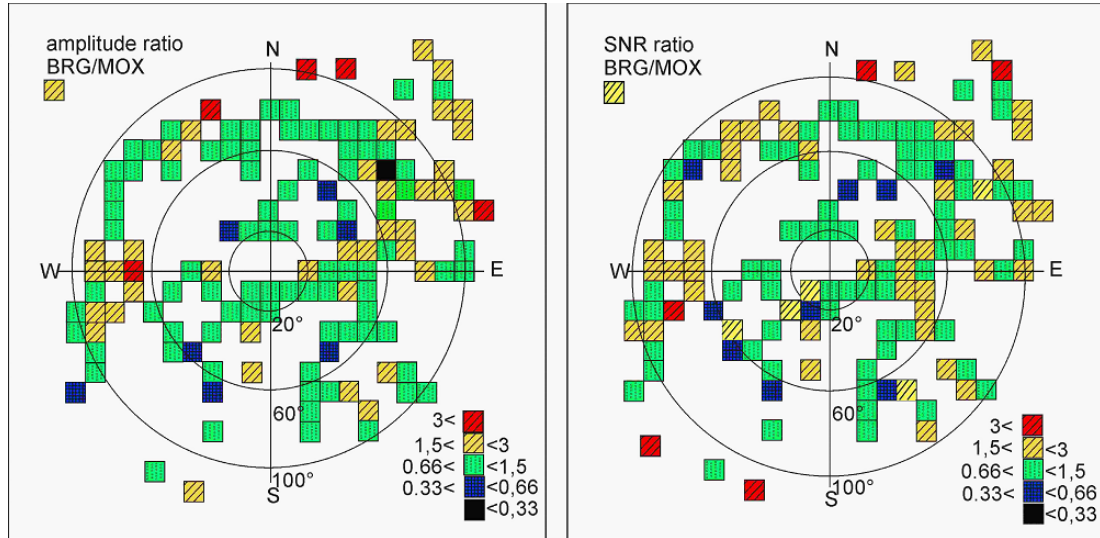
**Fig. 4.36** Records of a Semipalatinsk event at stations of the NORSAR seismic array (diameter about 90 km). The event is about  $37\text{--}38^\circ$  away. Note the remarkable variations of signal amplitudes by a factor up to 10 (the standard deviation is about a factor of 2) (from Tronrud, 1983a).

Also, while one station of a network may record events rather weakly from a certain source area, the same station may do as well as other stations (or even better) for events from another region, azimuth or distance (e.g., station GWS in Fig. 4.37 left and right, respectively).



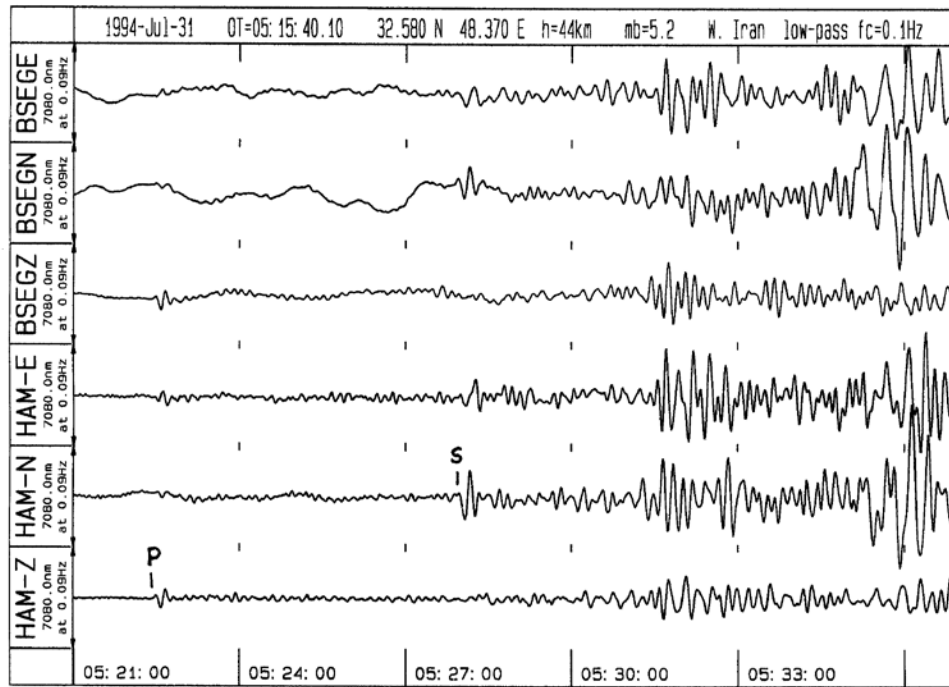
**Fig. 4.37** Short-period records of underground nuclear explosions at the test sites of Semipalatinsk (left,  $D$  about  $41^\circ \pm 1^\circ$ ) and Nevada (right;  $D$  about  $81^\circ \pm 1^\circ$ ) at stations of the former East German seismic network. Note the differences in signal amplitudes both amongst the stations for a given event and for the same station pairs, when comparing events in different backazimuth and distance. Also, at right, the compressive first motion is lost at several stations in the presence of noise due to the narrowband one-octave recording. Small numbers on the x-axis are seconds, while big numbers are minutes.

Fig. 4.38 compares for regional and teleseismic events the short-period P-wave amplitude ratio (left) and SNR (right) of two stations of the German Regional Seismic Network (GRSN). In the same azimuth range, but at different epicentral distances, station BRG may record both  $> 3$  times larger as well as  $> 3$  times smaller amplitudes than station MOX. This corresponds to magnitude differences up to one unit! The SNR ratio BGR/MOX also varies by a factor of 3 and more, depending on azimuth and distance of events. Therefore, optimal site selection can not be made only on the basis of noise measurements. Also, the signal conditions at possible alternative sites should be compared.



**Fig. 4.38** Pattern of the relative short-period P-wave amplitudes (left) and related SNR (right) at station BRG normalized to those of station MOX (170 km apart) in a distance-azimuth polar diagram. Modified version of Figure 7 from P. Bormann et al., 1992 © Elsevier; with permission from Elsevier.

However, differences in local signal conditions may become negligible in long-period recordings and thus play a lesser role in site selection for broadband networks and arrays. This is demonstrated with Fig. 4.39. Plotted are the low-pass ( $f_c = 0.1$  Hz) filtered 3-component traces of an earthquake in Iran, recorded at the stations HAM and BSEG. According to Fig. 4.34 the short-period noise at BSEG was about 20 to 40 dB lower than at station HAM, yet, the long-period SNR in the Z-component of HAM was comparably good and even better in the horizontal components because of better tilt noise avoidance than in the temporary installation at BSEG at the time of the records. Thermal shielding and reduction of tilt noise have the highest priority in long-period and very broadband installations (see IS. 5.4).

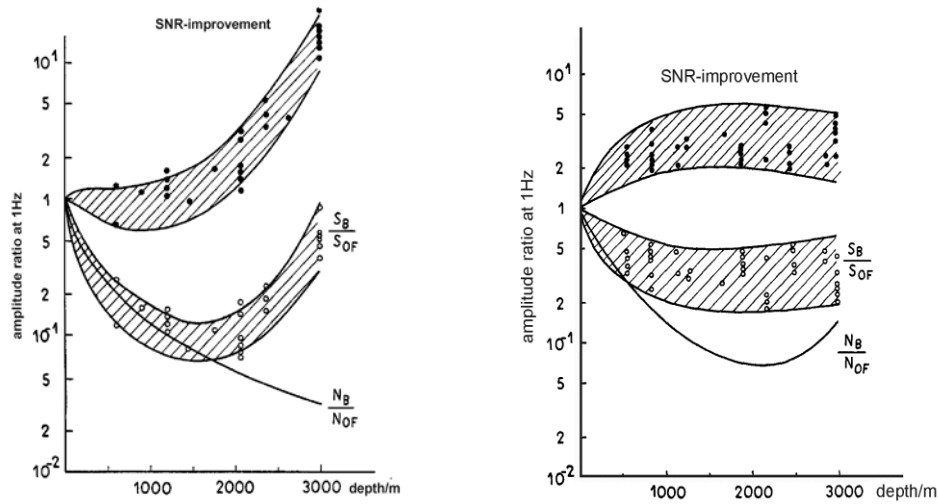


**Fig. 4.39** Low-pass filtered ( $f_c = 0.1$  Hz) long-period 3-component records at the seismic stations BSEG and HAM of the Iran earthquake. For the broadband noise spectra of these two station see Fig. 4.34. Copy of Figure 20 from Bormann et al., 1997 © Springer; with kind permission from Springer Science+Business Media.

#### 4.6.5 Installations in subsurface mines, tunnels and boreholes

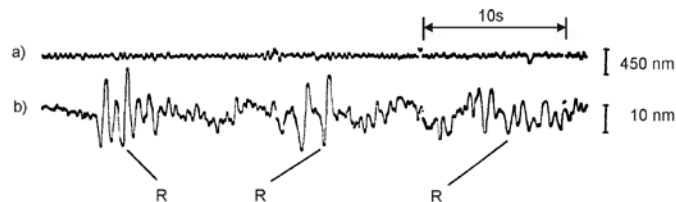
As shown in Fig. 4.29, Fig. 4.32 and Fig. 4.33, short-period seismic noise is strongly reduced with the depth in sensor installations in mines and boreholes. However, when installing seismometers at depth, one must also consider effects on the signal. Generally, amplitudes of seismic body waves recorded at the free surface are systematically increased by as much as a factor of two, depending on the angle of incidence and wavelength (see Exercise 3.4, Tab. 1). On the other hand, at a certain depth, destructive interference between incoming and surface-reflected waves may cause signal reduction. Therefore, because of the “free-surface effect”, peculiarities of the local noise field and geological conditions, the SNR does not necessarily increase steadily with depth. Fig. 4.40 compares two case studies of short-period signal and noise measurements in two deep boreholes in the USA.

The ratio of the noise in the borehole and at the surface,  $S_B/S_{OF}$ , differs in the two boreholes. Its mean value drops in the Texas borehole to  $1/10^{\text{th}}$  at about 1500 m depth and increases again to  $1/2$  of its surface value at 3000 m depth, while in the Oklahoma borehole it drops to about  $1/3$  at about 1000 m depth and then remains roughly constant thereafter. Accordingly, we have no SNR improvement (on average) in the Texas borehole down to about 1000 m depth, but then the SNR increases to a factor of about 15 at 3000 m depth. Contrary to this, the SNR increases by a factor of 3 in the Oklahoma borehole within the first 800 m, but then remains roughly constant (ranging between 1 and 5) up to 3000 m depth.



**Fig. 4.40** Depth dependence of the signal-to-noise ratio (SNR). The top curve in both figures shows the improvement in SNR. The abbreviations are: SB/SOF: Ratio of signal in borehole and at the surface; NB/NOF: Ratio of noise in borehole and at the surface. SNR improvement in a borehole in Texas is shown left and in Oklahoma right. (Figure copied from Bormann, 1966; redrawn from Douze, 1964).

Thus there is no straight-forward and continuous SNR improvement with depth. It may depend also on local geological and installation conditions. Nevertheless, we can generally expect a significant SNR improvement within the first few hundred meters depth. This applies particularly to borehole installations of long-period and broadband sensors which benefit already at much shallower depth greatly from the very stable temperature conditions and strongly reduced tilt noise, also in hard rock tunnels. For more information about tunnel installations see section 7.4.3 in Chapter 7.



**Fig. 4.41** Recording of a teleseismic event at  $D = 80^\circ$  at a) the surface and b) at 3000 m depth in the Texas borehole (see Fig. 4.40 left). The SNR in record b) has improved by a factor of about 10. Note that signal arrivals in the borehole record are followed by surface reflections (R) about 3 s later (Figure copied from Bormann, 1966; redrawn from Douze, 1964).

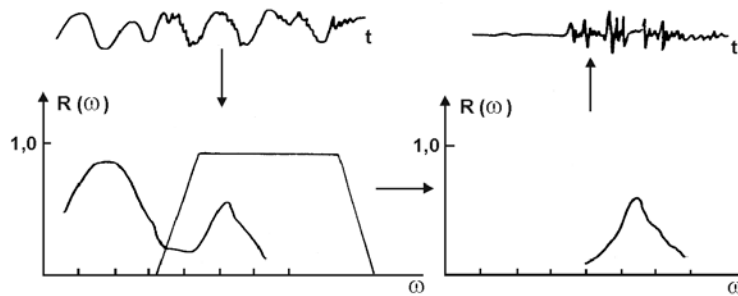
In mines or boreholes a depth of 100 m is generally sufficient to achieve most of the practicable reduction by -20 to -30 dB of long-period noise with periods between 30 s and 1000 s) (see Fig. 7.59 in section 7.4.5 of Chapter 7). According to Murphy and Savino (1975) the effect of atmospheric pressure fluctuations on horizontal component noise is reduced by more than 90% at a depth of 150 m. In this context one should also be aware that in records of deep borehole installation the superposition of the first arriving waves with their respective surface reflection may cause irritating signal distortions although they can be filtered out by tuned signal processing (Fig. 4.41). However, for installations less than 200 m deep the travel-time



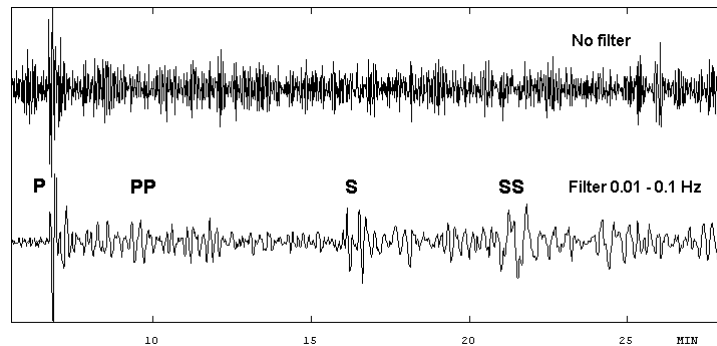
difference between direct and reflected waves is (in consolidated rock) less than 0.1 s and then usually negligible. Since the cost of drilling and installation increases greatly with depth, no deeper permanent seismic borehole installations have yet been made. In any event, the borehole should be drilled through the soil or weathered rock cover and penetrate well into the compacted underlying rock formations.

## 4.7 Improving the Signal-to-Noise Ratio by data processing

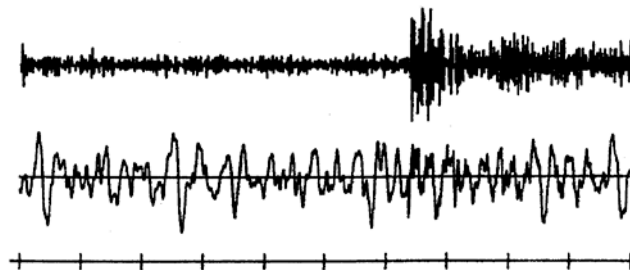
### 4.7.1 Frequency filtering



**Fig. 4.42** Principle of FOURIER transform and bandpass filtering of a seismic record.

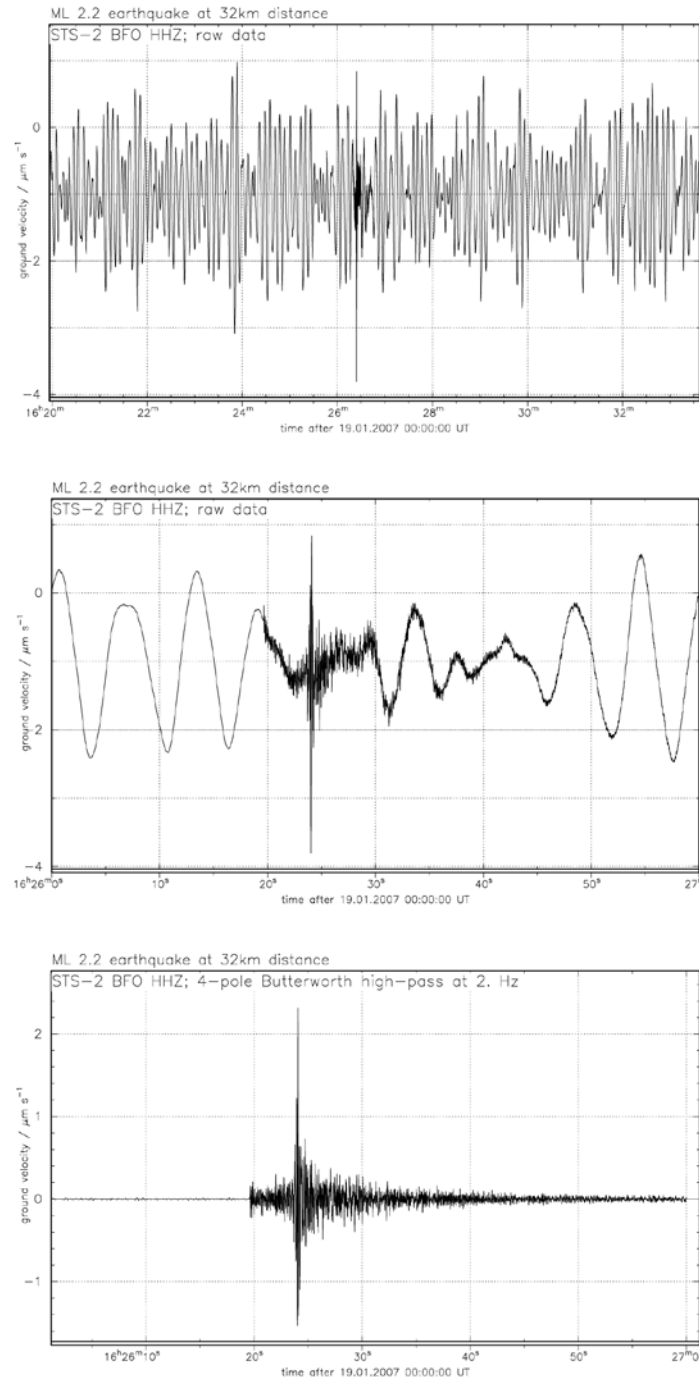


**Fig. 4.43** Recording of a long-period trace at a broadband station. The LP trace has a flat velocity response from 360 s to 0.5 s. On the unfiltered trace (top), only the P-phase might be identified, while on the filtered trace (bottom), the signal-to-noise ratio is much improved and several later phases are clearly recognizable since the microseisms have been removed by filtering (courtesy of J. Havskov, 2001).



**Fig. 4.44** Original (bottom) and frequency filtered record (top;  $f = 2.0 - 4.0$  Hz) of an underground nuclear explosion at the Semipalatinsk test site, Eastern Kazakhstan ( $D = 38^\circ$ ) at station 01A00 of the NORSAR array. Time marks in seconds (from Tronrud, 1983b).

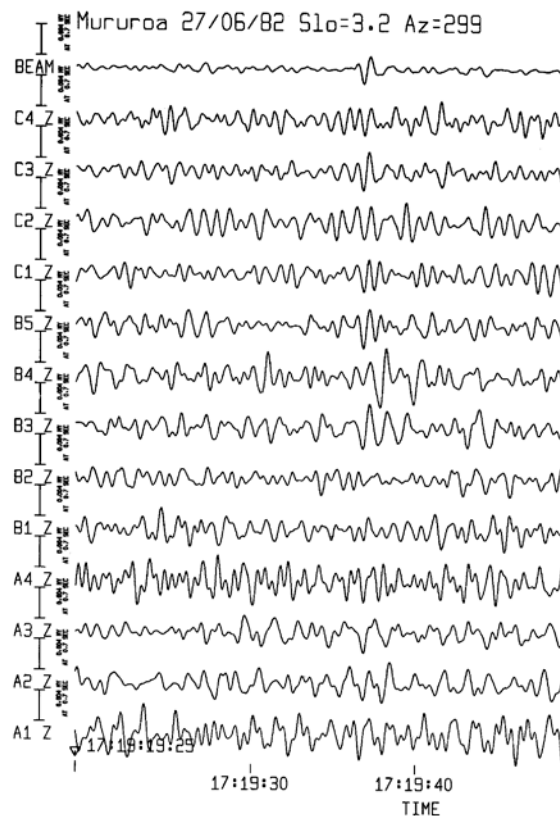
When the frequency spectrum of the seismic signal of interest differs significantly from that of the superposed seismic noise, band-pass filtering can help to improve the signal-to-noise ratio (SNR). Fig. 4.42 illustrates the principle and Fig. 4.43 to Fig. 4.45 show examples.



**Fig. 4.45** Isolation of a local  $M_l = 2.2$  earthquake that occurred 32 km away from the Black Forest Observatory (BFO), Germany, by frequency filtering of the raw data of a velocity broadband STS-2 record. Top: 10 min record of raw data; Middle: stretched 1 min record of raw data; Bottom: elimination of the 6 s ocean microseisms by 2-Hz highpass filtering. (Plots by courtesy of Thomas Forbriger).

## 4.7.2 Velocity filtering and beamforming

Often the dominant signal frequencies may coincide with that of strong noise. Then frequency filtering does not improve the SNR. On the other hand, the horizontal propagation velocity of noise, being dominantly surface waves of Rayleigh-wave type, is much lower than that of P waves and also lower than that of teleseismic S waves with a steep angle of incidence. This leads to frequency-wavenumber (f-k) filtering (see Chapter 9) as a way to improve SNR. To be able to determine the horizontal propagation direction and velocity of seismic signals by means of signal correlation, a group of seismic sensors must be deployed. If the aperture (diameter) of the sensor group is within the correlation radius of the signals it is called a seismic array (see Chapter 9); otherwise the group of sensors comprises a station network (see Chapter 8). Assuming that the noise within the array is random while the signal is coherent, even a simple direct summation of the  $n$  sensor outputs would already produce some modest SNR improvement. When the direction and velocity of travel of a signal through an array is known, one can compensate for the differences in arrival time at the individual sensors and then sum-up all the  $n$  record traces (beam forming). This increases the signal amplitude by a factor  $n$  while the random noise amplitudes increase in the beam trace only by  $\sqrt{n}$ , thus improving the SNR by  $\sqrt{n}$ . Fig. 4.46 compares the (normalized) individual records of 13 stations of the Gräfenberg array, Germany with the beam trace. A weak underground nuclear explosion at a distance of  $143.6^\circ$ , which is not recognizable in any of the single traces, is very evident in the beam trace.

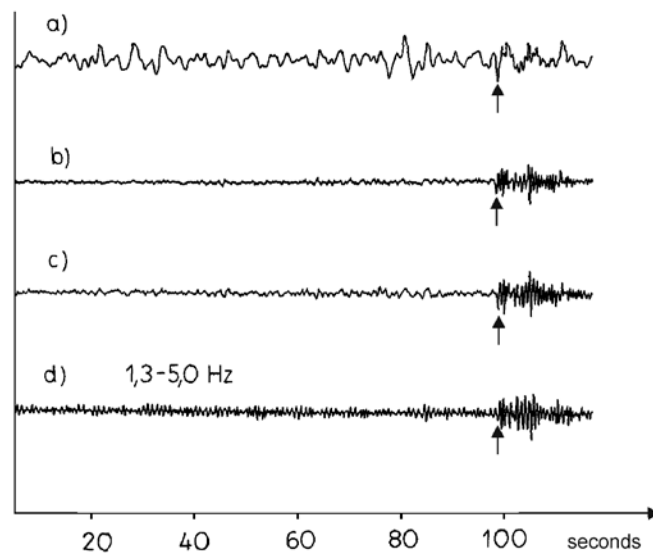


**Fig. 4.46** Detection of a weak underground nuclear explosion in the 10 kt range at the Mururoa Atoll test site ( $D = 145^\circ$ ) by beam forming (top trace). No signal is recognizable in any of the 13 individual record traces from stations of the Gräfenberg array, Germany (below) (from Buttkus, 1986; with permission of Schweizerbart [www.schweizerbart.de](http://www.schweizerbart.de)).

### 4.7.3 Noise prediction-error filtering

In near real time, it is possible to use a moving time-window to determine the characteristics of a given noise field by means of cross- and auto-correlation of array sensor outputs. This then allows the prediction of the expected random noise in a subsequent time interval. Subtracting the predicted noise time series from the actual record results in a much reduced noise level. Weak seismic signals, originally buried in the noise but not predicted by the noise “forecast” of the prediction-error filter (NPEF) may then stand out clearly. NPEFs have several advantages as compared to frequency filtering (compare with Fig. 4.47):

- No assumptions on the frequency spectrum of noise are required since actual noise properties are determined by the correlation of array sensor outputs;
- While frequency differences between signal and noise are lost in narrowband filtering, they are largely preserved in the case of the NPEF. This may aid signal identification and onset-time picking;
- Signal first-motion polarity is preserved in the NPEF whereas it is no longer certain after narrow-band or zero-phase band-pass filtering (see section 4.5).



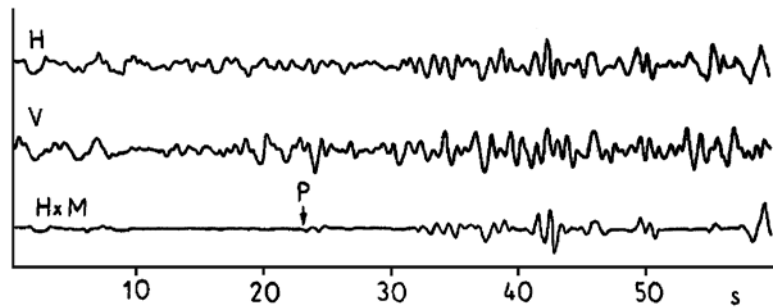
**Fig. 4.47** Records of an underground nuclear explosion recorded at the Uinta Basin small aperture seismic array a) in the beam trace (sum of 10 seismometers), b) and c) after noise prediction error filtering with and without cross correlation (see 4.4.2) and d) after frequency band-pass filtering (1.3 – 5 Hz) (compiled by Bormann, 1966, from data published by Claerbout, 1964).

### 4.7.4 Noise polarization filtering

3-component recordings allow one to reconstruct the ground particle motion and to determine its polarization. Shimshoni and Smith (1964) investigated the cross product

$$(4.20) \quad M_j = \sum_{i=-n}^n H_{i+j} \cdot V_{i+j}$$

in the time interval  $j - n$  to  $j + n$  with  $H$  and  $V$  as the horizontal and vertical component recordings, respectively.  $M$  is a measure of the total signal strength as well as of the degree of linear wave polarization. Eq. (4.20) vanishes for Rayleigh, Love and SH waves. On the other hand, for linearly polarized P and SV waves,  $H$  and  $V$  are exactly in phase and the correlation function becomes  $+1$  for P and  $-1$  for SV waves. The longer the integration time, the better the suppression of randomly polarized noise with a high Rayleigh-wave component. The optimal window length for good noise suppression, while still allowing good onset time picking, must be found by trial and error. Fig. 4.48 gives an example. One great advantage of polarization filtering is that it is independent of differences in the frequency and velocity spectrum of signal and noise and thus can be applied in concert with other procedures for SNR improvement.

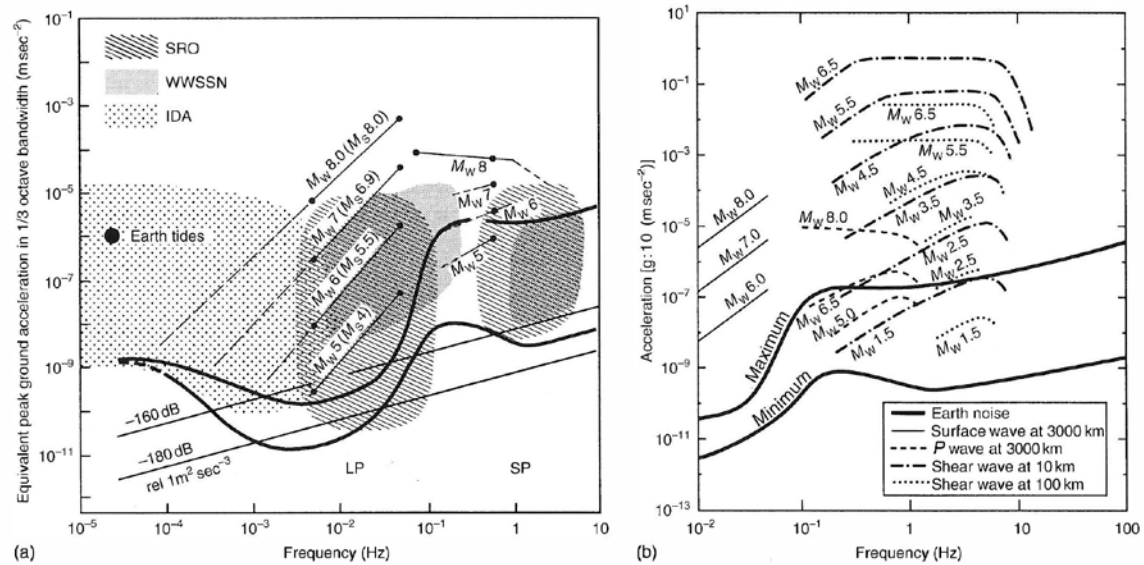


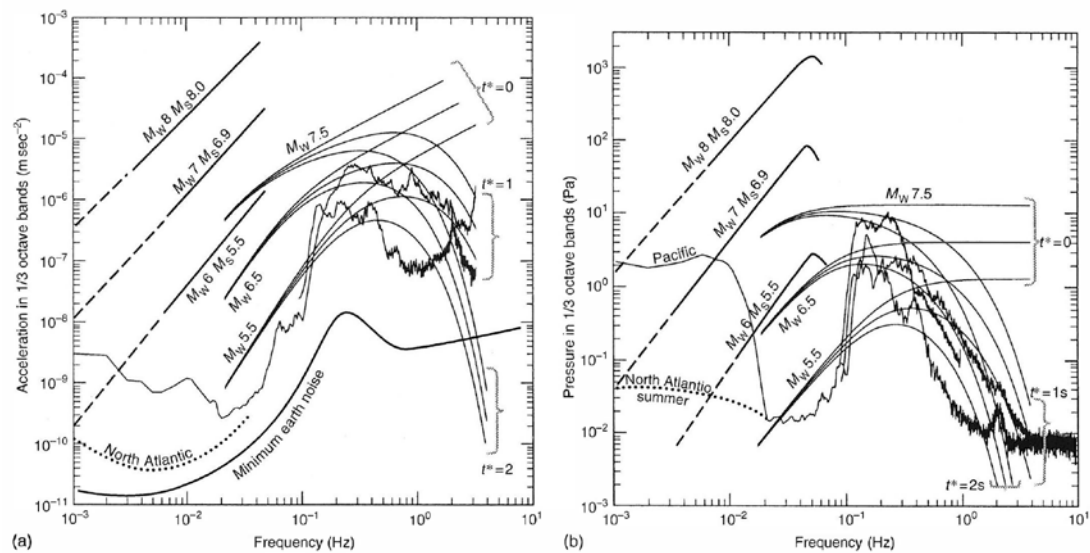
**Fig. 4.48** Example of SNR improvement by polarization filtering according to Eq. (4.20) (bottom trace).  $H$  – horizontal component record,  $V$  – vertical component record (modified from Shimshoni and Smith, 1964).

## 4.8 Global detection thresholds

### 4.8.1 Detection thresholds for surface waves and long-period body waves

Webb (2002) discusses on the basis of data and diagram compilations by several authors the detection thresholds on land-based and ocean bottom station sites for seismic events in terms of their moment magnitudes. According to Fig. 4.49a the global noise levels are sufficiently low at most land-based sites to allow the detection of long-period seismic arrivals from large distant earthquakes. Many of the Earth normal modes can be detected at the quieter sites from the largest events as well. Signal- to-noise ratios are estimated by comparing the root mean square (rms) noise levels with model signal amplitudes in  $1/3$  octave bands. Signal levels are estimated for teleseismic distances of  $30^\circ$ . The solid lines relate to surface wave and body wave amplitudes. The broken lines, extending the surface wave curves, represent normal mode amplitudes. Surface wave spectra, however, vary greatly with source mechanism and depth. Accordingly, surface waves from shallow events are barely detectable near  $0.04$  Hz at most sites from events exceeding about  $M_w = 4.5$ , but the detection of lower-frequency surface waves and normal modes requires larger events. Body waves can be detected at most sites at frequencies below the microseism peak from events with  $M_w > 4.5$ . Smaller events can be detected at smaller distances.





**Fig. 4.50** Same as Figure 4.49 for sea-floor sites for (a) vertical acceleration, and (b) pressure with noise levels from an atypically quiet (calm wind) interval (left) and from a more typical noisy (windy) interval (right). Signal amplitudes are modeled for three different values of attenuation ( $t^*$ ). High noise levels near 1 Hz make it difficult to detect short-period teleseismic body wave phases. (Copied from Figure 4 in Webb, 2002; © Academic Press, Amsterdam).

The detection thresholds depend on distance and travel-path as well as wave-type related attenuation. The rapid decrease in microseism noise at frequencies larger than the frequency of the microseism peak means that small changes in the frequency content of seismic phases may be associated with large differences in detection thresholds. S-waves are more attenuated than P-waves (see section 2.5.4.2 and Fig. 2.79 in Chapter 2 as well as the compressional and shear Q tables and formula in DS 2.1). The dominating frequency content of shear waves may therefore be too low in frequency at teleseismic distances to be seen at all above the large amplitudes of the microseism peak. P waves, however, have at teleseismic distances still enough energy around 1 Hz to be seen on records at land sites but not at generally more noisy sea-floor sites. According to Fig. 4.49 the detection thresholds at most land sites for P waves from distant events ( $>30^\circ$ ) will be roughly magnitude Mw 4.0, with slightly lower values for the quieter sites and much higher detection thresholds for the noisier island sites. According to Webb (2002) this is in rough agreement with more careful estimates of detection thresholds for groups of stations. Recent estimates for global event detection thresholds using seismic arrays range from about Mw 3.5 for heavily instrumented areas such as Europe to greater than Mw 4.5 for remote regions of the world's oceans (e.g., Kvaerna and Ringdal, 1999). Fig. 4.49 suggests that all stations will detect S wave arrivals from events as small as about Mw 2.5 at close range (100km).

Efforts to detect small underground nuclear explosions have focused on detecting high-frequency Pn and Sn arrivals at regional distances over continental platforms. Pn is then detectable at frequencies up to 50 Hz at 500 km and at distances up to 1500 km at 10 Hz from magnitude Ml 3.0 events. Also high-frequency oceanic Pn and Sn waves can still be detected over great distances from moderate oceanic events (lower detection threshold mb  $\approx$  4.6) because at their higher frequency the ocean microseism noise is low (Butler et al., 1987). However, at most sea-floor sites the high noise levels near 1 Hz and high attenuation in the oceanic upper mantle allow to detect teleseismic P waves with periods around 1 s only with poor SNR

from events with  $M_w > 7$  (Blackman et al., 1995). However, the lower noise levels within sea-floor boreholes may lower the detection thresholds for short-period teleseismic body waves from the  $M_w = 7.5$ , expected for noisy sea-floor sites with a saturated microseism spectrum, to about  $M_w 6.5$  and detection limits for quiet sea-floor noise conditions should be below  $M_w 5.5$ . Orcutt and Jordan (1986), e.g., reported on detection thresholds from a sensor in a borehole in the western Pacific, which is at a site where light winds are expected more than 73% of the time, suggesting low noise levels near 1 Hz. The lower detection limits increased from about  $m_b = 5.2$  at a distance of  $40^\circ$  to about  $m_b = 5.6$  at  $80^\circ$ . This is consistent with the low-noise model of Fig. 4.3. Fig. 4.50a, combined with a 10 dB SNR improvement due to the borehole installation, particularly if the observations are associated with deep earthquakes from the subduction zones of the western Pacific, with signals propagating along low-attenuation paths.

Wilcock et al. (1999) arrived at similar conclusions with respect to lower detection thresholds in an ocean environment. They investigated wind statistics over the world's ocean-floor ridges to look for regions where long intervals of calm winds are more common. In these quiet regions noise levels will often be 20 dB lower than at typical sites with wind-noise saturation, and thus their associated detection thresholds several magnitude units smaller. Figure 4.50a shows noise in 1/3 octave bands from a site on the sea floor during a typical noisy period and Fig. 4.50 during one of the rare low-wind days from the MELT experiment. Teleseismic detection thresholds at the sea floor depend on attenuation below a site which is represented by the value of  $t^*$  shown on the model curves for P-wave amplitude for earthquakes of magnitudes 5.5, 6.5, and 7.5. The microseism noise spectrum falls quickly with increasing frequency. Therefore detection thresholds for short-period teleseismic P waves are lower. The model curves for  $t^* = 0$  suggest that only the largest earthquakes ( $M_w > 7.5$ ) will be detected at typical noisy sea-floor sites near ridge crests (where attenuation is high). The detection threshold at sites on older oceanic crust, and for arrivals from deep earthquakes for which attenuation is less ( $t^* = 1$ ) are lower ( $M_w \approx 6$ ). Finally, at sea-floor sites that see long intervals with calm conditions overhead (see Fig. 4.50b) the detection threshold may fall as low as  $M_w = 5.5$ . The detection thresholds at local and regional distances for sites on the sea floor are even much lower because of the much higher frequency content of the wave arrivals as compared to that of the microseism peak. Local events with  $M_w$  between -1.5 and 20 have been detected during microearthquake studies on ridge crests.

## Acknowledgments

The authors thank E. Bergman, J. Havskov and E. Hjortenberget for carefully reading the draft for the first edition and for their valuable suggestions which helped to improve the text and a few figures. Still these parts make for most of the revised and amended 2<sup>nd</sup> edition. The new section 4.2 and parts of sections 4.3 and 4.4 are based on lecture notes by E. Wielandt, who also contributed the new Figs. 4.4, 4.5 and 4.6. Thanks go also to two reviewers of the upgraded version of this Chapter for their valuable comments. Th. Forbriger made lecture notes on seismic noise available as well, provided Fig. 4.45 and recommended additional references. J. R. R. Ritter checked the correctness of section 4.3.3.6 on urban noise. Several topically well-fitting figures were reproduced, with proper acknowledgment and permission by the authors, from other related NMSOP-2 contributions, e.g., Fig. 4.11 from L. Küperkoch, Fig. 4.18 from S. Wendt, Fig. 4.20 from Th. Diehl, and Fig. 4.24 and Fig. 4.25 from M. Shinohara.



## Recommended overview readings

Buttkus (2000)  
Galperin et al. (1978)  
Havskov and Alguacil (2002)  
Kumar and Ahmed (2011)  
Richardson et al. (1995)  
Scherbaum (2007)  
Tabulevich (1992)  
Webb (1998, 2002)

## References

- Agnew, D. C., Berger, J., Farrell, W. E., Gilbert, J. F., Masters, G., and Miller, D. (1986). Project IDA: a decade in review. *EOS Trans. Am. Geophys. Union*, **67**(10), 203-212.
- Aki, K., and Richards, P. G. (1980). *Quantitative seismology – theory and methods*, Volume II, Freeman and Company, ISBN 0-7167-1059-5(v. 2), 609-625.
- Aki, K., and Richards, P. G. (2002). *Quantitative seismology*. 2<sup>nd</sup> ed., ISBN 0-935702-96-2, University Science Books, Sausalito, CA., xvii + 700 pp.
- Aki, K., and Richards, P. G. (2009). *Quantitative seismology*. 2<sup>nd</sup> ed., corr. Print., Sausalito, Calif., Univ. Science Books, 2009, XVIII + 700 pp., ISBN 978-1-891389-63-4 (see [http://www.ldeo.columbia.edu/~richards/Aki\\_Richards.html](http://www.ldeo.columbia.edu/~richards/Aki_Richards.html)).
- Araki, E., Shinohara, M., Sacks, S., Linde, A., Kanazawa, T., Shiobara, H., Mikada, H., and K. Suyehiro (2004). Improvement of seismic observation in the ocean by use of seafloor boreholes. *Bull. Seism. Soc. Am.*, **94**, 678-690.
- Aster, R. C. and Shearer, P. M. (1991). High frequency seismograms recorded in the San Jacinto fault zone, Southern California, part 2. Attenuation and site effects. *Bull. Seismol. Soc. Am.*, **81**(4), 1081-1100.
- Bard, P.-Y. (1999). Microtremor measurements: A tool for site effect estimation? In: Irikura, K., Kudo, K., Okada, H., and Sasatani, T. (Eds.) (1999). The effects of surface geology on seismic motion. *Proc. 2<sup>nd</sup> International Symposium in Yokohama, Japan, 1-3 Dec. 1998*, A. A. Balkema, Rotterdam, ISBN 90 5809 030 2, 1251-1279.
- Beauduin, R. and J.-P. Montagner (1996). Time evolution of broad- band seismic noise during the French pilot experiment OFM/ SISOBS. *Geophys. Res. Lett.*, **23**(21), 2995-2998.
- Beauduin, R., Lognonné, P., Montagner, J. P., Cacho, S., Karczewski, J. F., and Morand, M. (1996). The effects of the atmospheric pressure changes on seismic signals or How to improve the quality of a station. *Bull. Seism. Soc. Am.*, **86**, 6, 1760-1769.
- Berger, J., Davis, P., Ekström, G. (2004). Ambient Earth noise: A survey of the Global Seismographic Network. *J. Geophys. Res.*, **109**, B11307, doi: 10.1029/2004JB003408.
- Blackman, E. K., Orcutt, J. A., and Forsyth, D. W. (1995). Recording teleseismic earthquakes using ocean-bottom seismographs at mid-ocean ridges. *Bull. Seism. Soc. Am.*, **85**, 6, 1648-1664.
- Bormann, P. (1966). *Recording and interpretation of seismic events (principles, present state and tendencies of development)* (in German). Publ. Inst. Geodyn., Jena, Akademie Verlag, Berlin, 158 pp.

- Bormann, P. (1973). Standardization and optimization of frequency-characteristics at Moxa station (GDR). Proceed. XII Assembly of ESC, Bukarest, *Technical and Economic Studies*, D-Series, **No. 10**, 133-145.
- Bormann, P. (1998). Conversion and comparability of data presentations on seismic background noise. *J. Seismology*, 2(1), 37-45; doi: 10.1023/A:1009780205669.
- Bormann, P., Wylegalla, K., Strauch, W., and Baumbach, M. (1992). Potsdam seismological station network: processing facilities, noise conditions, detection threshold and localization accuracy. *Phys. Earth Planet. Int.*, **69**, 311-321.
- Bormann, P., Wylegally, K., and Klinge, K. (1997). Analysis of broadband seismic noise at the German Regional Seismic Network and search for improved alternative station sites. *J. Seism.*, **1**, 357-381.
- Bormann, P. (1998). Conversion and comparability of data presentations on seismic background noise. *J. Seism.*, **2**, 37-45.
- Braddner, H., and Dodds, I. G. (1964). Comparative seismic noise on the ocean bottom and on land. *J. Geophys. Res.*, **69**, 20, 4339-4348.
- Bradley, C. R., and Stephen, R. A. (1996). Modeling of seafloor wave propagation and acoustic scattering in 3-D heterogeneous media. *J. Acoust. Soc. Am.*, **100**(1), 225-236.
- Bradley, C.R., Stephen, R. A., Dorman, L. M., and Orcutt, J. A. (1997). Very low frequency (0.2-10 Hz) seismoacoustic noise below the seafloor. *J. Geophys. Res.*, **102**(B6), 11703-11718.
- Broding, R. A., Bentley-Llewellyn, N. J., and Hearn, D. P. (1964). A study of three dimensional seismic detection system. *Geophysics*, **29**, 2, 221-249.
- Brune, J. N., and Oliver, J. (1959). The seismic noise of the Earth's surface. *Bull. Seism. Soc. Am.*, **49**, 349-353.
- Bungum, H., Mykkeltveit, S., and Kvaerna, T. (1985). Seismic noise in Fennoscandia, with emphasis on high frequencies. *Bull. Seismal. Soc. Am.*, **75**, 1489-1513.
- Butler, R., McCreery, C. S., Frazer, L. N., and Walker, D. A. (1987). High frequency seismic attenuation of oceanic P and S waves in the western Pacific. *J. Geophys. Res.*, **92**(B2), 1383-1396.
- Buttkus, B. (Ed.) (1986). Ten years of the Gräfenberg array: defining the frontiers of broadband seismology. *Geologisches Jahrbuch E-35*, Hannover, 135 pp.
- Buttkus, B. (2000). Spectral analysis and filter theory in applied geophysics. Springer Verlag, xv+667 pp., ISBN 3-540-62674-3.
- Carter, J.A., N. Barstow, P.W. Pomeroy, E.P. Chael, and P.J. Leahy (1991). High frequency seismic noise as a function of depth. *Bull. Seismal. Soc. Am.*, **81**(4), 1101-1114.
- Cessaro, R. K. (1994). Sources of primary and secondary microseisms. *Bull. Seism. Soc. Am.*, **84**, 1, 142-148.
- Diehl, T., Kissling, E., and Bormann, P. (2012). Tutorial for consistent phase picking at local to regional distances. In: Bormann, P. (Ed.). *New Manual of Seismological Observatory Practice (NMSOP-2)*, IASPEI, GFZ German Research Centre for Geosciences, Potsdam, 21 pp; doi: 10.2312/GFZ.NMSOP-2\_IS\_11.4; <http://nmsop.gfz-potsdam.de>,
- Douze, E. J. (1964). Signal and noise in deep wells. *Geophysics*, **29**, 5, 721-732.
- Duda, S. J., and Kaiser, D. (1989). Spectral magnitudes, magnitude spectra, and earthquake quantification: the stability issue of the corner period and of the maximum magnitude for a given earthquake. *Tectonophysics*, **166**, 205-219.

- Engdahl, E. R., and Gunst, R. H. (1966). Use of a high speed computer for the preliminary determination of earthquake hypocenters. *Bull. Seism. Soc. Am.*, **56**, 325-336.
- Evans, John R.; Followill, F.; Hutt, Charles R.; Kromer, R. P.; Nigbor, R. L.; Ringler, A. T.; Steim, J. M.; Wielandt, E. (2010). Method for calculating self-noise spectra and operating ranges for seismographic inertial sensors and recorders. *Seismological Research Letters*, **81**: 640 - 646
- Fix, J. E. (1972). Ambient Earth motion in the period range from 0.1 to 2560 sec. *Bull. Seism. Soc. Am.*, **62**, 1753-1760.
- Friedrich, A., Klinge, K., and Krüger, F. (1998). Ocean-generated microseismic noise located with the Gräfenberg array. *J. Seism.*, **2**, 47-64.
- Fyen, J. (1990). Diurnal and seasonal variations in the microseismic noise observed at the NORESS array. *Phys. Earth Planet. Inter.*, **63**, 252-268.
- Galperin, E.I., I.L. Nersesov, and R.M. Galperina (1978). *Borehole Seismology and the Study of the Seismic Regime of Large Industrial Centers*. D. Reidel, Dordrecht.
- Groos, J. C., and Ritter, J. R. R. (2009). Time domain classification and quantification of seismic noise in an urban environment. *Geophys. J. Int.*, **179**, 1213-1231; doi: 10.1111/j.1365-246X.2009.04343.x.
- Hasselmann, K.A. (1963). A statistical analysis of the generation of microseisms. *Rev. Geophys.*, **1**, 177-209.
- Haubrich, R. A., and McCamy, K. (1969). Microseisms coastal and pelagic sources. *Rev. Geophys. Space Phys.*, **7**, 539-571.
- Haubrich, R. A., Munk, W. H., and Snodgrass, F. E. (1963). Comparative spectra of microseisms and swell. *Bull. Seism. Soc. Am.*, **53**, 27-37.
- Havskov, J., and Alguacil, G. (2006). Instrumentation in earthquake seismology. *Springer*, Berlin, 2<sup>nd</sup> edition, 358 pp. plus CD.
- Heaton, T.H., Anderson, D. L., Arabasz, W. J., et al. (1989). National Seismic System Science Plan. *USGC Survey Circular* **1031**, 1-42.
- Holcomb, L. G. (1998). Spectral structure in the Earth's microseismic background between 20 and 40 seconds. *Bull. Seismol. Soc. Am.*, **88**(3), 744-757.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data.  
[http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf)
- IASPEI (2013). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data.  
[http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)
- Küperkoch, L., Meier, Th., and Diehl, T. (2012). Chapter 16: Automated event and phase detection. In: Bormann, P. (Ed.). *New Manual of Seismological Observatory Practice (NMSOP-2)*, IASPEI, GFZ German Research Centre for Geosciences, Potsdam, 52 pp; doi: 10.2312/GFZ.NMSOP-2\_ch16; <http://nmsop.gfz-potsdam.de>,
- Kulhanek, O. (2002). The structure and interpretation of seismograms. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*. Academic Press, Amsterdam, 333-348.
- Kumar, D., and Ahmed, I. (2011). Seismic noise. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, Vol. 2, 1157-1161; doi: 10.1007/978-90-481-8702-7.

- Kvaerna, T., and Ringdal, F. (1999). Seismic threshold monitoring for continuous assessment of global detection capability. *Bull. Seismal. Soc. Am.*, **89**(4), 949-959.
- Lay, T., and Wallace, T. C. (1995). Modern global seismology. ISBN 0-12-732870-X, Academic Press, 521 pp.
- Longuet-Higgins, M. S. (1950). A theory for the generation of microseisms. *Philos. Trans. R. Soc. Lond.*, **A 243**, 1-35.
- McCreery, C.S., F.K. Duennebie, and G.H. Sutton (1993). Correlation of deep ocean noise (0.4-20 Hz) with wind, and the Holu spectrum - a worldwide constant. *J. Acoust. Soc. Am.*, **93**(5), 2639-2648.
- Melton, B. S. (1978). The sensitivity and dynamic range of inertial seismographs. *Rev. Geophys. Space Phys.*, **14**, 393-116.
- Montagner, J. P., Karczewski, J. F., Romanowicz, B., Bouaricha, S., Lognonné, P., Roult, G., Stutzmann, E., Thiot, J. L., Brion, J., Dole, B., Fouassier, D., Koenig, J.C., Savary, J., Louri, L., Dupond, J., Echardour, A., and Floc'h, H. (1994). The French pilot experiment OFM-SISMOBS: first scientific results on noise level and event detection. *Physics Earth Planet. Inter.*, **84**, 321-336.
- Murphy, A.J. and J.R. Savino (1975). A comprehensive study of long period (20-200 s) Earth noise at the high gain worldwide seismograph stations. *Bull. Seismol. Soc. Am.*, **65**, 1827-1862.
- Nakamura, Y. (1989). A method for dynamic characteristics estimation of subsurface using microtremor on the ground surface. *Q. Rep. Railway Tech. Res. Inst.*, **30**, 25-33.
- Orcutt, J. A., and Jordan, T. H. (1986). Data from the Ngendei experiment in the southwest Pacific. In: *"The Vela Program,"* (Ed, A. Kerr), pp. 758-770. Def. Adv. Res. Proj. Agency, Washington, DC.
- Parolai S., Bormann, P., and Milkereit, C. (2001b). Assessment of the natural frequency of the sedimentary cover in the Cologne area (Germany) using noise measurements. *J. Earthq. Engrg.*, **5**, 541-564.
- Parolai S., Bormann, P., Milkereit, C. (2002). New relationships between  $V_s$ , thickness of sediments, and resonance frequency calculated by the H/V ratio of seismic noise for the Cologne area (Germany). *Bull. Seism. Soc. Am.*, **92**, 2521-2527.
- Peterson, J. (1993). Observations and modeling of seismic background noise. *U.S. Geol. Survey Open-File Report 93-322*, 95 pp.
- Richardson, W. J, Greene, C. R., Malme, C. I., and Thomson, D. H. (1995). *Marine mammals and noise*. Academic Press, San Diego.
- Romanowicz, B., D. Stakes, J.-P. Montagner, et al. (1998). MOISE: a pilot experiment toward long term seafloor geophysical observatories. *Earth Planets Space*, **50**, 927-937.
- Savino, J., McCamy, K., and Hade, G. (1972). Structures in Earth noise beyond twenty seconds - a window for earthquakes. *Bull. Seismal. Soc. Am.*, **62**(1), 141-176.
- Scherbaum, F. (2007). *Of poles and zeros; Fundamentals of Digital Seismometry*. 3<sup>rd</sup> revised edition, Springer Verlag, Berlin und Heidelberg.
- Schreiner, A. E., and Dorman, L. M. (1990). Coherence lengths of seafloor noise: effect of ocean bottom structure. *J. Acoust. Soc. Am.*, **88**(3), 1503-1514.
- Sleeman, R., van Wietum, A., and Trampert, J. (2006). Three-channel correlation analysis: A new technique to measure instrumental noise of digitizers and seismoc sensors. *Bull. Seism. Soc. Am.*, **96**(1), 258-271.

- Shimshoni, M., and Smith, S. W. (1964). Seismic signal enhancement with three-component detectors. *Geophysics*, **29**, 5, 664-671.
- Stephen, R. A., Collins, J. A., and Peal, K. R. (1999). Seafloor seismic stations perform well in study. *Eos, Trans. Am. Geophys. Un.*, **80**, 592.
- Tabulevich, V. (1992). Microseismic and infrasound waves. *Springer-Verlag Berlin*, 150 pp.
- Tronrud, L. B. (Ed.)(1983). Semiannual Technical Summary 1 October 1982 – 31 March 1983. *NORSAR Scientific Report No. 1-82/83*, 72 pp.
- Vernon, F. L. et al. (1998). Evaluation of teleseismic waveforms and detection thresholds from the OSN pilot experiment. *EOS Trans. Am. Geophys. Union*, **79**(45), 650.
- Webb S. C. (1998). Broadband seismology and noise under the ocean. *Rev. Geophys.* **36**, 105–142.
- Webb, S. C. (2002). Seismic noise on land and on the sea floor. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). International Handbook of Earthquake and Engineering Seismology, Part A, *Academic Press, Amsterdam*, 305-318.
- Webb, S.C., and Schultz, A. D. (1992). Very low frequency ambient noise at the seafloor under the Beaufort sea icecap. *J. Acoust. Soc. Am.*, **91**(3), 1429-1439.
- Webb, S. C., and Crawford, W. C. (1999). Long period seafloor seismology and deformation under ocean waves. *Bull. Seismal. Soc. Am.*, **89**(6), 1535-1542.
- Wiechert, E. (1904). Verhandlungen der zweiten internationalen Seismologischen Konferenz. *Beitr. Geophys.*, Ergänzungsband II, 41-43.
- Wilcock, W. S. D., Webb, S. C., and Bjarnason, I. Th. (1999). The effect of local wind on seismic noise near 1 Hz at the MELT site and on Iceland. *Bull. Seismal. Soc. Am.*, **89**(6), 1543-1557.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.
- Withers, M. M., Aster, R. C., Young, Ch. J., and Chael, E. P. (1996). High-frequency analysis of seismic background noise as a function of wind speed and shallow depth. *Bull. Seism. Soc. Am.*, **86**, 5, 1507-1515.
- Yanovskaya, T. G. (2012). Surface-wave tomography for upper mantle studies: methods and results. Keynote lecture at the 33<sup>rd</sup> General Assembly of the European Seismological Commission, Moscow, 19-24 Aug. 2012.
- Young, Ch. J., Chael, E. P., Withers, M. W., and Aster, R. C. (1996). A comparison of the high-frequency (>1Hz) surface and subsurface noise environment at three sites in the United States. *Bull. Seism. Soc. Am.*, **86**, 5, 1516-1528.
- Zürn, W. (1974). Detectability of small harmonic signals in digitized records. *J. Geophys. Res., Solid Earth and Planets*, **79**(29), 4433–4438.
- Zürn, W., and Widmer, R. (1995). On noise reduction in vertical seismic records below 2 mHz using barometric pressure. *Geophys. Res. Lett*, **22**(24), 3537-3540.
- Zürn, W., and Wielandt, E. (2007). On the minimum of vertical seismic noise near 3 mHz. *Geophys. J. Int.*, **168**, 647-658.

| Topic   | Bandwidth-dependent transformation of noise data from frequency into time domain and vice versa                                                                                                                                                                                                                                                            |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a><br><b>Erhard Wielandt</b> (formerly Institute of Geophysics, University of Stuttgart, D-70184 Stuttgart, Germany); E-mail: <a href="mailto:e.wielandt@t-online.de">e.wielandt@t-online.de</a> |
| Version | October 2011; DOI: 10.2312/GFZ.NMSOP-2_EX_4.1                                                                                                                                                                                                                                                                                                              |

## 1 Aim

The exercise aims at:

- deepening the understanding and developing skills in using the related equations presented in section 4.1 of Chapter 4, Vol. 1;
- application of the conversion program **noisecon**, which can be downloaded by right mouse click on this program name in the NMSOP-2 content overview folder *Download Programs & Files.*;
- demonstrating that the various data presentations given in this Exercise and in Chapter 4 on signal and noise spectra or amplitudes in different kinematic units are in fact all compatible or – if not – that reasons for it can be given.

## 2 Fundamentals

The underlying fundamentals have been outlined in detail in the introduction to Chapter 4. In summary, the following should be remembered:

When a broadband signal is split up into narrower frequency bands with ideal band-pass filters, then

- the instantaneous amplitudes in the individual bands add up to the instantaneous amplitude of the broadband signal,
- the signal powers (or energies in case of *transient* signals) in the individual bands add up to the power (or energy) of the broadband signal,
- the RMS amplitudes in the individual bands DO NOT add up to the RMS amplitude of the broadband signal. The rms amplitude must be calculated from the total power.

**A specification of noise amplitudes without a definition of the bandwidth is meaningless!**

Also: Signal energy is the time-integral of power. Accordingly, transient signals have a finite energy while stationary (noise) signals have an infinite energy but a finite and, in the time average, constant power. Transient signals and stationary signals must therefore be treated differently. The spectrum of a transient signal cannot be expressed in the same units as the spectrum of a stationary signal. Earthquake spectra and noise spectra can, therefore, not be represented in the same plot, unless the conversion between the units is explained. Also, band-pass filtered amplitudes in different spectral ranges are comparable only when having been filtered with the same relative bandwidth (RBW). Note that in signal analysis the

"power" of a signal is understood to be the mean square of its instantaneous amplitude. Physical power is proportional but not identical to what is called "power" in signal analysis - for example, the electric power is  $W = U^2 / R$ , not  $W = U^2$ .

### 3 Data, relationships and programs

The exercises are based on data presented in Figs. 4.5 to 4.8 in Chapter 4, and the Figures 1 to 3 below.

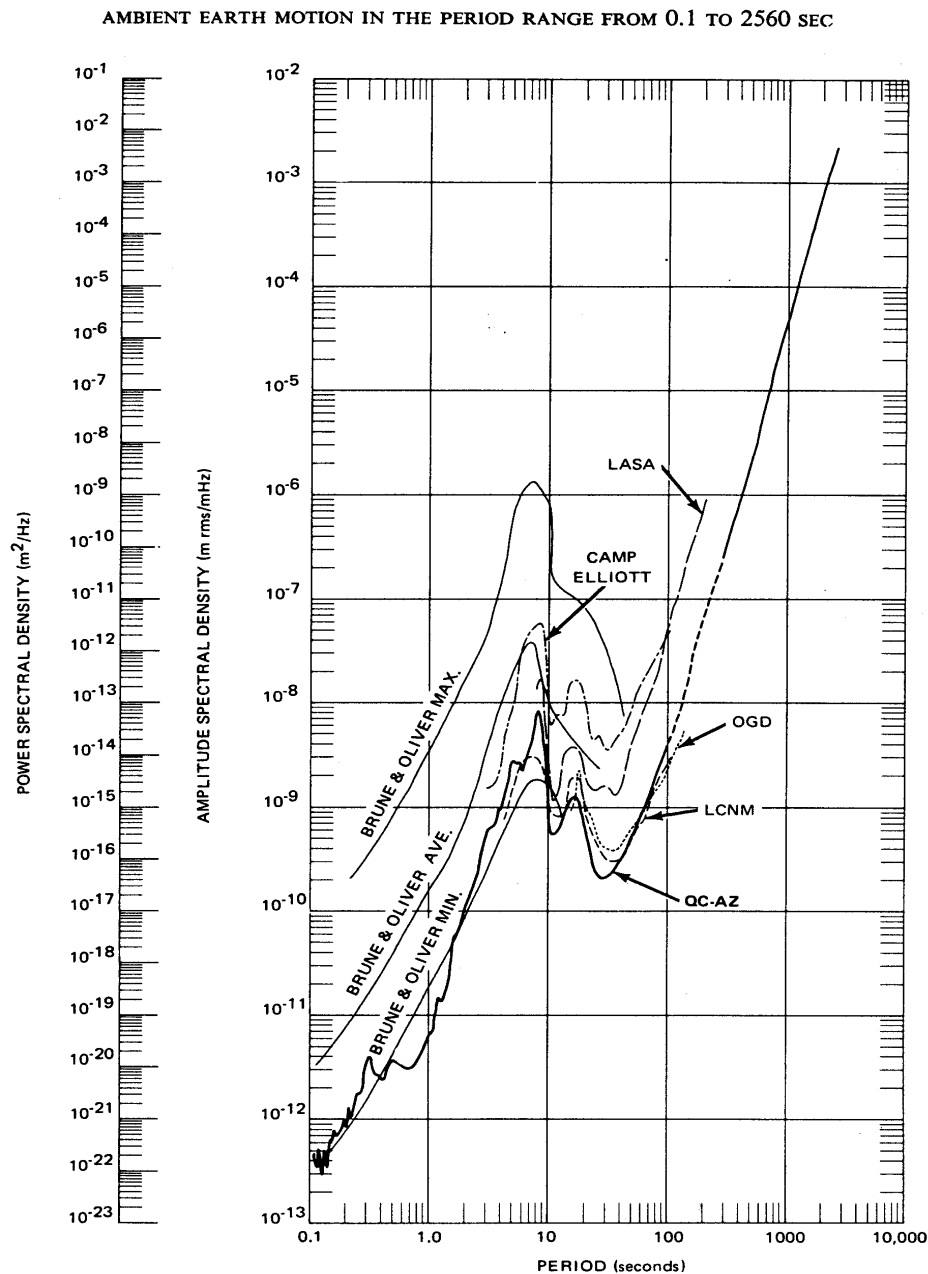


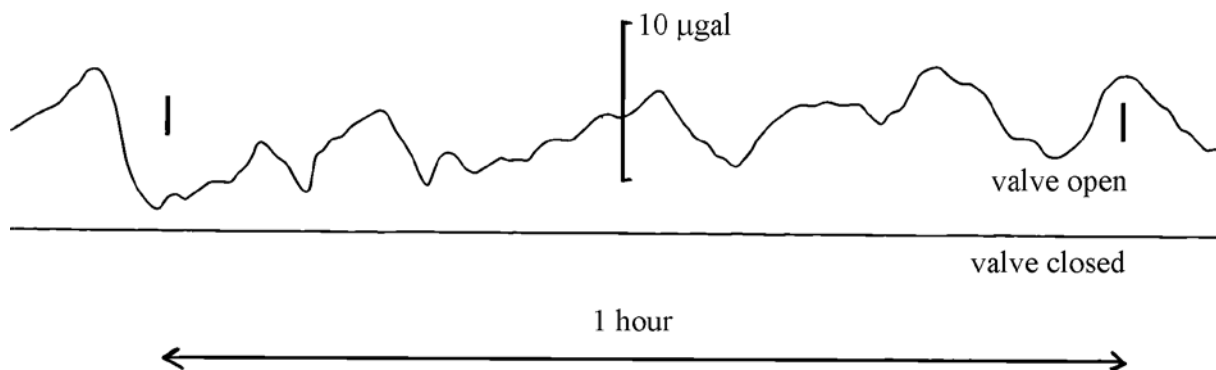
FIG. 1. Vertical Earth motion amplitude spectral density.

**Figure 1** Compilation of various noise amplitude and power spectral densities at various stations and according to the Brune and Oliver (1959) noise model as published by Fix (1972).





**Figure 2** Cut-out section of a record of the WWSSN-LP seismograph of strong secondary ocean microseisms caused by a winter storm over the Atlantic ocean, reproduced at original scale (30 mm = 1 minute). The magnification at the dominant period is about 400 times.



**Figure 3** Output signal of an STS1 seismometer with the vacuum bell valve open (upper trace) and closed (lower trace), respectively. The noise in the top trace is caused by changes in barometric pressure.

For manual solutions use the respective relationships given in Eqs. (4.4) to (4.17) of Chapter 4 and a pocket calculator with the required basic functions. Alternatively, you may use the program **noisecon**.



## 4 Tasks

**Task 1:** Determine the relative bandwidth (RBW) of an

- a) 2-octave filter
- b) 2/3-octave filter
- c) 1/3-octave filter
- d) 1/6-decade filter

by using Eq. (4.15) in Chapter 4 and

- e) express the bandwidth of an 1/3-decade filter in terms of octaves

by using Eq. (4.17).

**Task 2:** Calculate for the noise maximum of the upper curve in Fig. 4.5 the corresponding RMS ground motion (velocity and displacement).

- a) Estimate the velocity power maximum from Fig. 4.5 (Note the logarithmic scale!).
- b) Give this value also in units of  $(\text{m/s})^2/\text{Hz}$ .
- c) Estimate the frequency  $f_0$  related to this maximum.
- d) Calculate the RMS-velocity amplitude  $a_{v\text{RMS}}$  by considering Eqs. (4.15) and (4.16) and a relative bandwidth of 2/3 octaves.
- e) Transform this RMS velocity amplitude determined under d) into the corresponding RMS displacement amplitude  $a_{d\text{RMS}}$  considering Eq. (4.4).

**Task 3:** Transform the displacement power values of Fig. 4.6 at  $f = 1 \text{ Hz}$  and  $f = 10 \text{ Hz}$  in

- a) units of  $\text{m}^2/\text{Hz}$ ,
- b) acceleration power values with units  $(\text{m/s}^2)^2/\text{Hz}$  using Eq. (4.5),
- c) the values determined under b) in units of dB referred to  $1 (\text{m/s}^2)^2/\text{Hz}$  according to Eq. 4.6) and
- d) compare the result with the respective values in Fig. 4.7 for the New Low Noise Model (NLNM).

**Task 4:** Determine from Fig. 4.7 the respective ground acceleration power spectral density values of the NHNM in units of  $(\text{nm/s}^2)^2/\text{Hz}$  for

a)  $f = 1 \text{ Hz}$  ,

b)  $f = 0.1 \text{ Hz}$  .

using Eq. (4.6)

**Task 5:** Select any period between 0.01 and 10,000 sec (e.g.,  $T = 100\text{s}$ ) and confirm that the presentations in Figs. 4.7 and 4.8 in Chapter 4 are equivalent when assuming a relative bandwidth of 1/6 decades as used in Fig. 4.8.

**Task 6:** Transform selected velocity PSD values given in the lower curve of Fig. 4.23 into acceleration PSD  $\mathbf{P}_a[\text{dB}] = 10 \log (\mathbf{P}_a / 1 \text{ (m/s}^2\text{)}^2/\text{Hz})$  via Eq. (4.5) and compare them with the NLNM at

a) 1.5 Hz and

b) 10 Hz.

**Task 7:** Figure 1 is a historical compilation of observed low noise spectra, now made obsolete by the USGS New Low Noise Model (NLNM).

- a) One of the two vertical scales has an incorrect label; both the name and the unit are incorrect. Try to correct it.
- b) Compare the lowermost curve (QC-AZ) to the NLNM, for example by plotting the points of the NLNM curve for 30 s, 300 s and 3000 s into the figure and discuss the difference.
- c) What would be the rms acceleration of the QC-AZ noise in a half-octave frequency band from 1 to (roughly) 1.4 Hz and in a half-octave period band from 25 to 35 seconds? Compare to figure 5.18 of the NLNM. Note that  $\frac{1}{2}$  octave and  $\frac{1}{6}$  decade are nearly the same.

**Task 8:** Assess the noise level of the microseism storm in Figure 2 with respect to the NLNM

- a) Determine the range of periods of the microseisms.
- b) Estimate the bandwidth of the microseisms, their center frequency  $f_0$  and RBW.
- c) Estimate the displacement  $a_{\text{RMS}}$  from the *average peak amplitudes* (which are about  $1.25a_{\text{RMS}}$ ). The magnification of the record is about 400 at  $f_0$ .
- d) Transform the displacement  $a_{\text{RMS}}$  into acceleration  $a_{\text{RMS}}$  and  $\mathbf{P}_a [\text{dB}]$ .
- e) Compare with Fig. 4.7 and discuss possible differences.

**Taks 9:** Compare the noise level for the acceleration records of an STS1 seismometer shown in Figure 3 and compare it with the NLNM. Note that  $1 \text{ gal} = 10^{-2} \text{ m/s}^2$ .

- Estimate  $a_{\text{RMS}}$  from the *average peak amplitudes* in Figure 3, upper trace.
- Estimate the upper limit of  $a_{\text{RMS}}$  for the lower trace in Figure 3.
- Estimate the periods and bandwidth of the noise in Figure 3.
- Compare the  $a_{\text{RMS}}$  for the upper and the lower trace with the NLNM presentation in Fig. 4.8.
- Discuss the differences.

## 5 Solutions

**Note:** The errors in eye readings of the required parameters from the diagrams may be 10 to 30 %. Therefore, it is acceptable if your solutions differ from the ones given below in the same order or by about 1 to 3 dB in power. In case of larger deviations check your readings and calculations. Also: all power values given below in dB relate to the respective units in Fig. 4.7.

- Task 1:**
- 1.5
  - 0.466
  - 0.231
  - 0.386
  - 1.1 octaves

- Task 2:**
- $7 \times 10^{-8} (\text{cm/s})^2/\text{Hz}$
  - $7 \times 10^{-12} (\text{m/s})^2/\text{Hz}$
  - 0.16 Hz
  - $a_{\text{vRMS}} \approx 7 \times 10^{-7} \text{ m/s}$
  - $a_{\text{dRMS}} \approx 7 \times 10^{-7} \text{ m}$

- Task 3:**
- $2 \times 10^{-18} \text{ m}^2/\text{Hz}$  at 1 Hz and  $1.5 \times 10^{-22} \text{ m}^2/\text{Hz}$  at 10 Hz
  - $3.12 \times 10^{-15} (\text{m/s}^2)^2/\text{Hz}$  at 1 Hz and  $2.3 \times 10^{-15} (\text{m/s}^2)^2/\text{Hz}$  at 10 Hz

c) - 145 dB for 1 Hz and -146 dB for 10 Hz

d) The noise power at this site is for the considered frequencies about 20 dB higher than for the NLNM.

**Task 4:** a) and b)  $\approx -117$  dB, i.e.,  $\approx 2 \times 10^6 \text{ (nm/s}^2\text{)}^2/\text{Hz}$  ;

**Task 5:** For  $T = 100$  s we get from Fig. 4.7  $\mathbf{P_a}$  [dB] = -185 dB. With  $\text{RBW} = 0.3861$  for 1/6 octave bandwidth and  $f = 0.01$  Hz we calculate with Eq. (4.16)  $a_{\text{RMS}}^2 = 1.1 \times 10^{-21} \text{ (m/s}^2\text{)}^2$  which is about -210 dB, in agreement with Fig. 4.8.

**Task 6:** a)  $\mathbf{P_a} \approx -153$  db, 16 dB above the NLNM

b)  $\mathbf{P_a} \approx -153$  db, 15 dB above the NLNM

**Task 7:**

a) The amplitude-density scale in Figure 1 is inapplicable to noise and cannot be related to the power-density scale, which is correct.

b) It is convenient to express the amplitudes in decibels relative to  $1 \text{ m}^2/\text{Hz}$ . For example,  $10^{-13} \text{ m}^2/\text{Hz}$  corresponds to -130 dB,  $2 \times 10^{-13} \text{ m}^2/\text{Hz}$  to -127 dB,  $5 \times 10^{-13} \text{ m}^2/\text{Hz}$  to -123 dB,  $10^{-12} \text{ m}^2/\text{Hz}$  to -120 dB. In these units we obtain:

At a period of 30 s, the NLNM is at -157 dB while QC-AZ is at -163 dB

At a period of 300 s, the NLNM is at -120 dB while QC-AZ is at -100 dB

At a period of 3000 s, the NLNM is at -59 dB while QC-AZ is at -17 dB

Obviously, the authors of figure 1 were able to resolve minimum ground noise at 30 s but not at much longer periods where their instruments were too noisy.

c) Half-octave band around 1.2 Hz (0.83 s): QC-AZ has a displacement power density of  $10^{-20} \text{ m}^2/\text{Hz}$ . Multiply by the fourth power of the angular frequency,  $2\pi \times 1.2$  Hz, to obtain the approximate power density of acceleration,  $3.2 \times 10^{-17} \text{ m}^2/\text{s}^4/\text{Hz}$ . Multiply with the bandwidth of 0.4 Hz to obtain the total power,  $1.3 \times 10^{-17} \text{ m}^2/\text{s}^4$ . Take the square root to obtain the rms amplitude,  $3.6 \times 10^{-9} \text{ m/s}^2$ . The power can also be expressed as  $10 \times \log(1.3 \times 10^{-17}) \text{ dB} = -169 \text{ dB}$  relative to  $1 \text{ m}^2/\text{s}^4$ . One would attribute the same number of decibels to the rms amplitude, understanding that decibels always refer to power. The NLNM has -172 dB.

For the half-octave band around 27 s (0.037 Hz), the bandwidth is 0.011 Hz and the rms acceleration  $4.2 \times 10^{-11} \text{ m/s}^2$ , which can also be expressed as -207 dB. The NLNM has -200 dB.

- Task 8:**
- a) The periods of the microseisms in Figure 2 vary between  $T = 7$  s (for the smaller amplitudes) and  $T = 10$  s (for the largest amplitudes).
  - b) From this upper and lower period follows with Eq. (4.13)  $n \approx 0.5$  octaves or  $m \approx 1/6$  decade, a center frequency of  $f_0 \approx 0.119$  Hz ( $T_0 \approx 8.4$  s) and an RBW of  $\approx 0.36$
  - c) Maximum double trace amplitudes of the microseisms range between about 6 and 3 mm, average about 4.5 mm, corresponding to a “true” *average peak ground amplitude* of about  $5.6 \times 10^{-6}$  m and thus to a displacement  $a_{\text{RMS}} \approx 4.5 \times 10^{-6}$  m.
  - d) The acceleration  $a_{\text{RMS}} \approx 2.5 \times 10^{-6}$  m for  $f_0 \approx 0.119$  Hz and  $P_a \approx -98$  dB.
  - e)  $P_a \approx -98$  dB for this microseismic storm is close to the power at the NHNM peak around  $T = 5$  s ( $-96.5$  dB) but about 15 dB higher than the NHNM values at  $T \approx 8$  s. Thus, the record corresponds to a really strong microseism storm.
- Task 9:**
- a) From Figure 3, upper trace, the estimated *average peak amplitude* is about  $2.5 \mu\text{gal}$  and thus  $a_{\text{RMS}}$  about  $2 \times 10^{-8}$  m/s<sup>2</sup>.
  - b) The related upper limit of about  $1/100^{\text{th}}$  of a), i.e.,  $a_{\text{RMS}} < 2 \times 10^{-10}$  m/s<sup>2</sup>.
  - c) The periods of the noise in Figure 3 range between roughly 180 s and 750 s. This corresponds to a bandwidth of about 2 octaves or an RBW of 1.5.
  - d) The  $a_{\text{RMS}}$  for the open valve corresponds to  $-154$  dB, that for the closed valve to  $< -194$  dB.
  - e) Taking into account that Fig. 4.8 was calculated for  $1/6$  decade bandwidth only but the bandwidth of the considered noise signals being about 3 times larger we have to assume an about 5 dB higher noise level in Fig. 4.8. Therefore, for periods  $< 30$  s the barometric pressure noise is surely well below the NLNM when the sensor operates in a vacuum. A higher resolution of the record with the vacuum bell valve closed would be required in order to determine the noise level distance to the NLNM for  $T > 30$  s.

## References

- Brune, J. N., and Oliver, J. (1959). The seismic noise of the Earth's surface. *Bull. Seism. Soc. Am.*, **49**, 349-353.
- Fix, J. E. (1972). Ambient Earth motion in the period range from 0.1 to 2560 sec. *Bull. Seism. Soc. Am.*, **62**, 1753-1760.

# Chapter 5

## Seismic Sensors and their Calibration

(Version August 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch5)

Erhard Wielandt

Formerly Institute of Geophysics, University of Stuttgart;  
Now: Cranachweg 14/1, D-73230 Kirchheim unter Teck, Germany; E-mail: E.WIELANDT@t-online.de

|                                                                                    | <b>page</b> |
|------------------------------------------------------------------------------------|-------------|
| <b>5.1 Overview</b>                                                                | 2           |
| <b>5.2 Basic Theory</b>                                                            | 4           |
| 5.2.1 The complex notation                                                         | 4           |
| 5.2.2 The Laplace transformation                                                   | 5           |
| 5.2.3 The Fourier transformation                                                   | 7           |
| 5.2.4 The impulse response                                                         | 8           |
| 5.2.5 The convolution theorem                                                      | 9           |
| 5.2.6 Specifying a system                                                          | 10          |
| 5.2.7 The mechanical pendulum                                                      | 11          |
| 5.2.8 Transfer functions of pendulums and electromagnetic seismometers             | 12          |
| <b>5.3 Design of seismic sensors</b>                                               | 15          |
| 5.3.1 Pendulum-type seismometers                                                   | 15          |
| 5.3.2 Decreasing the restoring force                                               | 16          |
| 5.3.3 Sensitivity of horizontal seismometers to tilt                               | 18          |
| 5.3.4 Direct effects of barometric pressure                                        | 19          |
| 5.3.5 Effects of temperature                                                       | 20          |
| 5.3.6 Sensitivity to magnetic fields                                               | 20          |
| 5.3.7 The homogeneous triaxial arrangement                                         | 20          |
| 5.3.8 Electromagnetic velocity sensing and damping                                 | 21          |
| 5.3.9 Electronic displacement sensing                                              | 22          |
| 5.3.10 Electrochemical (MET) transducers                                           | 23          |
| <b>5.4 Force-balance accelerometers and seismometers</b>                           | 23          |
| 5.4.1 The force-balance principle                                                  | 23          |
| 5.4.2 Force-balance accelerometers                                                 | 24          |
| 5.4.3 Velocity broadband seismometers                                              | 25          |
| 5.4.4 Other methods of bandwidth extension                                         | 26          |
| <b>5.5 Seismic noise, site selection, installation and instrumental self-noise</b> | 27          |
| 5.5.1 The USGS low-noise model                                                     | 27          |
| 5.5.2 Site selection                                                               | 28          |
| 5.5.3 Seismometer installation                                                     | 29          |
| 5.5.4 Magnetic shielding                                                           | 30          |
| 5.5.5 Instrumental self-noise                                                      | 30          |
| 5.5.6 Self-noise of electromagnetic short-period seismographs                      | 31          |
| 5.5.7 Self-noise of force-balance seismometers                                     | 32          |
| 5.5.8 Coherency analysis                                                           | 33          |
| 5.5.9 Transient disturbances                                                       | 33          |

|            |                                               |    |
|------------|-----------------------------------------------|----|
| <b>5.6</b> | <b>Seismometer Calibration</b>                | 34 |
| 5.6.1      | Electrical and mechanical calibration         | 34 |
| 5.6.2      | General conditions                            | 34 |
| 5.6.3      | Specific procedures for geophones             | 34 |
| 5.6.4      | Calibration with sinewaves (obsolete)         | 36 |
| 5.6.5      | Step response and weight-lift test (obsolete) | 37 |
| 5.6.6      | Calibration with arbitrary signals            | 38 |
| 5.6.7      | Specific procedure for triaxial seismometers  | 40 |
| 5.6.8      | Calibration on a shake table                  | 42 |
| 5.6.9      | Calibration by stepwise motion                | 42 |
| 5.6.10     | Calibration with tilt                         | 44 |
| <b>5.7</b> | <b>Testing for non-linear distortions</b>     | 45 |
| <b>5.8</b> | <b>Free software</b>                          | 46 |
|            | <b>Acknowledgments</b>                        | 46 |
|            | <b>References</b>                             | 47 |

## 5.1 Overview

There are two basic types of seismic sensors: inertial seismometers which measure ground motion relative to an inertial reference (a suspended mass), and strainmeters or extensometers which measure the motion of one point of the ground relative to another. Since the motion of the ground relative to an inertial reference is in most cases much larger than the differential motion within a vault of reasonable dimensions, inertial seismometers are generally more sensitive to earthquake signals. However, at very low frequencies it becomes increasingly difficult to maintain an inertial reference, and for the observation of low-order free oscillations of the Earth, tidal motions, and quasi-static deformations, strainmeters may outperform inertial seismometers. Strainmeters are conceptually simpler than inertial seismometers although their technical realization and installation may be more difficult (see IS 5.1). This Chapter is concerned with inertial seismometers only. For a more comprehensive description of inertial seismometers, recorders and communication equipment see Havskov and Alguacil (2002).

An inertial seismometer converts ground motion into an electric signal but its properties can not be described by a single scale factor, such as output volts per millimeter of ground motion. The response of a seismometer to ground motion depends not only on the amplitude of the ground motion (how large it is) but also on its time scale (how sudden it is). This is because the seismic mass has to be kept in place by a mechanical or electromagnetic restoring force. When the ground motion is slow, the mass will move with the rest of the instrument, and the output signal for a given ground motion will therefore be smaller. The system is thus a high-pass filter for the ground displacement. This must be taken into account when the ground motion is reconstructed from the recorded signal, and is the reason why we have to go to some length in discussing the dynamic transfer properties of seismometers.

The dynamic behavior of a seismograph system within its linear range can, like that of any linear time-invariant (LTI) system, be described with the same degree of completeness in four different ways: by a linear differential equation, the Laplace transfer function (5.2.2), the complex frequency response (5.2.3), or the impulse response of the system (5.2.4). The first two are usually obtained by a mathematical analysis of the physical system (the hardware). The latter two are directly related to certain calibration procedures (5.6.4 and 5.6.5) and can

therefore be determined from calibration experiments where the system is considered as a “black box”(this is sometimes called an identification procedure). However, since all four are mathematically equivalent, we can derive each of them either from a knowledge of the physical components of the system or from a calibration experiment. The mutual relations between the “time-domain” and “frequency-domain” representations are illustrated in Fig. 5.1. Practically, the mathematical description of a seismometer is limited to a certain bandwidth of frequencies that should at least include the bandwidth of seismic signals. Within this limit then any of the four representations describe the system's response to arbitrary input signals completely and unambiguously. The viewpoint from which they differ is how efficiently and accurately they can be implemented in different signal-processing procedures.

In digital signal processing, seismic sensors are often represented with other methods that are efficient and accurate but not mathematically exact, such as recursive (IIR) filters. Digital signal processing is however beyond the scope of this section. A wealth of textbooks is available both on analog and digital signal processing, for example Oppenheim and Willsky (1983) for analog processing, Oppenheim and Schaffer (1975) for digital processing, and Scherbaum (1996, 2007) for seismological applications.

The most commonly used description of a seismograph response in the classical observatory practice has been the “*magnification curve*”, i.e. the frequency-dependent magnification of the ground motion. Mathematically this is the modulus (absolute value) of the complex frequency response, usually called the *amplitude response*. It specifies the steady-state harmonic responsivity (amplification, magnification, conversion factor) of the seismograph as a function of frequency. However, for the correct interpretation of seismograms, also the phase response of the recording system must be known. It can in principle be calculated from the amplitude response, but is normally specified separately, or derived together with the amplitude response from the mathematically more elegant description of the system by its *complex transfer function* or its *complex frequency response*.

While for a purely electrical filter it is usually clear what the amplitude response is - a dimensionless factor by which the amplitude of a sinusoidal input signal is multiplied - the situation is not always as clear for seismometers because different authors may prefer to measure the input signal (the ground motion) in different ways: as a displacement, a velocity, or an acceleration. Both the physical dimension and the mathematical form of the transfer function depend on the definition of the input signal, and one must sometimes guess from the physical dimension to what sort of input signal it applies. The output signal, traditionally a needle deflection, is now normally a voltage, a current, or a number of counts.

Calibrating a seismograph means measuring (and in some cases adjusting) its transfer properties and expressing them as a complex frequency response or one of its mathematical equivalents. For most applications the result must be available as parameters of a mathematical formula, not as raw data; so determining parameters by fitting a theoretical curve of known shape to the data is usually part of the procedure. Practically, seismometers are calibrated in two steps.

The first step is an electrical calibration (5.6.1) in which the seismic mass is excited with an electromagnetic force. Most seismometers have a built-in calibration coil that can be connected to an external signal generator for this purpose. Usually the response of the system to different sinusoidal signals at frequencies across the system's passband (5.6.4), to impulses or steps (5.6.5), or to arbitrary broadband signals (5.6.6) is observed while the absolute



magnification or gain remains unknown. For the exact calibration of sensors with a large dynamic range such as those employed in modern seismograph systems, the latter method is most appropriate. Shake tables are not suitable to measure the response of a seismometer over a large bandwidth.

The second step, the determination of the absolute gain, is more difficult because it requires mechanical test equipment in all but the simplest cases (5.6.3). The most direct method is to calibrate the seismometer on a shake table (5.6.9) or step table (5.6.10). The frequency at which the absolute gain is measured must be chosen so as to minimize noise and systematic errors, and is often predetermined by these conditions within narrow limits. Other mechanical devices such as mechanical balances and machine tools can also provide a suitable mechanical input for an absolute calibration (5.6.10, 5.6.11).

## 5.2 Basic theory

This section introduces some basic concepts of the theory of linear systems. For a more complete and rigorous treatment, the reader should consult a textbook such as by Oppenheim and Willsky (1983). Digital signal processing is based on the same concepts but the mathematical formulations are different for discrete (sampled) signals (see Oppenheim and Schaffer, 1999, 2009; Scherbaum, 1996, 2007; Plešinger et al., 1996). Readers who are familiar with the mathematics may proceed to section 5.3.

### 5.2.1 The complex notation

A fundamental mathematical property of linear time-invariant (LTI) systems such as seismographs (as long as they are not driven out of their linear operating range) is that they do not change the waveform of sinewaves and of exponentially decaying or growing sinewaves. A more abstract mathematical formulation of this statement is that these waveforms are eigenfunctions of the differential operators describing LTI systems. An input signal of the form

$$f(t) = e^{\sigma t} (a_1 \cos \omega t + b_1 \sin \omega t) \quad (5.1)$$

will produce an output signal

$$g(t) = e^{\sigma t} (a_2 \cdot \cos \omega t + b_2 \cdot \sin \omega t) \quad (5.2)$$

with the same  $\sigma$  and  $\omega$ . Note that  $\omega$  is the angular frequency, which is  $2\pi$  times the common frequency. Using Euler's identity

$$e^{j\omega t} = \cos \omega t + j \sin \omega t \quad (5.3)$$

and the rules of complex algebra, we may write our input and output signals as

$$f(t) = \Re[c_1 \cdot e^{(\sigma + j\omega)t}] \text{ and } g(t) = \Re[c_2 \cdot e^{(\sigma + j\omega)t}] \quad (5.4)$$

respectively, where  $\Re[.]$  denotes the real part and  $c_1 = a_1 - jb_1$ ,  $c_2 = a_2 - jb_2$  are complex amplitudes. It can now be seen that the only difference between the input and output signal lies in the amplitude, not in the waveform. The ratio  $c_2 / c_1$  is the complex gain of the system, and for  $\sigma = 0$ , it is the value of the complex frequency response at the angular frequency  $\omega$ . What we have outlined here may be called the engineering approach to complex notation. The sign  $\Re[.]$  for the real part is often omitted but always understood.

The mathematical approach is slightly different in that real signals are not considered to be the real parts of complex signals but the sum of two complex-conjugate signals with positive and negative frequencies:

$$f(t) = c_1 \cdot e^{(\sigma + j\omega)t} + c_1^* \cdot e^{(\sigma - j\omega)t} \quad (5.5)$$

where the asterisk  $*$  denotes the complex conjugate. The mathematical notation is slightly less concise, but since for real signals only the term with  $c_1$  must be explicitly written down (the other one being its complex conjugate), the two notations become very similar. However, the  $c_1$  term describes the whole signal in the engineering convention but only half of the signal in the mathematical notation! This may easily cause confusion, especially in the definition of power spectra. Power spectra computed after the engineer's method (such as the USGS Low Noise Model, see 5.5.1 and Chapter 4) attribute all power to positive frequencies and therefore have twice the power appearing in the mathematical notation.

## 5.2.2 The Laplace transformation

A signal that has a definite beginning in time (such as the seismic waves from an earthquake) can be decomposed into exponentially growing, stationary, or exponentially decaying sinusoidal signals with the *Laplace integral transformation*:

$$f(t) = \frac{1}{2\pi j} \int_{\sigma - j\infty}^{\sigma + j\infty} F(s) e^{st} ds, \quad F(s) = \int_0^\infty f(t) e^{-st} dt. \quad (5.6)$$

The first integral defines the inverse transformation (the synthesis of the given signal) and the second integral the forward transformation (the analysis). It is assumed here that the signal begins at or after the time origin.  $s$  is a complex variable that may assume any value for which the second integral converges; depending on  $f(t)$ , it may not converge when  $s$  has a negative real part. The Laplace transform  $F(s)$  is then said to “exist” for this value of  $s$ . The real parameter  $\sigma$  which defines the path of integration for the inverse transformation (the first integral) can be arbitrarily chosen as long as the path remains on the right side of all singularities of  $F(s)$  in the complex  $s$  plane. This parameter decides whether  $f(t)$  is synthesized from decaying ( $\sigma < 0$ ), stationary ( $\sigma = 0$ ) or growing ( $\sigma > 0$ ) sinusoids. Remember that the mathematical expression  $e^{st}$  with complex  $s$  represents a growing or decaying sinewave, and with imaginary  $s$  a pure sinewave.

The time derivative  $\dot{f}(t)$  has the Laplace transform  $s \cdot F(s)$ , the second derivative  $\ddot{f}(t)$  has  $s^2 \cdot F(s)$ , etc. Suppose now that an analog data-acquisition or data-processing system is characterized by the linear differential equation

$$c_2 \ddot{f}(t) + c_1 \dot{f}(t) + c_0 f(t) = d_2 \ddot{g}(t) + d_1 \dot{g}(t) + d_0 g(t) \quad (5.7)$$

where  $f(t)$  is the input signal,  $g(t)$  is the output signal, and the  $c_i$  and  $d_i$  are constants. We may then subject each term in the equation to a Laplace transformation and obtain

$$c_2 s^2 F(s) + c_1 s F(s) + c_0 F(s) = d_2 s^2 G(s) + d_1 s G(s) + d_0 G(s) \quad (5.8)$$

from which we get

$$G(s) = \frac{c_2 s^2 + c_1 s + c_0}{d_2 s^2 + d_1 s + d_0} F(s) \quad (5.9)$$

We have thus expressed the Laplace transform of the output signal by the Laplace transform of the input signal, multiplied by a known rational function of  $s$ . From this we obtain the output signal itself by an inverse Laplace transformation. This means, we can solve the differential equation by transforming it into an algebraic equation for the Laplace transforms. Of course, this is only practical if we are able to evaluate the integrals analytically, which is the case for a wide range of “mathematical” signals. Real signals must be approximated by suitable mathematical functions for a transformation. The method can obviously be applied to linear and time-invariant differential equations of any order. (Time-invariant means that the properties of the system, and hence the coefficients of the differential equation, do not depend on time.)

The rational function

$$H(s) = \frac{c_2 s^2 + c_1 s + c_0}{d_2 s^2 + d_1 s + d_0} \quad (5.10)$$

is the (Laplace) transfer function of the system described by the differential equation (5.7). It contains the same information on the system as the differential equation itself.

Generally, the transfer function  $H(s)$  of an LTI system is the complex function for which

$$G(s) = H(s) \cdot F(s) \quad (5.11)$$

with  $F(s)$  and  $G(s)$  representing the Laplace transforms of the input and output signals.

A rational function like  $H(s)$  in (5.10), and thus an LTU system, can be characterized up to a constant factor by its poles and zeros. This is discussed in section 5.2.6.

### 5.2.3 The Fourier transformation

Somewhat closer to intuitive understanding but mathematically less general than the Laplace transformation is the Fourier transformation

$$f(t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \tilde{F}(\omega) e^{j\omega t} d\omega, \quad \tilde{F}(\omega) = \int_{-\infty}^{\infty} f(t) e^{-j\omega t} dt \quad (5.12)$$

The signal is here assumed to have a finite energy so that the integrals converge. The condition that no signal is present at negative times can be dropped in this case. The Fourier transformation decomposes the signal into purely harmonic (sinusoidal) waves  $e^{j\omega t}$ . The direct and inverse Fourier transformation are also known as a harmonic analysis and synthesis.

Although the mathematical concepts behind the Fourier and Laplace transformations are different, we may consider the Fourier transformation as a special version of the Laplace transformation for real frequencies, i.e. for  $s = j\omega$ . In fact, by comparison with Eq. (5.6), we see that  $\tilde{F}(\omega) = F(j\omega)$ , i.e. the Fourier transform for real angular frequencies  $\omega$  is identical to the Laplace transform for imaginary  $s = j\omega$ . For practical purposes the two transformations are thus nearly equivalent, and many of the relationships between time signals and their transforms (such as the convolution theorem) are similar or the same for both. The function  $\tilde{F}(\omega)$  is called the complex frequency response of the system. Some authors use the name “transfer function” for  $\tilde{F}(\omega)$  as well; however,  $\tilde{F}(\omega) = F(j\omega)$  is not the same function as  $F(\omega)$ , so a different name are appropriate. The distinction between  $\tilde{F}(\omega)$  and  $F(s)$  is essential when systems are characterized by their poles and zeros. These are equivalent but not identical in the complex  $s$  and  $\omega$  planes, and it is important to know whether the Laplace or Fourier transform is meant. Usually, poles and zeros are given for the Laplace transform. In case of doubt, check the symmetry of the poles and zeros in the complex plane: those of the Laplace transform are symmetric to the real axis as in Figure 1 of exercise EX 5.7 while those of the Fourier transform are symmetric to the imaginary axis.

The absolute value  $|\tilde{F}(\omega)|$  is called the amplitude response, and the phase of  $\tilde{F}(\omega)$  the phase response of the system. Note that amplitude and phase do not form a symmetric pair; however a certain mathematical symmetry (expressed by the Hilbert transformation) exists between the real and imaginary parts of a rational transfer function, and between the phase response and the natural logarithm of the amplitude response.

The definition of the Fourier transformation according to Eq. (5.12) applies to continuous transient signals. For other mathematical representations of a signal, different definitions must be used:

$$f(t) = \sum_{v=-\infty}^{\infty} b_v e^{2\pi j v t / T}, \quad b_v = \frac{1}{T} \int_0^T f(t) e^{-2\pi j v t / T} dt \quad (5.13)$$

for periodic signals  $f(t)$  with a period  $T$ , and

$$f_k = \frac{1}{M} \sum_{l=0}^{M-1} c_l e^{2\pi j k l / M}, \quad c_l = \sum_{k=0}^{M-1} f_k e^{-2\pi j k l / M} \quad (5.14)$$

for time series  $f_k$  consisting of  $M$  equidistant samples (such as digital seismic data). We have written the inverse transform (the synthesis) first in each case. The successive approximation of arbitrary signals by sums of sine waves is demonstrated in the **fourierdemo** program (section 5.8).

The Fourier integral transformation (Eq. 5.12) is mainly an analytical tool; the integrals are not normally evaluated numerically because the discrete Fourier transformation (Eq. 5.14) permits more efficient computations. Eq. (5.13) is the Fourier series expansion of periodic functions, also mainly an analytical tool but also useful to represent periodic test signals. The discrete Fourier transformation (Eq. 5.13) is sometimes considered as being a discretized, approximate version of Eqs. (5.12) or (5.14) but is actually a mathematical tool in its own right: it is a mathematical identity that does not depend on any assumptions on the series  $f_k$ . Its relationship with the other two transformations, and especially the interpretation of the subscript  $l$  as representing a single frequency, do however depend on the properties of the original, continuous signal. The most important condition is that the bandwidth of the signal before sampling must be limited to less than half of the sampling rate  $f_s$ ; otherwise the sampled series will not contain the same information as the original. The bandwidth limit  $f_n = f_s/2$  is called the *Nyquist frequency*. Whether we consider a signal as periodic or as having a finite duration (and thus a finite energy) is to some degree arbitrary since we can analyze real signals only for finite intervals of time, and it is then a matter of definition whether we assume the signal to have a periodic continuation outside the interval or not.

The Fast Fourier Transformation or FFT (Cooley and Tukey, 1965) is a recursive algorithm to compute the sums in Eq. (5.14) efficiently, and does not constitute a mathematically different definition of the discrete Fourier transformation.

### 5.2.4 The impulse response

A useful (although mathematically difficult) fiction is the Dirac “needle” pulse  $\delta(t)$  (e.g. Oppenheim and Willsky, 1983), supposed to be an infinitely short, infinitely high, positive pulse at the time origin whose integral over time equals 1. It can not be realized, but its time-integral, the unit step function, can be approximated by switching a current on or off or by suddenly applying or removing a force. According to the definitions of the Laplace and Fourier transforms, both transforms of the Dirac pulse have the constant value 1. The amplitude spectrum of the Dirac pulse is “white”, this means, it contains all frequencies with equal amplitude. In this case Eq. (5.11) reduces to  $G(s)=H(s)$ . The transfer function  $H(s)$  is thus the Laplace transform of the impulse response  $g(t)$ . Likewise, the complex frequency response is the Fourier transform of the impulse response. All information contained in these complex functions is also contained in the impulse response of the system. The same is true for the step response, which is often used to test or calibrate seismic equipment.

Explicit expressions for the response of a linear system to impulses, steps, ramps and other simple waveforms can be obtained by evaluating the inverse Laplace transform over a suitable contour in the complex  $s$  plane, provided that the poles and zeros are known. The result, generally a sum of decaying complex exponential functions (sinusoids), can then be

numerically evaluated with a computer or even a calculator. Although this is an elegant way of computing the response of a linear system to simple input signals with any desired precision, a warning is necessary: the numerical samples so obtained are not the same as the samples obtained with a digitizer. The digitizer must limit the bandwidth before sampling and therefore does not generate instantaneous samples but some sort of time-averages. For computing samples of band-limited signals, different mathematical concepts must be used (Schuessler, 1981).

Specifying the impulse or step response of a system in place of its transfer function is not practical because the analytic expressions are cumbersome to write down and represent signals of infinite duration that can not be tabulated in full length.

### 5.2.5 The convolution theorem

Any signal may be understood as consisting of a sequence of pulses. This is obvious in the case of sampled signals, but can be generalized to continuous signals by representing the signal as a continuous sequence of Dirac pulses. We may construct the response of a linear system to an arbitrary input signal as a sum over suitably delayed and scaled impulse responses. This process is called a convolution:

$$g(t) = \int_0^{\infty} h(t') f(t-t') dt' = \int_0^{\infty} h(t-t') f(t') dt' \quad (5.15)$$

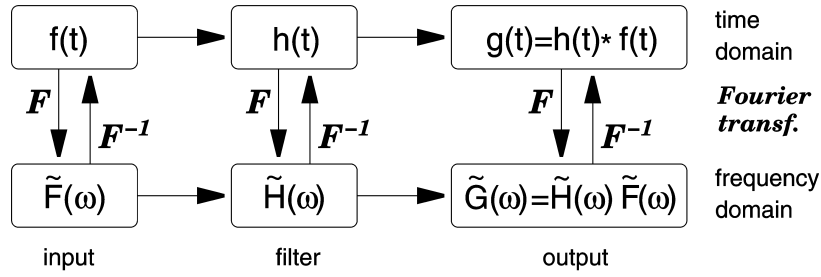
Here  $f(t)$  is the input signal and  $g(t)$  the output signal while  $h(t)$  characterizes the system. We assume that the signals are causal (i.e. zero at negative time), otherwise the integration would have to start at  $-\infty$ . Taking  $f(t) = \delta(t)$ , i.e. using a single impulse as the input, we get

$$g(t) = \int h(t') \delta(t-t') dt' = h(t),$$

so  $h(t)$  is in fact the impulse response of the system.

The response of a linear system to an arbitrary input signal can thus be computed either by convolution of the input signal with the impulse response in time domain, or by multiplication of the Laplace-transformed input signal with the transfer function, or by multiplication of the Fourier-transformed input signal with the complex frequency response in frequency domain.

Since instrument responses are often specified as a function of frequency, the FFT algorithm has become a standard tool to compute output signals. The FFT method assumes, however, that all signals are periodic, and is therefore mathematically inaccurate when this is not the case. Signals must in general be tapered to avoid spurious results. (A taper is a weight function that is zero or small at the beginning and end of the time interval). Fig. 5.1 illustrates the interrelations between signal processing in the time and frequency domains.



**Fig. 5.1** Pathways of signal processing in the time and frequency domains. The asterisk between  $h(t)$  and  $f(t)$  indicates a convolution. An interactive version of this scheme with a number of test signals is available as a BASIC program **filtdemo** (section 5.8). Since the complex frequency response of a layered elastic medium can be expressed by a mathematical formula, this scheme can also be used for computation of synthetic seismograms.

In digital processing, these methods translate into convolving discrete time series or transforming them with the FFT method and multiplying the transforms. For impulse responses with more than 100 samples, the FFT method is usually more efficient. The convolution method is also known as a FIR (finite impulse response) filtration. A third method, the recursive or IIR (infinite impulse response) filtration (e.g. Oppenheim and Schaffer, 2009) is often preferred for its flexibility and efficiency. The design of IIR filters requires special attention because for mathematical reasons they cannot exactly represent rational transfer functions (see the remarks under 5.6.6).

### 5.2.6 Specifying a system

When  $P(s)$  is a polynomial of  $s$  and  $\alpha$  is a specific value of  $s$  for which  $P(\alpha) = 0$ , then  $\alpha$  is called a zero, or a root, of the polynomial. A polynomial of order  $n$  has  $n$  complex zeros  $\alpha_i$ , and can be factorized as  $P(s) = p \cdot \prod (s - s_i)$ . Thus, the zeros of a polynomial together with the constant  $p$  determine the polynomial completely. Since our transfer functions  $H(s)$  are the ratio of two polynomials as in Eq. (5.10), they can be specified by their zeros (the zeros of the numerator  $G(s)$ ), their poles (the zeros of the denominator  $F(s)$ ), and a gain factor (or equivalently the total gain at a given frequency). The whole system, as long as it remains in its linear operating range and does not produce noise, can thus be described by a small number of discrete parameters.

Transfer functions are usually specified according to one of the following concepts:

1. The real coefficients of the polynomials in the numerator and denominator are listed.
2. The denominator polynomial is decomposed into normalized first-order and second-order factors with real coefficients (a total decomposition into first-order factors would require complex coefficients). The factors can in general be attributed to individual modules of the system. They are preferably given in a form from which corner periods and damping coefficients can be read, as in Eqs. (5.23) to (5.25). The numerator often reduces to a gain factor times a power of  $s$ .

3. The poles and zeros of the transfer function are listed together with a gain factor. Poles and zeros must either be real or symmetric to the real axis, as mentioned above. When the numerator polynomial is  $s^m$ , then  $s = 0$  is an  $m$ -fold zero of the transfer function, and the system is a high-pass filter of order  $m$ . (Zeros at nonzero frequency do normally not appear in the transfer function of broadband seismographs because, if they occur mathematically, their effect must practically be cancelled by nearby poles; otherwise the response would not be called broadband.) Depending on the order  $n$  of the denominator and accordingly on the number of poles, the response may be flat at high frequencies ( $n = m$ ), or the system may act as a low-pass filter there ( $n > m$ ). The case  $n < m$  can occur only as an approximation in a limited bandwidth because no practical system can have an unlimited gain at high frequencies.

In the header of the widely used SEED-format data (10.4), the gain factor is split up into a normalization factor bringing the gain to unity at a specified normalization frequency in the passband of the system, and a gain factor representing the actual gain at this frequency. Text versions of dataless SEED headers, named response files or RESP files, can be downloaded for a large number of seismic stations from the IRIS Data Management Center: <http://www.iris.edu/dms/mda.htm>. EX\_5.5 contains an exercise in determining the response from given poles and zeros. An interactive, tutorial program **polzero** in BASIC is available for this purpose (section 5.8). RESP files are normally evaluated under LINUX with the EVALRESP software offered by the IRIS Data Management Center. The first section of a RESP file that describes the seismometer can also be interpreted in a MS-DOS environment with the **winresp** program (section 5.8, **polzero** folder). See also the exercises and worksheets mentioned at the end of section 5.2.

### 5.2.7 The mechanical pendulum

The simplest physical model for an inertial seismometer is a mass-and-spring system with viscous damping (Fig. 5.2).

We assume that the seismic mass is constrained to move along a straight line without rotation (i.e., it performs a pure translation). The mechanical elements are a mass of  $M$  kilograms, a spring with a stiffness  $S$  (measured in Newtons per meter), and a damping element with a constant of viscous friction  $D$  (in Newtons per meter per second). Let the time-dependent ground motion be  $x(t)$ , the absolute motion of the mass  $y(t)$ , and its motion relative to the ground  $z(t) = y(t) - x(t)$ . An acceleration  $\ddot{y}(t)$  of the mass results from any external force  $f(t)$  acting on the mass, and from the forces transmitted by the spring and the damper:

$$M \ddot{y}(t) = f(t) - S z(t) - D \dot{z}(t). \quad (5.16)$$

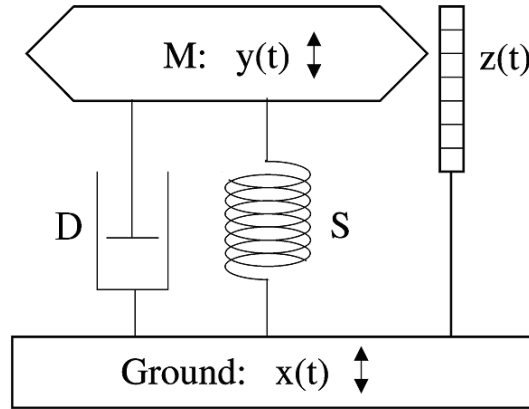
Since we are interested in the relationship between  $z(t)$  and  $x(t)$ , we rearrange this into

$$M \ddot{z}(t) + D \dot{z}(t) + S z(t) = f(t) - M \ddot{x}(t). \quad (5.17)$$

We observe that an acceleration  $\ddot{x}(t)$  of the ground has the same effect as an external force of magnitude  $f(t) = -M \ddot{x}(t)$  acting on the mass in the absence of ground acceleration. We may thus simulate a ground motion  $x(t)$  by applying a force  $-M \ddot{x}(t)$  to the mass while the



ground is not moving. The force is normally generated by sending a current through an electromagnetic transducer, but it may also be applied mechanically.



**Fig. 5.2** Elements of a mechanical harmonic oscillator.

### 5.2.8 Transfer functions of pendulums and electromagnetic seismometers

According to Eqs. (5.7) and (5.8), Eq. (5.17) can be rewritten as

$$(s^2 M + sD + S) Z = F - s^2 M X \quad (5.18)$$

or

$$Z = (F / M - s^2 X) / (s^2 + sD / M + S / M). \quad (5.19)$$

From this we can obtain directly the transfer functions  $T_f = Z/F$  for the external force  $F$  and  $T_d = Z/X$  for the ground displacement  $X$ . We arrive at the same result, expressed by the Fourier-transformed quantities, by simply assuming a time-harmonic motion  $x(t) = \tilde{X}e^{j\omega t} / 2\pi$  as well as a time-harmonic external force  $f(t) = \tilde{F}e^{j\omega t} / 2\pi$ , for which Eq. (5.17) reduces to

$$(-\omega^2 M + j\omega R + S)\tilde{Z} = \tilde{F} + \omega^2 M\tilde{X} \quad (5.20)$$

or

$$\tilde{Z} = (\tilde{F} / M + \omega^2 \tilde{X}) / (-\omega^2 + j\omega R / M + S / M). \quad (5.21)$$

While in mathematical derivations it is convenient to use the angular frequency  $\omega = 2\pi f$  to characterize a sinusoidal signal of frequency  $f$ , and some authors omit the word „angular“ in this context, we reserve the term „frequency“ to the number of cycles per second.

By checking the behavior of  $\tilde{Z}(\omega)$  in the limit of low and high frequencies, we find that the mass-and-spring system is a second-order high-pass filter for displacements and a second-order low-pass filter for accelerations and external forces (see Fig. 5.3). Its corner frequency is  $f_0 = \omega_0 / 2\pi$  with  $\omega_0 = \sqrt{S / M}$ . This is at the same time the „eigenfrequency“ or „natural frequency“ with which the mass oscillates when the damping is negligible. At the angular frequency  $\omega_0$ , the ground motion  $\tilde{X}$  is amplified by a factor  $\omega_0 M / R$  and phase shifted by

$\pi/2$ . The imaginary term in the denominator is usually written as  $2j\omega\omega_0h$  where  $h = D/(2\omega_0M)$  is the numerical damping, i.e., the ratio of the actual to the critical damping.

In order to convert the motion of the mass into an electric signal, the mechanical pendulum in the simplest case is equipped with an electromagnetic velocity transducer (see 5.3.8) whose output voltage we denote with  $\tilde{U}$ . We then have an electromagnetic seismometer (or geophone when designed for seismic exploration). When the responsivity of the transducer is  $E$  (volts per meter per second;  $\tilde{U} = -Ej\omega\tilde{Z}$ ) we get

$$\tilde{U} = -j\omega E(\tilde{F}/M + \omega^2\tilde{X})/(-\omega^2 + 2j\omega\omega_0h + \omega_0^2) \quad (5.22)$$

from which, in the absence of an external force (i.e.  $f(t) = 0$ ,  $\tilde{F} = 0$ ), we obtain the frequency-dependent complex response functions

$$\tilde{H}_d(\omega) := \tilde{U}/\tilde{X} = -j\omega^3 E/(-\omega^2 + 2j\omega\omega_0h + \omega_0^2) \quad (5.23)$$

for the displacement,

$$\tilde{H}_v(\omega) := \tilde{U}/(j\omega\tilde{X}) = -\omega^2 E/(-\omega^2 + 2j\omega\omega_0h + \omega_0^2) \quad (5.24)$$

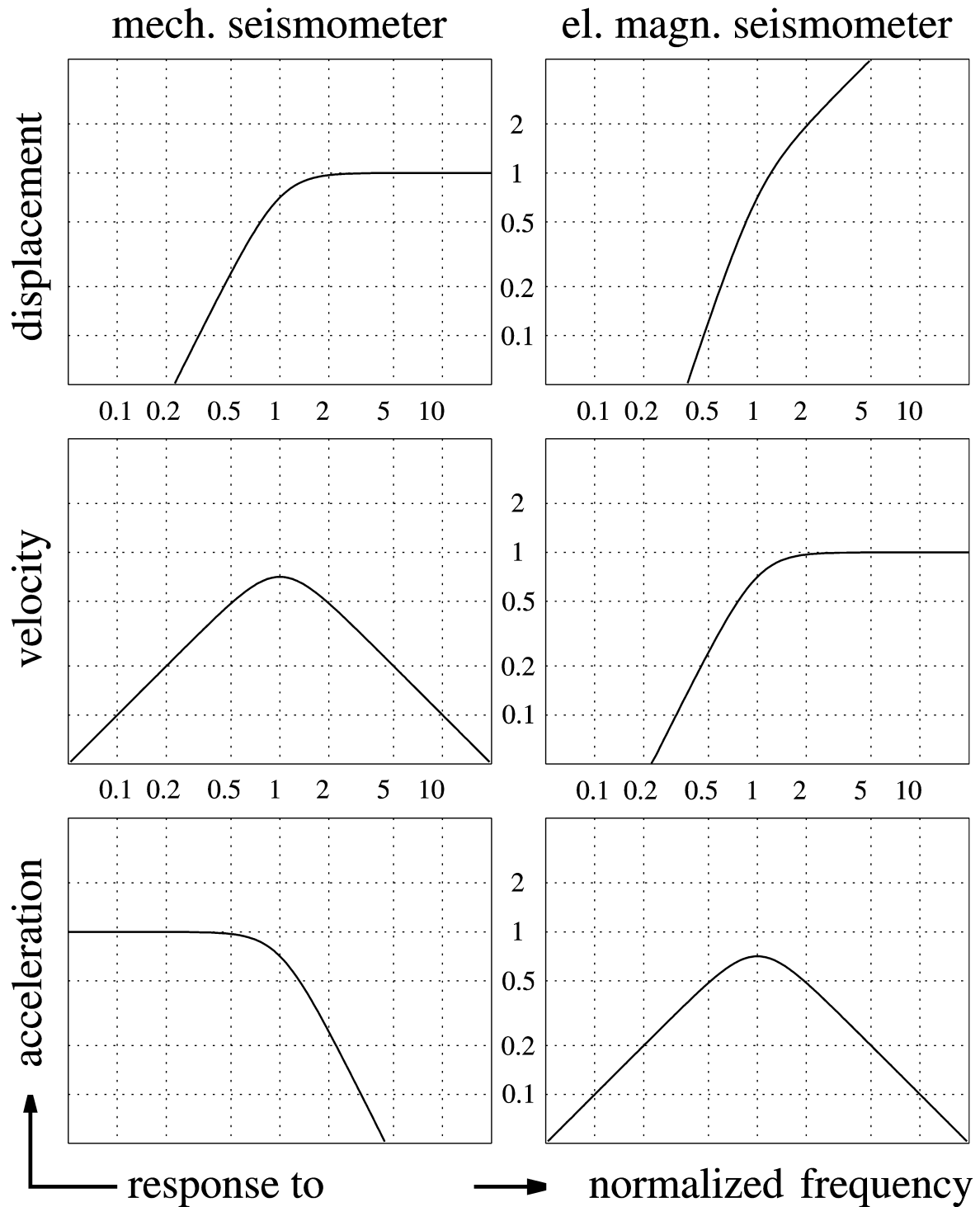
for the velocity, and

$$\tilde{H}_a(\omega) := \tilde{U}/(-\omega^2\tilde{X}) = j\omega E/(-\omega^2 + 2j\omega\omega_0h + \omega_0^2) \quad (5.25)$$

for the acceleration.

With respect to its frequency-dependent response, the electromagnetic seismometer is a second-order high-pass filter for the velocity, and a band-pass filter for the acceleration. Its response to displacement has no flat part and no concise name. These responses (or, more precisely speaking, the corresponding amplitude responses) are illustrated in Fig. 5.3.

The mathematical and graphical representation of the response is the subject of several exercises and information sheets in this manual. EX 5.5 requires different mathematical descriptions of a broadband seismograph and EX 5.6 derives such descriptions for the now historical WWSSN-LP seismograph. IS 5.2 and EX 5.1 by J. Bribach explain the construction of Bode diagrams, a standardized asymptotic representation of the amplitude response.



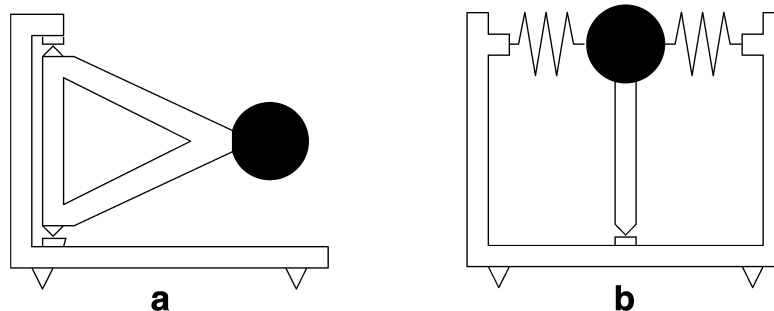
**Fig. 5.3** Response curves of a mechanical seismometer (spring pendulum, left) and electrodynamic seismometer (geophone, right) with respect to different kinds of input signals (displacement, velocity and acceleration). The normalized frequency is the signal frequency divided by the eigenfrequency (corner frequency) of the seismometer. All of these response curves have a second-order corner at the normalized frequency 1. Step responses of second-order high-pass, band-pass and low-pass filters are shown in Fig. 5.23.

## 5.3 Design of seismic sensors

Although the mass-and-spring system of Fig. 5.2 is a useful mathematical model for a seismometer, it is incomplete as a practical design. The suspension must suppress five out of the six degrees of freedom of the seismic mass (three translational and three rotational) but the mass must still move as freely as possible in the remaining direction. This section discusses some of the mechanical concepts by which this can be achieved. In principle it is also possible to let the mass move in all directions and observe its motion with three orthogonally arranged transducers, thus creating a three-component sensor with only one suspended mass. Indeed, some historical instruments have made use of this concept. However, it is difficult to minimize the restoring force and to suppress parasitic rotations of the mass when its translational motion is unconstrained. Modern three-component seismometers therefore have separate mechanical sensors for the three axes of motion.

### 5.3.1 Pendulum-type seismometers

Most seismometers are of the pendulum type, i.e., they let the mass rotate around an axis rather than move along a straight line (Fig. 5.4 to Fig. 5.7). The point bearings in our figures are for illustration only; most seismometers have crossed flexural hinges. Pendulums are not only sensitive to translational but also to angular acceleration. Forbriger (2009) shows however that this sensitivity depends on an arbitrary definition. In order to decompose the motion of the pendulum into a translational and a rotational part, we must define an axis of rotation. When it is properly chosen, the rotational sensitivity disappears. The rotational component of seismic signals is however normally so small that there is no practical difference between linear-motion and pendulum-type seismometers.



**Fig. 5.4** (a) Garden-gate suspension; (b) Inverted pendulum.

For small translational ground motions the equation of motion of a pendulum is formally identical to Eq. (5.17) but  $z$  must then be interpreted as the angle of rotation. Since the rotational counterparts of the constants  $M$ ,  $D$ , and  $S$  in Eq. (5.17) are of little interest in modern electronic seismometers, we will not discuss them further and refer the reader instead to the older literature, such as Berlage (1932) or Willmore (1979).

The simplest example of a pendulum is a mass suspended with a string or wire (like Foucault's pendulum). When the mass has small dimensions compared to the length  $\ell$  of the string so that it can be idealized as a point mass, then the arrangement is called a

mathematical pendulum. Its period of oscillation is  $T = 2\pi\sqrt{\ell/g}$  where  $g$  is the gravitational acceleration. A mathematical pendulum of 1 m length has a period of nearly 2 seconds; for a period of 20 seconds the length has to be 100 m. Clearly, this is not a suitable design for a long-period seismometer.

### 5.3.2 Decreasing the restoring force

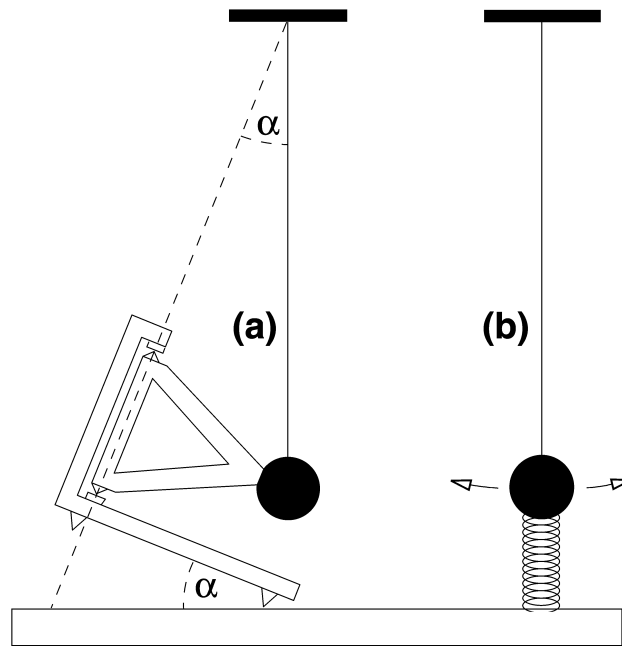
At low frequencies and in the absence of an external force, Eq. (5.17) can be simplified to  $Sz = -M\ddot{x}$  and read as follows: A relative displacement  $-\Delta z$  of the seismic mass indicates a ground acceleration

$$\ddot{x} = (S/M) \Delta z = \omega_0^2 \Delta z = (2\pi/T_0)^2 \Delta z \quad (5.26)$$

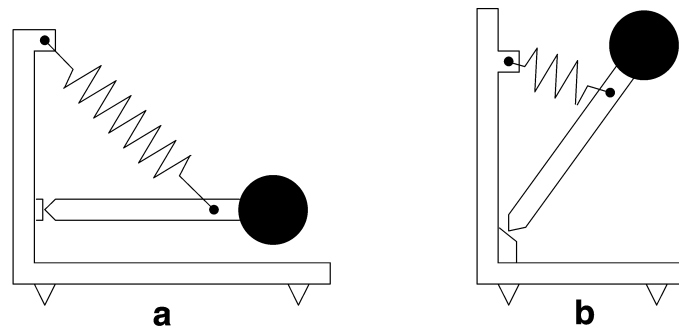
where  $\omega_0$  is the angular eigenfrequency of the pendulum, and  $T_0$  its eigenperiod. If  $\Delta z$  is the smallest mechanical displacement that can be measured electronically, then the formula determines the smallest ground acceleration that can be observed at low frequencies. For a given transducer, it is inversely proportional to the square of the free period of the suspension. A sensitive long-period seismometer therefore requires either a pendulum with a low eigenfrequency or a very sensitive transducer. Since the eigenfrequency of an ordinary pendulum is essentially determined by its size, and seismometers must be reasonably small, astatic suspensions have been invented that combine small overall size with a long free period.

The simplest astatic suspension is the “garden-gate” pendulum used in horizontal seismometers (Fig. 5.4a). The mass moves in a nearly horizontal plane around a nearly vertical axis. Its free period is the same as that of a mass suspended from the point where the plumb line through the mass intersects the axis of rotation (Fig. 5.5a). The eigenperiod  $T_0 = 2\pi\sqrt{\ell/g \sin \alpha}$  is infinite when the axis of rotation is vertical ( $\alpha = 0$ ), and is usually adjusted by tilting the whole instrument. This is one of the earliest designs for long-period horizontal seismometers. Another early design is the inverted pendulum held in stable equilibrium by springs or by a stiff hinge (Fig. 5.4b); a famous example is Wiechert's horizontal pendulum built around 1905.

An astatic spring geometry for vertical seismometers invented by LaCoste (1934) is shown in Fig. 5.6a. The mass is in neutral equilibrium and has therefore an infinite free period when three conditions are met: the spring is pre-stressed to zero length (i.e. the spring force is proportional to the total length of the spring), its end points are seen under a right angle from the hinge, and the mass is balanced in the horizontal position of the boom. A finite free period is obtained by making the angle of the spring slightly smaller than  $90^\circ$ , or by tilting the frame accordingly. By simply rotating the pendulum, astatic suspensions with a horizontal or oblique (Fig. 5.6b) axis of sensitivity can be constructed as well.

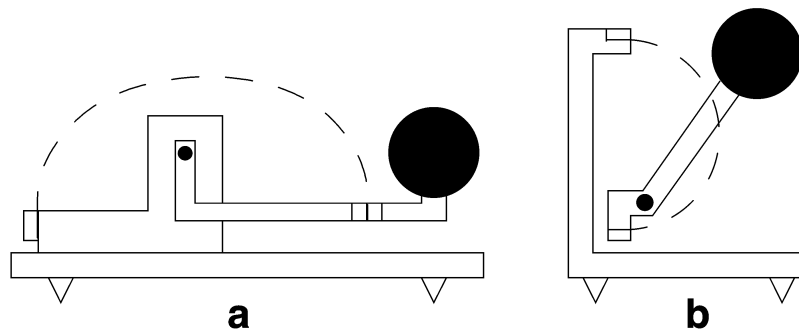


**Fig. 5.5** Equivalence between a tilted “garden-gate” pendulum and a string pendulum. For a free period of 20 s, the string pendulum must be 100 m long. The tilt angle  $\alpha$  of a garden-gate pendulum with the same free period and a length of 30 cm is about  $0.2^\circ$ . The longer the period is made, the less stable it will be under the influence of small tilt changes. (b) Period-lengthening with an auxiliary compressed spring.



**Fig. 5.6** LaCoste suspensions.

The astatic leaf-spring suspension (Fig. 5.7a, Wielandt, 1975), in a limited range around its equilibrium position, is comparable to a LaCoste suspension but is much simpler to manufacture. A similar spring geometry is used in the triaxial seismometer Streckeisen STS2 (Fig. 5.7b, DS 5.1 and Wielandt and Streckeisen, 1982). The delicate equilibrium of forces in astatic suspensions makes them susceptible to external disturbances such as changes in temperature; they are difficult to operate without a stabilizing feedback system.

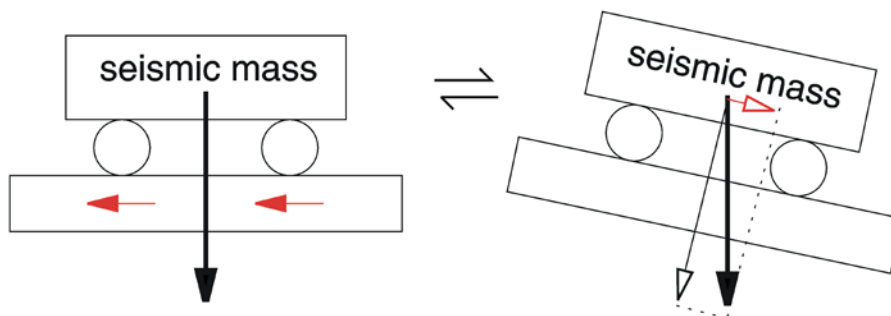


**Fig. 5.7** Leaf-spring astatic suspensions.

Apart from genuinely astatic designs, almost any seismic suspension can be made astatic with an auxiliary spring acting normal to the line of motion of the mass and pushing the mass away from its equilibrium (Fig. 5.5b). The long-period performance of such suspensions, however, is quite limited. Neither the restoring force of the original suspension nor the destabilizing force of the auxiliary spring can be made perfectly linear (i.e. proportional to the displacement). While the linear components of the force may cancel, the nonlinear terms remain and cause the oscillation to become non-harmonic and even unstable at large amplitudes. Viscous and hysteretic behavior of the springs may also cause problems. The additional spring (which has to be soft) may introduce parasitic resonances. Modern seismometers do not use this concept and rely either on a genuinely astatic spring geometry or on the sensitivity of electronic transducers.

### 5.3.3 Sensitivity of horizontal seismometers to tilt

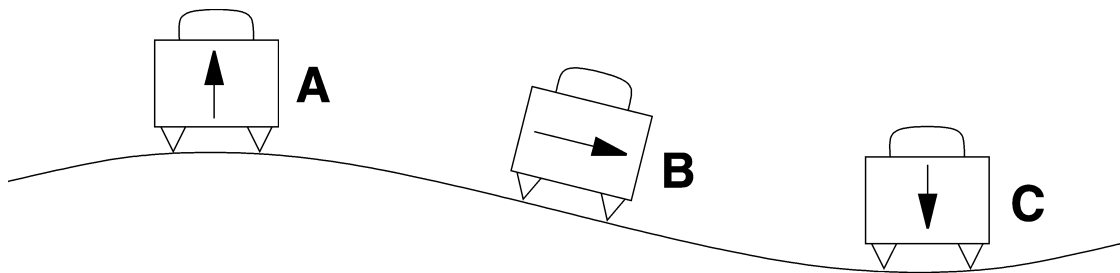
We have already seen (Eq. (5.17)) that a seismic acceleration of the ground has the same effect on the seismic mass as an external force. The largest such force is gravity. It is normally cancelled by the suspension, but when the seismometer is tilted, the projection of the vector of gravity onto the axis of sensitivity changes, producing a force that is in most cases undistinguishable from a seismic signal (Fig. 5.8).



**Fig. 5.8** The relative motion of the seismic mass is the same when the ground is accelerated to the left as when it is tilted to the right.

Undesired tilt at seismic frequencies may be caused by moving or variable surface loads such as cars, people, and atmospheric pressure. The resulting disturbances are a second-order effect in well-adjusted vertical seismometers but otherwise a first-order effect (see Rodgers, 1968; Rodgers, 1969). This explains why horizontal long-period seismic traces are always noisier than vertical ones. A short, impulsive tilt excursion is equivalent to a step-like change of ground velocity and therefore will cause a long-period transient in a horizontal broadband seismometer. For periodic signals, the apparent horizontal displacement associated with a given tilt increases with the square of the period. At tidal and lower frequencies, all horizontal seismometers act as tiltmeters.

Fig. 5.9 illustrates the effect of barometrically induced ground tilt. Let us assume that the ground is vertically deformed by as little  $\pm 1 \mu\text{m}$  over a distance of 3 km, and that this deformation oscillates with a period of 10 minutes. A simple calculation then shows that seismometers A and C see a vertical acceleration of  $\pm 10^{-10} \text{ m/s}^2$  while B sees a horizontal acceleration of  $\pm 10^{-8} \text{ m/s}^2$ . The horizontal noise is thus 100 times larger than the vertical one. In absolute terms, even the vertical acceleration is by a factor of four above the minimum ground noise in one octave as specified by the USGS Low Noise Model (see 5.5.1)



**Fig. 5.9** Ground tilt caused by the atmospheric pressure is the main source of very-long-period noise on horizontal seismographs.

### 5.3.4 Direct effects of barometric pressure

Besides tilting the ground, the continuously fluctuating barometric pressure affects seismometers in at least three different ways: (1) when the seismometer is not enclosed in a hermetic housing, the mass will experience a variable buoyancy which can cause large disturbances in vertical sensors; (2) changes of pressure also produce adiabatic changes of temperature which affect the suspension (see the next subsection). Both effects can be greatly reduced by making the housing airtight or installing the sensor inside an external pressure jacket; however, then (3) the housing or jacket may be deformed by the pressure and these deformations may be transmitted to the seismic suspension as stress or tilt. While it is always worthwhile to protect vertical long-period seismometers from changes of the barometric pressure, it has often been found that horizontal long-period seismometers are less sensitive to barometric noise when they are not hermetically sealed. This, however, may cause other problems such as corrosion.



### 5.3.5 Effects of temperature

The equilibrium between gravity and the spring force in a vertical seismometer is disturbed when the temperature changes. Although thermally self-compensated alloys are available for springs, such a spring does not make a compensated seismometer. The geometry of the whole suspension changes with temperature; the seismometer must therefore be compensated as a whole. However, the different time constants involved prevent an efficient compensation at seismic frequencies. Short-term changes of temperature, therefore, must be suppressed by the combination of thermal insulation and thermal inertia. Special caution is required with seismometers where electronic components are enclosed with the mechanical sensor: these instruments heat themselves up when insulated and are then very sensitive to air drafts, so the insulation must at the same time suppress any possible air convection (5.5.3). Long-term (seasonal) changes of temperature do not interfere with the seismic signal (except when they cause convection in the vault) but may drive the seismic mass out of its operating range. Eq. (5.26) can be used to calculate the thermal drift of a vertical seismometer when the temperature coefficient of the spring force is formally assigned to the gravitational acceleration.

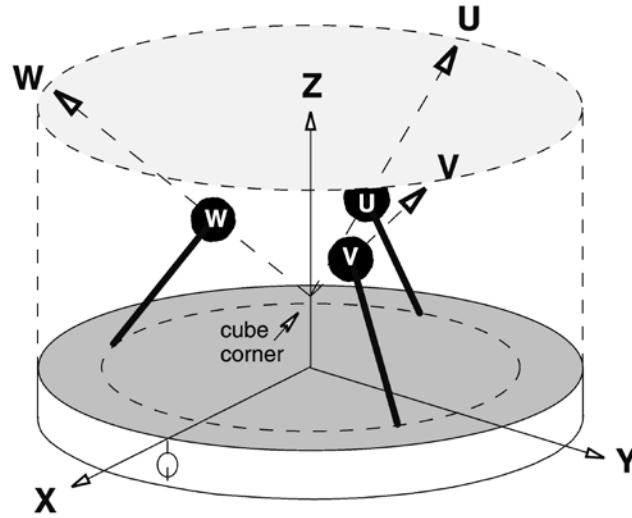
### 5.3.6 Sensitivity to magnetic fields

Broadband seismometers are to some degree sensitive to magnetic fields because all thermally self-compensated spring materials are magnetic. This may be noticeable when seismometers are operated in industrial areas or in the vicinity of dc-powered railway lines. Magnetic interferences by trains must especially be suspected when the long-period noise follows a regular timetable. Magnetic storms have frequently been seen in seismograms. At a very quiet site, the natural background variations of the geomagnetic field may limit the long-period resolution of a vertical sensor when its magnetic sensitivity exceeds  $0.5 \text{ m/s}^2$  per Tesla (Forbriger 2007, Forbriger et al. 2010). It is apparently difficult for manufacturers to avoid this level of magnetic sensitivity. Seismometers can also accidentally acquire a remanent magnetization during transportation or installation. Magnetic shielding (see 5.5.4 and IS 5.4) is therefore recommended at quiet sites.

### 5.3.7 The homogeneous triaxial arrangement

In order to observe ground motion in all directions, a triple set of seismometers oriented towards North, East, and upward (Z) has been the standard for a century. However, horizontal and vertical seismometers differ in their construction, and it takes some effort to make their responses equal. An alternative way of manufacturing a three-component set is to use three sensors of identical construction whose sensitive axes are inclined against the vertical like the edges of a cube standing on its corner (Fig. 5.10), by an angle of  $\arctan \sqrt{2}$ , or 54.7 degrees. A technical advantage of this concept is that the spring has to support only a fraction of the pendulum's weight, so the spring can be lighter and have a higher parasitic resonance.

The homogeneous-triaxial geometry was, with different intentions, introduced by Gal'perin (1955, 1977), Knothe (1963), and Melton and Kirkpatrick (1970), and is presently used in the Streckeisen STS2 and Nanometrics Trillium broadband seismometers.



**Fig. 5.10** The homogeneous triaxial geometry of the STS2 seismometer

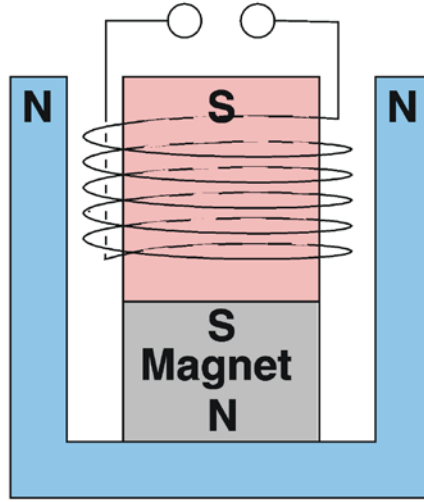
Since most seismologists want finally to see the conventional E, N and Z components of motion, the oblique components U, V, W of the STS2 are electrically recombined according to

$$\begin{pmatrix} X \\ Y \\ Z \end{pmatrix} = \frac{1}{\sqrt{6}} \begin{pmatrix} -2 & 1 & 1 \\ 0 & \sqrt{3} & -\sqrt{3} \\ \sqrt{2} & \sqrt{2} & \sqrt{2} \end{pmatrix} \begin{pmatrix} U \\ V \\ W \end{pmatrix}. \quad (5.27)$$

The X axis of the STS2 seismometer is normally oriented towards East; the Y axis then points North. Noise originating in one of the sensors of a triaxial seismometer will appear on all three outputs (except for Y being independent of U). Its origin can be traced by transforming the X, Y and Z signals back to U, V and W with the inverse (transposed) matrix (programs **triax** and **rectax** (**lincomb** folder, section 5.8). Disturbances affecting only the horizontal outputs are unlikely to originate in the seismometer and are, in general, due to tilt. Disturbances of the vertical output only may be related to temperature, barometric pressure, or electrical problems common to all three sensors such as an unstable supply voltage.

### 5.3.8 Electromagnetic velocity sensing and damping

The simplest transducer both for sensing motions and for exerting forces is an electromagnetic (electrodynamic) device where a coil moves in the field of a permanent magnet, as in a loudspeaker (Fig. 5.11). The motion induces a voltage in the coil; a current flowing in the coil produces a force. From the conservation of energy it follows that the responsivity of the coil-magnet system as a force transducer, in Newtons per Ampere, and its responsivity as a velocity transducer, in Volts per meter per second, are identical. The units are in fact the same (remember that  $1\text{Nm} = 1\text{Joule} = 1\text{VAs}$ ). When such a velocity transducer is loaded with a resistor, permitting a current to flow, then according to Lenz's law it generates a force, opposing the motion. This effect is used to damp the mechanical free oscillation of passive seismic sensors (geophones and electromagnetic seismometers).



**Fig. 5.11** Electromagnetic velocity and force transducer.

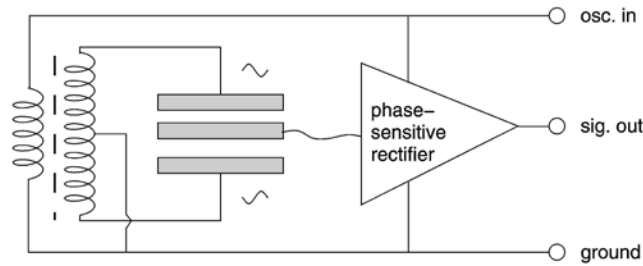
We have so far treated the damping of passive sensors as if it were a viscous effect in the mechanical receiver. Actually, only a small part  $h_m$  of the damping is due to mechanical causes. The main contribution normally comes from the electromagnetic transducer, which is suitably shunted for this purpose. Its contribution is

$$h_{el} = E^2 / 2M\omega_0 R_d \quad (5.28)$$

where  $R_d$  is the total damping resistance (the sum of the resistances of the coil and of the external shunt). The total damping  $h_m + h_{el}$  is preferably chosen as  $1/\sqrt{2}$ , a value that defines a second-order Butterworth filter characteristic, and gives a maximally flat response in the passband (such as the velocity-response of the electromagnetic seismometer in Fig. 5.3).

### 5.3.9 Electronic displacement sensing

At very low frequencies, the output signal of electromagnetic transducers becomes too small to be useful for seismic sensing. One then uses active electronic transducers where a carrier signal, usually in the audio frequency range, is modulated by the motion of the seismic mass. The basic modulating device is an inductive or capacitive half-bridge. Inductive half-bridges are detuned by a movable magnetic core. They require no electric connections to the moving part and are environmentally robust; however their sensitivity appears to be limited by the granular nature of magnetism to something like  $10^{-10}$  m. Capacitive half-bridges (Fig. 5.12) are realized as three-plate capacitors where either the central plate or the outer plates move with the seismic mass. Their resolution is limited by the ratio between the electrical noise of the demodulator and the electrical field strength, and is, for modern broadband seismometers, typically better than  $10^{-12}$  m in the short-period teleseismic band. The comprehensive paper by Jones and Richards (1973) on the design of capacitive transducers still represents state-of-the-art in all essential aspects.



**Fig. 5.12** Capacitive displacement transducer (Blumlein bridge).

### 5.3.10 Electrochemical (MET) transducers

The motion of a liquid electrolyte in a tube can be sensed with fine mesh electrodes through which the liquid flows, by utilizing electrochemical effects at the interface between the electrodes and the liquid. According to one source ([www.mettechnology.com](http://www.mettechnology.com)) such sensors were first developed for the inertial guidance of German rocket weapons in the second world war, then investigated in the US but soon abandoned there, finally developed to practical usefulness in Russia based on theoretical work by V. A. Kozlov and V. Agafonof. The transducers were named Solions (from solution and ion) in the US and Molecular-Electronic (MET) by Russian authors.

In a partly filled circular tube or in a linear tube closed by elastic membranes, the liquid acts as a seismic mass, resulting in mechanically simple and very robust seismic sensors. Their transfer function is however not so simple because hydrodynamic and diffusive processes are involved. A description by poles and zeros as for pendulum-type sensors is mathematically inadequate although it can serve as an approximation. MET seismometers have some practical advantages: they are small and rugged, have a low power consumption, and don't need mass locking, mass centering or leveling. This makes them especially useful for ocean-bottom seismographs (see Chapter 7, section 7.5). They cannot, however, compete with observatory-grade pendulum instruments in other respects: resolution, precision, linearity. Force feedback is difficult to combine with the MET principle.

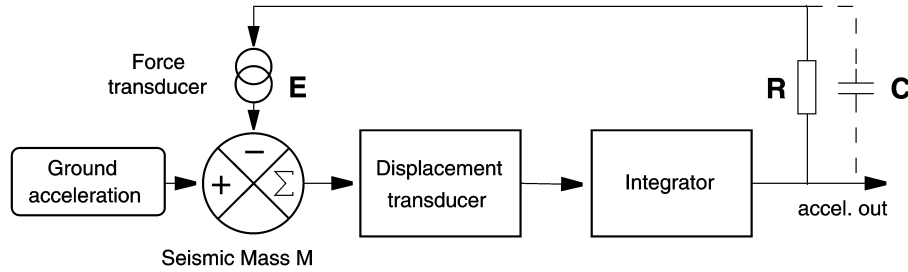
The MET principle is also used in rotational sensors where a circular tube is completely filled with the electrolyte. In this application they appear to be superior to mechanical devices; their symmetric design makes them virtually insensitive to linear acceleration. Rotational components of ground motion have been observed with MET sensors in the near-field of seismic sources, and with costly laser-gyroscopic devices at teleseismic distances from large earthquakes (see IS 5.3).

## 5.4 Force-balance accelerometers and seismometers

### 5.4.1 The force-balance principle

In a conventional passive seismometer, the inertia of the mass makes it move against the frame when the frame is accelerated, and the relative displacement or velocity of the mass is then converted into an electric signal. This principle of measurement is now used for short-period seismometers only. Broadband seismometers are built according to the force-balance principle. The inertial force is compensated (or 'balanced') with an electrically generated force

so that the seismic mass follows the motion of the frame; of course some small relative motion must remain because otherwise the inertial force could not be observed. The feedback force is generated with an electromagnetic force transducer or ‘forcer’ (Fig. 5.11). The electronic circuit (Fig. 5.13) is a servo loop, like in an analog chart recorder, and adjusts the feedback force so that the mass follows the motion of the frame.



**Fig. 5.13** Feedback circuit of a force-balance accelerometer (FBA).

The servo loop is most effective when it contains an integrator, in which case the offset of the mass is exactly nulled in the time average. (In a chart recorder, the difference between the input signal and a voltage indicating the pen position, is nulled). Due to unavoidable delays in the feedback loop, force-balance systems have a limited bandwidth; however, at frequencies where they are effective, they generate a feedback force that is proportional to ground acceleration. When the force is proportional to the current in the transducer, then the current, the voltage across the feedback resistor  $R$ , and the output voltage are all proportional to ground acceleration. Thus we have converted the acceleration into an electric signal without depending on the precision of a mechanical suspension.

The response of a force-balance system is approximately inverse to the gain of the feedback path. It can be easily modified by giving the feedback path a frequency-dependent gain. For example, if we make the capacitor  $C$  large so that it determines the feedback current, then the gain of the feedback path increases linearly with frequency and we have a system whose responsivity to acceleration is inverse to frequency and thus flat to velocity over a certain passband. We will look more closely at this option in section 5.4.3.

## 5.4.2 Force-balance accelerometers

Fig. 5.13 without the capacitor  $C$  represents the circuit of a force-balance accelerometer (FBA), a device that is widely used for earthquake strong-motion recording, for measuring tilt, and for inertial navigation. By equating the inertial and the electromagnetic force, it is easily seen that the responsivity (the output voltage per ground acceleration) is

$$U_{out} / \ddot{x} = MR / E \quad (5.29)$$

where  $M$  is the seismic mass,  $R$  the total resistance of the feedback path, and  $E$  the responsivity of the forcer (in  $N/A$ ). The conversion is determined by only three passive components of which the mass is error-free by definition (it defines the inertial reference), the resistor is a nearly ideal component, and the force transducer very precise because the motion is small. Some accelerometers do not have a built-in feedback resistor; the user can insert a

resistor of his own choice and thus select the gain. The responsivity in terms of current per acceleration is simply  $I_{out} / \ddot{x} = M / E$ .

FBA's work down to zero frequency but the servo loop becomes ineffective at some upper corner frequency  $f_0$  (usually a few hundred to a few thousand Hz), above which the arrangement acts like an ordinary inertial displacement sensor. The feedback loop behaves like an additional stiff spring; the response of the FBA sensor corresponds to that of a mechanical pendulum with the eigenfrequency  $f_0$ , as schematically represented in the left panels of Fig. 5.3.

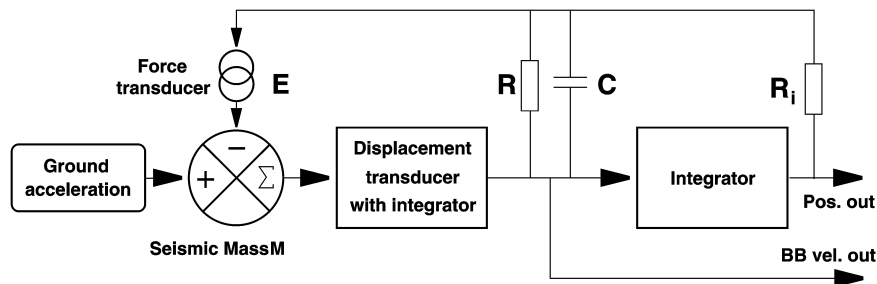
### 5.4.3 Velocity broadband seismometers

For broadband seismic recording with high sensitivity, an output signal proportional to ground acceleration is unfavorable. At high frequencies, sensitive accelerometers are easily saturated by traffic noise or impulsive disturbances. At low frequencies, a system with a response flat to acceleration generates a permanent voltage at the output as soon as the suspension is not completely balanced. The system might then be saturated by the offset voltage resulting from thermal drift or tilt. What we need is a band-pass response in terms of acceleration, or equivalently a high-pass response in terms of ground velocity, like that of a normal electromagnetic seismometer (geophone, right panels in Fig. 5.3) but with a lower corner frequency.

The desired velocity broadband (VBB) response is obtained from the FBA circuit by adding paths for differential feedback and integral feedback (Fig. 5.14). A large capacitor  $C$  is chosen so that the differential feedback dominates throughout the desired passband. While the feedback current is still proportional to ground acceleration as before, the voltage across the capacitor  $C$  is a time integral of the current, and thus proportional to ground velocity. This voltage serves as the output signal. The output voltage per ground velocity, i.e. the apparent generator constant  $E_{app}$  of the feedback seismometer, is

$$E_{app} = V_{out} / \dot{x} = M / EC . \quad (5.30)$$

Again the response is essentially determined by three passive components. Although a capacitor with a solid dielectric is not quite as ideal a component as a good resistor, the response is still linear and very stable.



**Fig. 5.14** Feedback circuit of a VBB (velocity-broadband) seismometer. As in Figure 5.13, the seismic mass is the summing point of the inertial force and the negative feedback force.

The output signal of the second integrator is normally accessible at the „mass position“ output. It does not indicate the actual position of the mass but indicates where the mass would go if the feedback were switched off. „Centering“ the mass of a feedback seismometer has the effect of discharging the integrator so that its full operating range is available for the seismic signal. The mass-position output is not normally used for seismic recording but is useful as a state-of-health diagnostic, and is used in some calibration procedures.

The relative strength of the integral feedback increases at lower frequencies while that of the differential feedback decreases. These two components of the feedback force are of opposite phase ( $-\pi/2$  and  $\pi/2$  relative to the output signal, respectively). At certain low frequency, the two contributions are of equal strength and cancel each other out. This is the lower corner frequency of the closed-loop system. Since the closed-loop response is inverse to that of the feedback path, one would expect to see a resonance in the closed-loop response at this frequency. However, the proportional feedback remains and damps the resonance; the resistor  $R$  acts as a damping resistor. At lower frequencies, the integral feedback dominates over the differential feedback, and the closed-loop response to ground velocity decreases with the square of the frequency. As a result, the feedback system behaves like a conventional electromagnetic seismometer and can be described by the usual three parameters: free period, damping, and generator constant. In fact, electronic broadband seismometers, even if their actual electronic circuit is more complicated than presented here, follow the simple theoretical response of electromagnetic seismometers more closely than those ever did.

As far as the response is concerned, a force-balance circuit as described here may be seen as a means to convert a moderately stable short- to medium-period suspension into a stable electronic long-period or very-long-period seismometer. The corner period may be increased by a large factor, for example 24-fold (from 5 to 120 sec) in the STS2 seismometer or even 200-fold (from 0.6 to 120 sec) in CMG3 and Trillium seismometers. But this factor says little about the performance of the system. Feedback does not reduce the instrumental noise; a large extension of the bandwidth is useless when the system is noisy. According to Eq. (5.26), short-period suspensions must be combined with extremely sensitive transducers for a satisfactory sensitivity at long periods.

At some high frequency, the loop gain falls below unity. This is the upper corner frequency of the feedback system which marks the transition from a response flat to velocity to one flat to displacement. A well-defined and nearly ideal behavior of the seismometer, like at the lower corner frequency, should not be expected here both because the feedback becomes ineffective and because most suspensions have parasitic resonances slightly above the electrical corner frequency (otherwise they could have been designed for a larger bandwidth). The detailed response at the high-frequency corner, however, rarely matters since the upper corner frequency is usually outside the passband of the record. Its effect on the transfer function in most cases can be modeled as a small, constant delay (a few milliseconds) over the whole VBB passband.

#### **5.4.4 Other methods of bandwidth extension**

The force-balance principle permits the construction of high-performance, broadband seismic sensors but is not easily applicable to geophone-type sensors because fitting a displacement transducer to these is difficult. Sometimes it is desirable to broaden the response of an existing geophone without a mechanical redesign.

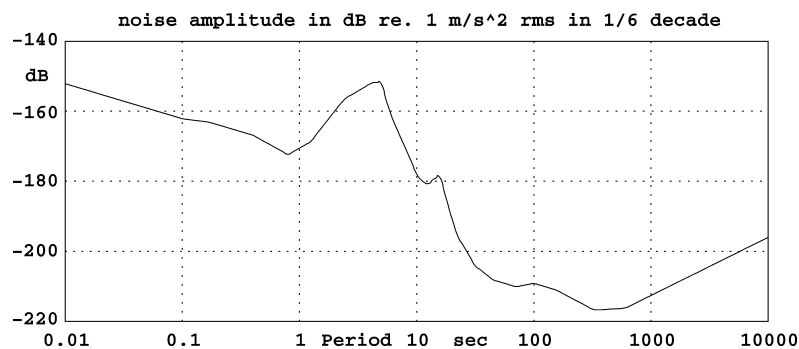
The simplest solution is to send the output signal of the geophone through a filter that removes its original response (this is called an inverse filtration) and replaces it by some other desired response, preferably that of a geophone with a lower eigenfrequency. The analog, electronic version of this process would only be used in connection with direct visible recording; for all other purposes, one would implement the filtration digitally as part of the data processing.

Alternatively, the bandwidth of a geophone may be enlarged by strong damping. This does not enhance the gain outside the passband but rather reduces it at and around the eigenfrequency; nevertheless, after appropriate amplification, the net effect is an extension of the bandwidth towards longer periods. Strong damping is obtained by connecting the coil to a preamplifier whose input impedance is negative. The total damping resistance, which is otherwise limited by the resistance of the coil (Eq. (5.28)), can then be made arbitrarily small. The response of the over-damped geophone is flat to acceleration around its free period. It can be made flat to velocity by an approximate (band-limited) integration. This technique is used in the Lennartz Le-1d and Le-3d seismometers (see DS 5.1) whose electronic corner period can be up to 40 times larger than the mechanical one. Although these are not strictly force-balance sensors, they take advantage of the fact that active damping (which is a form of negative feedback) greatly reduces the relative motion of the mass.

## 5.5 Seismic noise, site selection and installation

Electronic seismographs can be designed for any desired magnification of the ground motion. A practical limit, however, is imposed by the presence of undesired signals which must not be magnified so strongly as to obscure the record. Such signals are usually referred to as noise and may be of seismic, instrumental, or environmental origin. Seismic noise is treated in Chapter 4; see also exercise EX\_4.1. Instrumental self-noise may have mechanical and electronic sources and will be discussed in the next section. Here we focus on those general aspects of site selection and of seismometer installation aimed at the reduction of environmental noise. For technical details on site selection as well as vault, tunnel and borehole installations see Chapter 7. More recommendations for proper deployment and shielding of seismometers are summarized in the information sheet IS 5.4.

### 5.5.1 The USGS low-noise model



**Fig. 5.15** The USGS New Low Noise Model (NLNM), here expressed as RMS amplitude of ground acceleration in a constant relative bandwidth of one-sixth decade.



The USGS low-noise model (Peterson, 1993; see Fig. 5.15) is a graphical and numerical representation of the lowest vertical seismic noise levels observed worldwide, and is extremely useful as a reference for the quality of a site or of an instrument. A recent compilation of minimum noise levels by Berger et al. (2004) has essentially confirmed the validity of the NLNM below 5 Hz; at higher frequencies the NLNM appears to be somewhat too low. Origin and properties of seismic noise are discussed in Chapter 4. The **noisecon** program (section 5.8) can be used to convert power spectral densities of the NLNM into other units (such as rms amplitudes in a given bandwidth) and vice versa.

## 5.5.2 Site selection

Site selection for a permanent station is always a compromise between two conflicting requirements: infrastructure and low seismic noise. The noise level depends on the geological situation and on the proximity of sources, some of which are usually associated with the infrastructure. A seismograph installed on solid basement rock can be expected to be fairly insensitive to local disturbances while one sitting on a thick layer of soft sediments will be noisy even in the absence of identifiable sources. As a rule, the distance from potential sources of noise, such as roads and inhabited houses, should be very much larger than the thickness of the sediment layer. Broadband seismographs can be successfully operated in major cities when the geology is favorable; in unfavorable situations, such as in sedimentary basins, only deep mines and boreholes may offer acceptable noise levels (see 4.3.2, 7.4.3 and 7.4.5).

By definition of the Low Noise Model, most sites have a noise level above the NLNM, sometimes by a large factor. This factor, however, is not uniform over time or over the seismic frequency band. At short periods ( $< 2$  s), a noise level within a factor of 10 of the NLNM may be considered very good in most areas. Short-period noise at most sites is predominantly man-made, lower during nighttime, and somewhat larger in the horizontal components than in the vertical. At intermediate periods (2 to 20 s), marine microseisms dominate. They have similar amplitudes in the horizontal and vertical components and have large seasonal variations. In winter they may be 50 dB above the NLNM. At longer periods, the vertical ground noise is often within 10 or 20 dB of the NLNM even at otherwise noisy stations. Horizontal long-period noise may nevertheless be horrible at the same station due to tilt-gravity coupling (5.3.3). It may be larger than vertical noise by a factor of up to 300, the factor increasing with period. Therefore, a site can be considered as favourable when the horizontal noise at 100 to 300 s is within 20 dB (i.e., a factor of 10 in amplitude) above the vertical noise. Tilt may be caused by traffic, wind, or local fluctuations of the barometric pressure. Large tilt noise is sometimes observed on concrete floors when an unventilated cavity exists underneath; the floor then acts like a membrane. Such noise can be identified by its linear polarization and its correlation with the barometric pressure. Even on an apparently solid foundation, the long-period noise often correlates with the barometric pressure (see Beauduin et al., 1996). If the situation can not be remedied otherwise, the barometric pressure should be recorded with the seismic signal and used for a correction. An example of barometric noise is shown in Fig. 2.21 of Chapter 2. For very-broadband seismographic stations, barometric recording is generally recommended.

Besides ground noise, environmental conditions must be considered. An aggressive atmosphere may cause corrosion, wind and short-term variations of temperature may induce

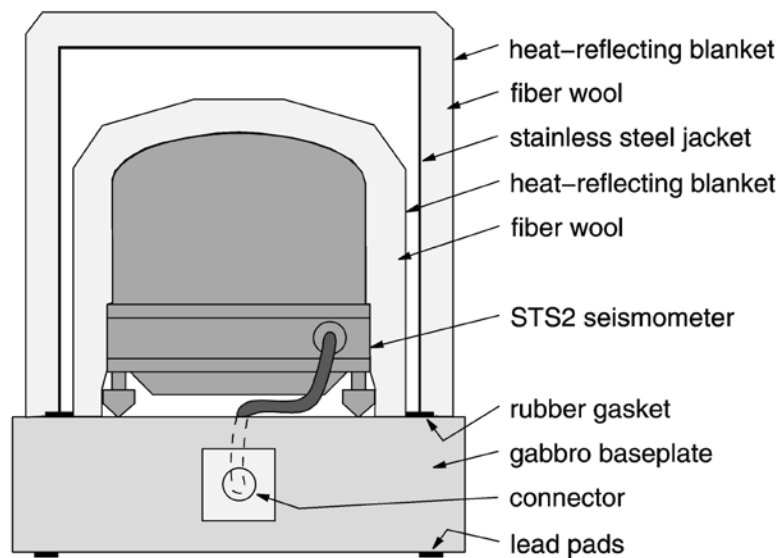
noise, and seasonal variations of temperature may exceed the manufacturer's specifications for unattended operation. Seismometers must be protected against these conditions, sometimes by hermetic containers as described in the next subsection. Suggestions for vault design have been given by Uhrhammer and Karavas (1997) and Trnkoczy (1998) and more recently by the PASSCAL Instrument Center (2009a). Since it is difficult to prevent water from accumulating in vaults, installation of a drainage or a sump pump should be considered (Passcal Instrument Center 2009b).

### 5.5.3 Seismometer installation

We briefly describe the installation of a portable broadband seismometer inside a building, vault, or cave. First, the orientation of the sensor is marked on the floor. This is best done with a geodetic gyroscope, but a magnetic compass will do in most cases. The magnetic declination must be taken into account. Since a compass may be deflected inside a building, the direction should be taken outside and transferred to the site of installation. Spirit levels combined with a laser, and especially laser cross levels, are most convenient tools for orienting seismometers, and are available at low cost from do-it-yourself stores. When the magnetic declination is unknown or unpredictable (such as at high latitudes or in volcanic areas), the orientation should be determined with a sun compass.

To isolate the seismometer from stray currents, small glass or Plexiglas plates should be cemented to the ground under its feet. Then the seismometer is installed, tested, and wrapped with a layer of soft, thermally insulating material such as fiber wool or a synthetic fleece blanket. It is essential that the inner heat shield is so soft that it cannot transport substantial forces to the sensor. An additional heat-reflecting blanket (commonly sold as "space blanket" or "rescue blanket") protects the sensor from thermal radiation and air drafts. The thermal shield should also cover the floor around the seismometer (see Figures 1 to 5 in IS 5.4). The use of Styrofoam seeds is not recommended; they have been observed to cause mechanical noise. Stiff shields such as Styrofoam boxes provide additional protection but must not touch the sensor. The self-heating of electronic seismometers can induce convection in any open space inside the insulation; it is therefore important that the insulation leaves no gap around the seismometer, or at most a gap that is only a few millimeters wide.

For a permanent installation under unfavourable environmental conditions, the seismometer should be enclosed in a hermetic container. A problem with such containers (as with all seismometer housings) is, however, that they cause tilt noise when they are deformed by the barometric pressure. Essentially three precautions are possible: (1) either the base-plate is carefully cemented to the floor, or (2) it is made so massive that its deformation is negligible, or (3) a "warp-free" design is used, as described by Holcomb and Hutt (1992) for the STS1 seismometers. Some fresh desiccant (Silica gel) should be placed inside the container, even into the vacuum bell of STS1 seismometers. Cable connectors corrode easily in a humid environment. All extra connectors and auxiliary electronic equipment should therefore be protected with closed (as far as possible) plastic bags and desiccant. Fig. 5.16 illustrates the shielding of the STS2 seismometers (see DS 5.1) in the German Regional Seismic Network (GRSN). For more details see Figures 1 to 7 in IS 5.4. Guidelines for the installation of broadband seismometers can also be found in Uhrhammer et al. (1998) and more specifically for the installation of both weak-motion (short-period and broadband) and strong-motion seismometers on land, in vaults, tunnels, mines and boreholes as well as in the ocean environment in the Sub-chapters 7.4 and 7.5 of this Manual.



**Fig. 5.16** The STS2 seismometer of the GRSN inside its shields.

#### 5.5.4 Magnetic shielding

Magnetic shields can be manufactured from Permalloy (Mu-metal) but they are expensive and of limited efficiency. An active compensation may be preferable. Such a device might consist of a three-component fluxgate magnetometer that senses the field near the seismometer, an electronic driver circuit in which the signal is integrated with a short time constant (a few milliseconds) and amplified, and a three-component set of Helmholtz coils that compensate changes of the magnetic field (see Figures 6 and 7 in IS 5.4). The permanent geomagnetic field should not be compensated; the resulting offsets of the fluxgate outputs can be compensated electrically before the integration, or with a small permanent magnet mounted near the fluxgate.

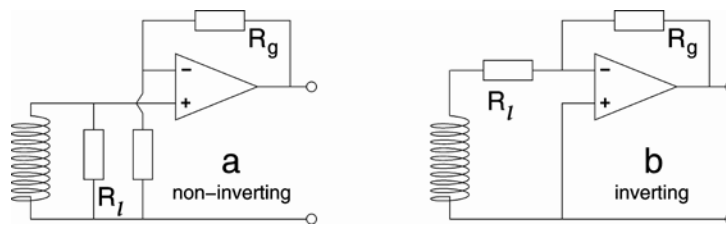
#### 5.5.5 Instrumental self-noise

All modern seismographs use semiconductor amplifiers which, like other active (power-dissipating) electronic components, produce continuous electronic noise whose origin is manifold but ultimately related to the quantization of the electric charge. Electromagnetic transducers, such as those used in geophones, also produce thermal electronic noise (resistor noise, Johnson noise). The contributions from semiconductor noise and resistor noise are often comparable, and together limit the sensitivity of the system. Another source of continuous noise, the Brownian (thermal) motion of the seismic mass, may be noticeable when the mass is very small (less than a few grams). Presently manufactured observatory-grade seismometers have sufficient mass to make the Brownian noise negligible against noise from other sources and we will therefore not discuss it here. Seismographs may also suffer from transient disturbances originating in slightly defective semiconductors or in stressed mechanical parts of the seismometer. The present section is mainly concerned with identifying and measuring instrumental noise.

### 5.5.6 Self-noise of electromagnetic short-period seismographs

Electromagnetic seismometers and geophones are passive sensors whose self-noise is of purely thermal origin and does not increase at low frequencies as it does in active (power-dissipating) devices. Their output signal level, however, is comparatively low, so a low-noise preamplifier (Fig. 5.17) must be inserted between the geophone and the recorder. We will call this combination an electromagnetic seismograph or EMS. Unfortunately the preamplifier noise does increase at low frequencies and limits the overall sensitivity. EMSs are now rarely used for long-period or broadband recording because of the superior performance of feedback instruments.

The sensitivity of an EMS is normally limited by amplifier noise. However, this noise does not depend on the amplifier alone but also on the impedance of the electromagnetic transducer coil (which can be chosen within wide limits). Up to a certain impedance the amplifier noise voltage is nearly constant, but then it increases linearly with the impedance, due to a noise current flowing out of the amplifier input. On the other hand, the signal voltage increases with the square root of the coil impedance. The best signal-to-noise ratio is therefore obtained with an optimum source impedance defined by the corner between voltage and current noise in the graph of total noise vs. source impedance, and is different for each type of amplifier and also depends on frequency. Vice versa, when the transducer is given, the amplifier must be selected for low noise at the relevant impedance and frequency.

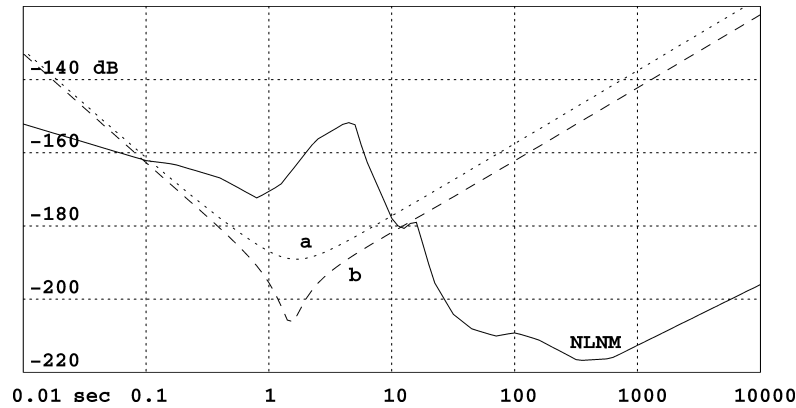


**Fig. 5.17** Two alternative circuits for an EMS preamplifier with a low-noise op-amp. The non-inverting circuit is generally preferable when the damping resistor  $R_I$  is much larger than the coil resistance and the inverting circuit when it is comparable or smaller. However, the relative performance also depends on the noise specifications of the op-amp. The gain is adjusted with  $R_g$ .

The electronic noise of an EMS can be predicted when the technical data of the sensor and the amplifier are known. Semiconductor noise increases at low frequencies; amplifier specifications must apply to seismic rather than audio frequencies. In combination with a given sensor, the noise can then be expressed as an equivalent seismic noise level and compared to real seismic signals or to the NLNM (Fig. 5.15). As an example, Fig. 5.18 shows the self-noise of one of the better seismometer-amplifier combinations. It resolves minimum ground noise between 0.1 and 10 s period. Discussions and more examples are found in Riedesel et al. (1990) and in Rodgers (1992, 1993 and 1994). The result is easily summarized:

Most well-designed seismometer-amplifier combinations resolve minimum ground noise up to 6 or 8 s period, that is, to the microseismic peak. A few of them may make it to about 15 s; they marginally resolve the secondary microseismic peak. To resolve minimum ground noise up to 30 s is hopeless, as is obvious from Fig. 5.18. Ground noise falls and electronic noise

risers so rapidly beyond a period of 20 s that the crossover point can not be substantially moved towards longer periods. Of course, at a reduced level of sensitivity, restoring long-period signals from short-period sensors may make sense, and the long-period surface waves of sufficiently large earthquakes may well be recorded with short-period electromagnetic seismometers.



**Fig. 5.18** Electronic self-noise of the input stage of a short-period seismograph. The EMS is a Sensonics Mk3 with two 8 kOhm coils in series and tuned to a free period of 1.5 s. The amplifier is the LT1012 op-amp. The curves a and b refer to the circuits of Fig. 5.17. NLNM is the USGS New Low Noise Model (Fig. 5.15). The ordinate gives rms noise amplitudes in dB relative to  $1 \text{ m/s}^2$  in 1/6 decade.

Amplifier noise can be observed by locking the sensor or tilting it so that the mass is firmly at a stop, or by replacing it with a resistor that has the same resistance as the coil. If these manipulations do not significantly reduce the noise, then obviously the EMS does not resolve seismic noise. However, this is only a test, not a way to precisely measure the electronic self-noise. A locked sensor or a resistor do not exactly represent the electric impedance of the unlocked sensor.

### 5.5.7 Self-noise of force-balance seismometers

Although the self-noise of force-balance seismometers can theoretically be predicted from that of its components, such a prediction may be unrealistic because certain sources of noise appear only under operating conditions. Anyhow, the user can hardly test the components without destroying the instrument. The electronic circuit cannot be tested when the mass is locked. The instrumental noise can thus only be observed under operating conditions, in the presence of seismic signals and seismic noise.

Although seismic noise is generally a nuisance in this context, natural signals may also be useful as test signals. Marine microseisms should be visible on any sensitive seismograph whose seismometer has a free period of one second or longer; they normally are the strongest continuous signal in a broadband trace. However, their amplitude exhibits large seasonal and geographical variations. For broadband seismographs at quiet sites, the tides of the solid Earth are a reliable and predictable test signal. They have a predominant period of slightly less than 12 hours and an amplitude in the order of  $10^{-6} \text{ m/s}^2$ . While normally invisible in the raw data, they may be extracted by low-pass filtration with a corner frequency of 1 mHz. For this

purpose it is helpful to have the data available with a sampling rate of 1 per second or less. By comparison with the predicted tides, the gain and polarity of the seismograph may be checked (e.g. Davis and Berger 2007). A seismic broadband station that records Earth's tides is likely to be up to international standards.

### 5.5.8 Coherency analysis

For a quantitative determination of instrumental noise, two or three instruments must be operated side by side (Holcomb, 1989, 1990; Sleeman et al. 2006). One can then determine the coherency between the records and assume that coherent noise is seismic and incoherent noise is instrumental. This works well if one has a quiet site and a good reference instrument, but the method is not safe. The seismometers may respond coherently to environmental disturbances caused by barometric pressure, temperature, the supply voltage, magnetic fields, vibrations, or electromagnetic waves. Nonlinear behavior (intermodulation) may produce coherent but spurious long-period signals. When no good reference instrument is available, then different instrument types should be used in the test that are unlikely to respond in the same way to environmental disturbances.

The coherency analysis is somewhat tricky in detail when only two instruments are available. When the transfer functions of both instruments are precisely known, it is in fact theoretically possible to determine the seismic signal and the instrumental noise of each instrument separately as a function of frequency. Alternatively, one may assume that the transfer functions are not so well known but the reference instrument is noise-free; in this case the noise and the relative transfer function of the other instrument can be determined. The coherency test with three instruments after Sleeman et al. (2006) permits to determine the relative transfer functions and the instrumental noise of each instrument at the same time, so it requires no questionable assumptions. It may fail, however, when physically different sources of noise are present such as seismic and magnetic noise to which the instruments do not have a uniform response (they may, for example, have the same response to seismic but not to magnetic noise). As with all statistical methods, very long time series or multiple observations are required for significant results. We offer the computer programs **twocrosp** and **tricrosp** for the analysis (see 5.8, **sleeman** folder).

### 5.5.9 Transient disturbances

Most new seismometers produce spontaneous transient disturbances, quasi miniature earthquakes caused by stresses in the mechanical components. Although they do not necessarily originate in the spring, their waveform at the output seems to indicate a sudden and permanent (step-like) change in the spring force. Long-period seismic records are sometimes severely degraded by such disturbances. The transients often die out within months or years; if they do not, and especially when their frequency increases, corrosion must be suspected. Manufacturers try to mitigate the problem with a low-stress design and by aging the components or the finished seismometer (by extended storage, vibrations, or heating and cooling cycles). It is sometimes possible to release stresses and eliminate transient disturbances by hitting the pier around the seismometer with a hammer, a procedure that is recommended in each new installation.

## 5.6 Seismometer calibration

### 5.6.1 Electrical and mechanical calibration

The calibration of a seismometer establishes knowledge of the relationship between its input (the ground motion) and its output (an electric signal), and is a prerequisite for a reconstruction of the ground motion. Since precisely known ground motions are difficult to generate, one makes use of the equivalence between ground acceleration and an external force on the seismic mass, and calibrates seismometers with an electromagnetic force generated in a calibration coil. This is called an electrical or relative calibration. In the case of feedback seismometers an electrical calibration essentially characterizes the electronic feedback circuit but not the mechanical receiver. Only if the factor of proportionality between the current in the coil and the equivalent ground acceleration is known, the absolute responsivity to ground motion can be determined from an electric calibration. Otherwise, it must be determined from a mechanical experiment in which the seismometer is subject to a known mechanical motion or a tilt. This is called a mechanical or absolute calibration. Since precise mechanical calibration signals are difficult to generate over a large bandwidth, one does not normally attempt to determine the complete transfer function in this way.

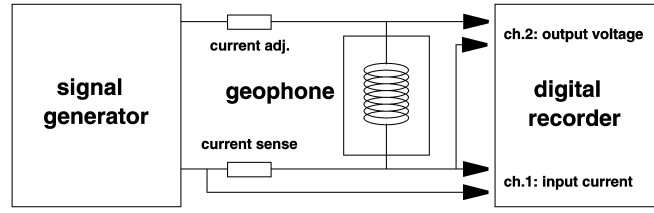
Sub-sections 5.6.2 to 5.6.7 are is mainly concerned with the electrical (relative) calibration although some methods may also be used for the mechanical calibration on a shake table (5.6.9). Procedures for the absolute mechanical calibration that do not require a shake table are presented in in 5.6.10 and 5.6.11.

### 5.6.2 General conditions

Calibration experiments are disturbed by seismic noise and tilt and should therefore be carried out in a basement room. However, the large operating range of modern seismometers permits a calibration with relatively large signal amplitudes, making background noise less of a problem than one might expect. Thermal drift is more serious because it interferes with the long-period response of broadband seismometers. For a calibration at long periods, seismometers must be protected from draft and allowed sufficient time to reach thermal equilibrium. Visible and digital recording in parallel is recommended. Recorders must be absolutely calibrated before they can serve to absolutely calibrate seismometers. **The input impedance of recorders and the source impedance of sensors must be measured so that a correction can be applied for the voltage drop in the source impedance.**

### 5.6.3 Specific procedures for geophones

Geophones usually have no calibration coil. The calibration current must then be sent through the signal coil where it produces an ohmic voltage in addition to the output signal generated by the motion of the mass. The undesired voltage can be compensated in a bridge circuit (Willmore 1959); the bridge is zeroed with the seismic mass locked or at a stop. When the calibration current and the output voltage are digitally recorded, it is more convenient to use only a half-bridge (Fig. 5.19) and to compensate the ohmic voltage numerically. The program **calex** (section 5.8) can do this automatically. The residual (the difference between the actual and the modeled response) is often dominated by nonlinear distortions (5.7).



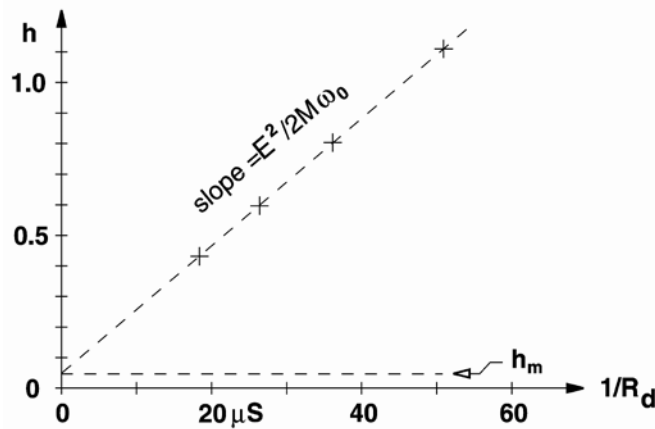
**Fig. 5.19** Half-bridge circuit for calibrating electromagnetic seismometers

Geophones can be absolutely calibrated without a mechanical input provided that the total moving mass  $M$  is known and its motion is linear. In an electric calibration a geophone behaves like a resonant electric circuit. Its electrical impedance is

$$\tilde{Z}(\omega) = R_C + \frac{E^2}{\omega_0 M} \cdot \frac{j\omega\omega_0}{-\omega^2 + 2j\omega\omega_0 h + \omega_0^2} \quad (5.31)$$

$R_C$  is the ohmic resistance of the coil,  $E$  is the generator constant of the electromagnetic transducer as in 5.2.8. The term after the plus sign is the response of a resonant electric circuit consisting of a capacitor  $C=M/E^2$ , an inductor  $L=E^2/S$ , and a resistor  $R_D=E^2/D$  in parallel.  $M$ ,  $S$ , and  $D$  are the mechanical components of the pendulum as in 5.2.7. The analysis of the resonant response, either in time or in frequency domain, supplies all desired parameters when  $M$  is known. For an analysis in time domain, it is convenient to excite a transient response by interrupting a current through the signal coil (Willmore 1979, Rodgers et al. 1995). In this case the ohmic voltage disappears when the transient response begins. **calex** can also be used here.

Another approach is illustrated in fig. 5.20. The electromagnetic part of the numerical damping is inversely proportional to the total damping resistance, the factor of proportionality being  $E^2 / 2M\omega_0$ . The generator constant  $E$  can thus be calculated from relative calibrations with different resistive loads, independent of the method used for the relative calibration.



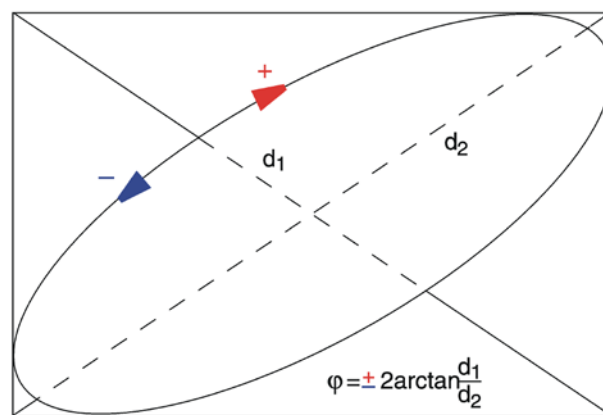
**Fig. 5.20** Determining the generator constant from a plot of damping versus total damping resistance  $R_d = R_{coil} + R_{load}$ . The horizontal units are microsiemens (reciprocal Megohms).



### 5.6.4 Calibration with sinewaves (obsolete)

With a sinusoidal input, the output of a linear system is also sinusoidal, and the ratio of the two signal amplitudes is the absolute value of the transfer function. An experiment with sinewaves therefore permits a direct check of the transfer function, without any a-priori knowledge of its mathematical form and without waveform modeling. This is often the first step in the identification of an unknown system. A computer program would however be required to derive a parametric representation of the response from the measured values. A calibration with arbitrary signals, as described later, is more straightforward for this purpose.

Calibration with sinewaves is time-consuming because the system must reach a steady state after each change of frequency, which can take more than 10 minutes for some broadband seismometers. The gain and phase delay can be read manually from a plotted Lissajous ellipse as in Fig.5.21. The accuracy of the evaluation depends on the purity of the sinewave. A better accuracy is obtained by numerical analysis of digitally recorded data. By fitting sinewaves to the signals, amplitudes and phases can be extracted for just one precisely known frequency at a time; distortions of the input signal don't matter then. If the test frequency is not digitally controlled, then it should be fitted as well. The fit should be computed for an integer number of cycles, and offsets should be removed from the data. We offer a computer program **sinfit** for this purpose (section 5.8, **sinical** folder). Although the method is now obsolete for the purpose of calibration, it is useful for investigating unmodelled details of the response such as parasitic resonances; these might be lost in a more time-efficient broadband calibration. Another surviving application is the calibration of passive short-period seismometers in seismic stations where sinewaves can be remotely applied to the calibration coil but no other test signals are available. The evaluation can be done with a **sinical** program (section 5.8).



**Fig. 5.21** Measuring the phase between two sine-waves with a Lissajous ellipse.

Eigenfrequency  $f_0$  and damping  $h$  of seismometers with a conventional response can be determined graphically with a set of standard resonance curves on double-logarithmic paper. The measured amplitude ratios are plotted as a function of frequency  $f$  on the same type of paper and overlain with the standard curves (Fig. 5.22). The desired quantities can be read directly. The method is simple but not very accurate.

EX 5.3 by J. Bribach and Ch. Teupser can be used for seismometer calibration by harmonic drive.

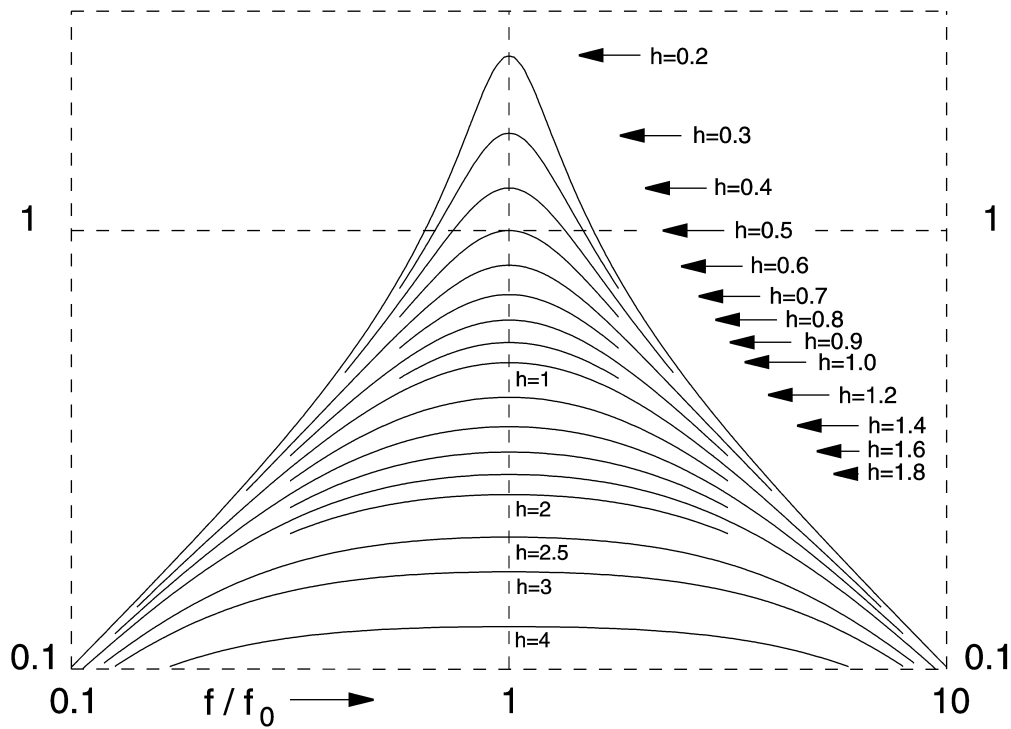


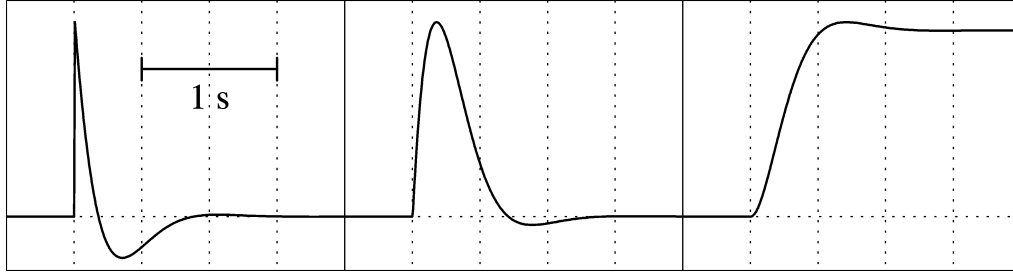
Fig. 5.22 Normalized resonance curves.

### 5.6.5 Step response and weight-lift test (obsolete)

The simplest, but only moderately accurate and now historical, calibration method is to observe the response of the system to a step input. The step signal can be generated by switching on or off a current through the calibration coil, or by applying or removing a constant mechanical force on the seismic mass, usually by lifting a weight. Horizontal sensors used to be absolutely calibrated with a V-shaped thread attached to the mass at one end, to a fixed point at the other end, and to the test weight at half length. The thread was then burned off for a soft release.

The step-response experiment can be used both for a relative and an absolute calibration; when applicable, it is probably the simplest method for the latter. Using a known test weight  $w$  and knowing the seismic mass  $M$ , we also know the test signal: it is a step in acceleration whose magnitude is  $w/M$  times gravity (times a geometry factor when the force is applied through a thread). In case of a rotational pendulum, a correction factor must be applied when the force does not act at the center of gravity. The method has lost its former importance because the seismic mass of modern seismometers is not easily accessible, and the correction factor for rotational motion is not supplied by the manufacturers.

In the context of relative calibration, the step-response method is still useful as a quick and intuitive test, and has the advantage that it can be visually evaluated. The **calex** method covers the step response as well. Fig. 5.23 shows the characteristic step responses of second-order high-pass, band-pass, and low-pass filters with  $1/\sqrt{2}$  of critical damping.



**Fig. 5.23** Normalized step responses of second-order high-pass, band-pass and low-pass filters.

Each response is a strongly damped oscillation around its asymptotic value. With the specified damping, the systems are Butterworth filters, and the amplitude decays to  $e^{-\pi}$  or 4.3% within one half-wave. The ratio of two subsequent amplitudes of opposite polarity is known as the overshoot ratio. It can be evaluated for the numerical damping  $h$ : when  $x_i$  and  $x_{i+n}$  are two (peak-to-peak) amplitudes  $n$  periods apart, with integer or half-integer  $n$ , then

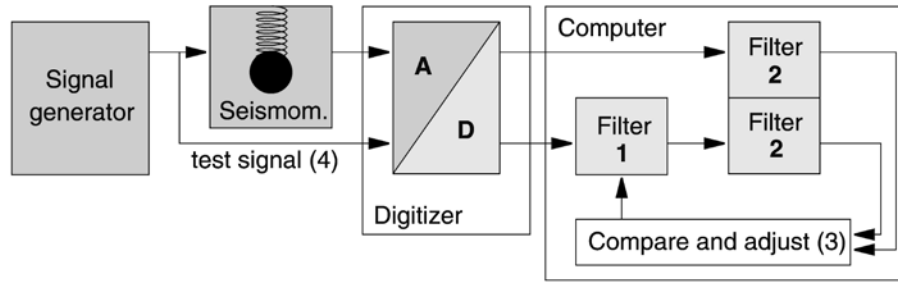
$$\frac{1}{h^2} = 1 + \left( \frac{2\pi n}{\ln x_i - \ln x_{i+n}} \right)^2. \quad (5.32)$$

The free period, in principle, can also be determined from the impulse or step response of the damped system but should be measured preferably without electrical damping so that more oscillations can be observed. A system with the free period  $T_0$  and damping  $h$  oscillates with the period  $T_0 / \sqrt{1 - h^2}$  and the overshoot ratio  $\exp(-\pi h / \sqrt{1 - h^2})$ .

### 5.6.6 Calibration with arbitrary signals

In most cases, the purpose of calibration is to obtain the parameters of an analytic representation of the transfer function. Assuming that its mathematical form is known, the task is to determine the parameters from an experiment in which both the input and the output signals are known. As compared to other methods where a predetermined input signal is used and only the output signal is recorded, recording both signals has the additional advantage of eliminating the transfer function of the recorder from the analysis.

Calibration is a classical inverse problem that can be solved with standard least-squares methods. The general solution is schematically depicted in Fig. 5.24. A computer algorithm (filter 1) is implemented that represents the seismometer as a filter and permits the computation of its response to an arbitrary input. An inversion scheme (3) is programmed around the filter algorithm in order to find best-fitting filter parameters for a given pair of input and output signals. The purpose of filter 2 is explained below. The sensor is then calibrated with a test signal (4) for which the response of the system is sensitive to the unknown parameters but which is otherwise arbitrary. When the system is linear, parameters obtained from one test signal will also predict the response to the other signal.



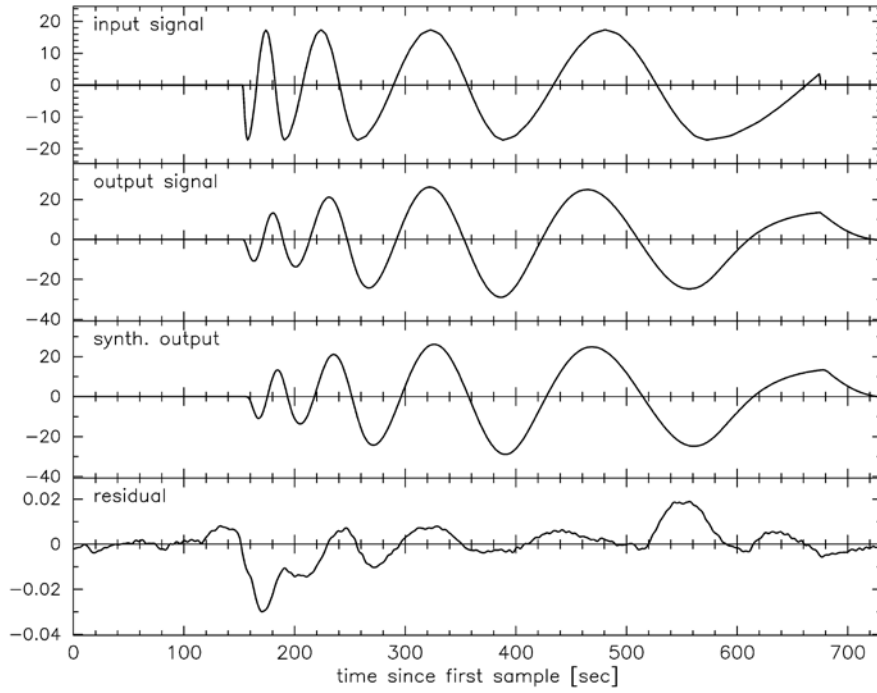
**Fig. 5.24** Block diagram of the **calex** procedure. Storage and retrieval of the data are omitted from the figure.

When the transfer function has been correctly parameterized and the inversion has converged, then the residual error consists mainly of noise, drift, and nonlinear distortions. At a signal level of about one-third of the operating range, typical residuals are 0.03% to 0.05% rms for force- balance seismometers and  $\geq 1\%$  for passive electrodynamic sensors.

The approximation of a rational transfer function with a discrete filtering algorithm is not trivial. The program **calex** (section 5.8) uses an impulse-invariant recursive filter (Schuessler, 1981). This method formally requires that the seismometer has a negligible response at frequencies outside the Nyquist bandwidth of the recorder, a condition that is severely violated by most digital seismographs; but this problem can be circumvented with an additional digital low-pass filtration (Filter 2 in Fig. 5.24) that limits the bandwidth of the simulated system. Signals from a typical calibration experiment are shown in Fig. 5.25. A sweep as a test signal permits the residual error to be visualized as a function of time or frequency. Since essentially only one frequency is present at a time, the time axis may as well be interpreted as a frequency axis.

With an appropriate choice of the test signal, other methods like the calibration with sine-waves step functions, random noise or random telegraph signals, can be duplicated and compared to each other. An advantage of the **calex** algorithm is that it makes no use of special properties of the test signal, such as being sinusoidal, periodic, step-like or random. Therefore, test signals can be short (a few times the free period of the seismometer) and can be generated with the most primitive means, even by hand (you may turn the dial of a sinewave generator by hand, or even produce the test signal with a battery and a switch or potentiometer). A breakout box or a special cable may, however, be required for feeding the calibration signal into the digital recorder.

For a quick and easy check of the transfer function, the simple method of spectral division may be sufficient. When the input signal (the stimulus) and the output signal (the response) have both been recorded, and if the system was quiet at the beginning and the end of the record, then dividing the Fourier transform of the output signal by that of the input signal may result in a reasonable approximation to the actual response, at least in a limited bandwidth. A parametric (mathematical) representation of the response, such as by poles and zeros, is however more difficult to obtain in this way.



**Fig. 5.25** Electrical calibration of an STS2 seismometer with **calex**. Traces from top to bottom: input signal (a sweep with a total duration of 10 min); output signal; synthetic output signal; residual. The rms residual is 0.05 % of the rms output. See also EX 5.4.

Some other routines for seismograph calibration and system identification are contained in the PREPROC software package (Plešinger et al., 1996).

### 5.6.7 Specific procedure for triaxial seismometers

In homogeneous-triaxial seismometers such as the Streckeisen STS2 and the Nanometrics Trillium models, transfer functions in a strict sense can only be attributed to the individual U, V, W sensors, not to the X, Y, Z outputs. Formally, the response of a triaxial seismometer to arbitrary ground motions is described by a nearly diagonal 3 x 3 matrix of transfer functions relating the X, Y, Z output signals to the X, Y, Z ground motions. This is also true for conventional three-component sets if they are not perfectly aligned; only the composition of the matrix is different.

However, if the U, V, W sensors are reasonably well matched, then we need not care about a matrix of transfer functions. The X, Y, Z channels can then each be described by a single transfer function that is a weighted average of those of the U, V, W sensors:

$$\begin{pmatrix} T_X \\ T_Y \\ T_Z \end{pmatrix} = \frac{1}{6} \begin{pmatrix} 4 & 1 & 1 \\ 0 & 3 & 3 \\ 2 & 2 & 2 \end{pmatrix} \begin{pmatrix} T_U \\ T_V \\ T_W \end{pmatrix} \quad (5.33)$$

This formula applies to transfer functions or parameters but not to signals; the matrix elements in (5.33) are the squares of those in 5.27. For any real instrument they will differ slightly from their nominal values but this is negligible in the present context.

The X, Y, Z outputs can thus be calibrated as if they represented single-component sensors. In order to simulate a ground acceleration in one of the X, Y, or Z directions, simultaneous currents must be sent through the U, V, W calibration coils so that an output signal appears only at one of the X, Y, Z outputs. For the Z component this requires equal currents through the U, V, W coils. For the X or Y direction, the three currents must however have different magnitudes and polarities, which requires a slightly more complicated test arrangement. This inconvenience is circumvented by a procedure introduced by Peter Davis (UCSD): calibrate the vertical output with equal currents into the U, V, W coils but record the X and Y output signals as well. Unless all transfer functions happen to be identical, small residual signals will appear at the X and Y outputs. They contain information on the X and Y transfer functions. Use the inverse (transposed) of the signal transformation matrix (5.27) to reconstruct the U, V, W signals, analyze these for their transfer functions, and recombine the results according to (5.33). A computer program **trical**, essentially a combination of **triax** and **calex**, is available for this procedure (section 5.8, **calex** folder). The experiment normally requires four digitizer channels; if only three are available, one must use a predetermined stimulus such as a pseudo-random telegraph signal that can be numerically duplicated.

Using ground noise or other seismic signals, an unknown sensor can be calibrated against a known one by operating the two sensors side by side (Pavlis and Vernon, 1994). The method is limited to a frequency band where suitable seismic signals occur well above the instrumental noise level and are spatially coherent between the two instruments. The instruments must be close to each other on the same pier. The frequency response and the gain of the unknown instrument can be determined at the same time. We offer a program **inverseif** (section 5.8) for this analysis. If the frequency response of both sensors is already known or can be measured electrically, then it will suffice to deconvolve both records to a common response and compare the signal amplitudes in a frequency band where the waveforms are identical.

In a similar way, the orientation of a three-component borehole seismometer may be determined by comparison with a reference instrument at the surface. The mathematical problem can be formulated as follows: for each component  $y_i$  of the borehole seismometer find a set of three directional coefficients  $a_{i1}$ ,  $a_{i2}$ ,  $a_{i3}$  so that the output signal  $y_i$  is best represented by  $y_i = a_{i1} x_1 + a_{i2} x_2 + a_{i3} x_3$  in a least-squares sense, where  $x_1$ ,  $x_2$ ,  $x_3$  are the output signals of the three components of the reference sensor. Almost any seismic signal that is recorded with a good signal-to-noise ratio can be used as a test signal. Instrumental responses must be normalized (deconvolved to a common frequency response) and an additional band-pass filtration is recommended. The 3\*3 matrix  $A = (a_{ik})$  contains information both on the orientation and on the orthogonality of the borehole sensor (assuming proper orientation and orthogonality of the reference sensor). The x and y components can also be interchanged in the formulation of the problem; one then obtains the inverse matrix that is needed to correct the borehole signals. We offer the **linreg3** software (section 5.8, **linregress** folder) for calculating the  $a_{ik}$  coefficients. Commercial packages for linear algebra like MATLAB also offer convenient solutions.

### 5.6.8 Calibration on a shake table

Using a shake table is the most direct way of obtaining an absolute calibration. In practice, however, precision is usually poor outside a frequency band roughly from 0.5 to 5 Hz. At higher frequencies, a shake table loaded with a broadband seismometer may develop parasitic resonances, and inertial forces may cause undesired rotations. At low frequencies, the maximum displacement and thus the signal-to-noise ratio may be insufficient, and the motion may be non-uniform due to friction or roughness in the bearings. Still worse, most shake tables do not produce a purely translational motion but also some tilt. Gravity is then coupled into the seismic signal. Its relative contribution increases with the square of the signal period and causes an intolerable error in the horizontal components at long periods. One might think that a tilt of 10  $\mu$ rad per mm of linear motion should not matter; however, at periods longer than 20 s, such a tilt will cause a larger output signal than the linear motion. At a period of 1 s, the effect of the same tilt would be negligible. Long-period measurements on a horizontal shake table, if possible at all, require extreme care.

Electromagnetic shake tables may have large stray fields. They can be picked up by electromagnetic transducers or electronic circuits in the sensor so strongly that no calibration is possible at any frequency.

Although most calibration methods mentioned in the previous section are applicable on a shake table, the preferred method is to record both the motion of the table (as measured with a displacement transducer) and the output signal of the seismometer, and to analyze these signals by linear modeling with **calex** (section 5.8) or equivalent software. Depending on the definition of active and passive parameters, one might determine only the absolute gain (responsivity, generator constant) or any number of additional parameters of the frequency response. **calex** permits the elimination of tilt effects from a shake-table calibration, under the assumption that the tilt is proportional to the displacement.

### 5.6.9 Calibration by stepwise motion

The movable tables of machine tools like lathes and milling machines, and of mechanical balances, can replace a shake table for the absolute calibration of seismometers. Also, a portable “step table” for seismometer calibration is now commercially available. The idea is to place the seismometer on the table, let it come to equilibrium, then move the table by a known amount and let it rest again. The apparent motion of the frame can then be calculated by inverse filtration of the output signal and compared with the known mechanical displacement. Since the calculation involves triple integrations, offset and drift must be carefully removed from the seismic trace. The main contribution to drift in the apparent horizontal velocity comes from tilt associated with the motion of the table. With the method subsequently described, it is possible to separate the contributions of displacement and tilt from each other so that the displacement can be reconstructed with good accuracy. This method of calibration is most convenient because it uses only normal workshop equipment. The inherent precision of machine tools and the use of relatively large motions eliminate the difficulty of measuring small mechanical displacements. A FORTRAN program **dispcal** is available for the evaluation (section 5.8).

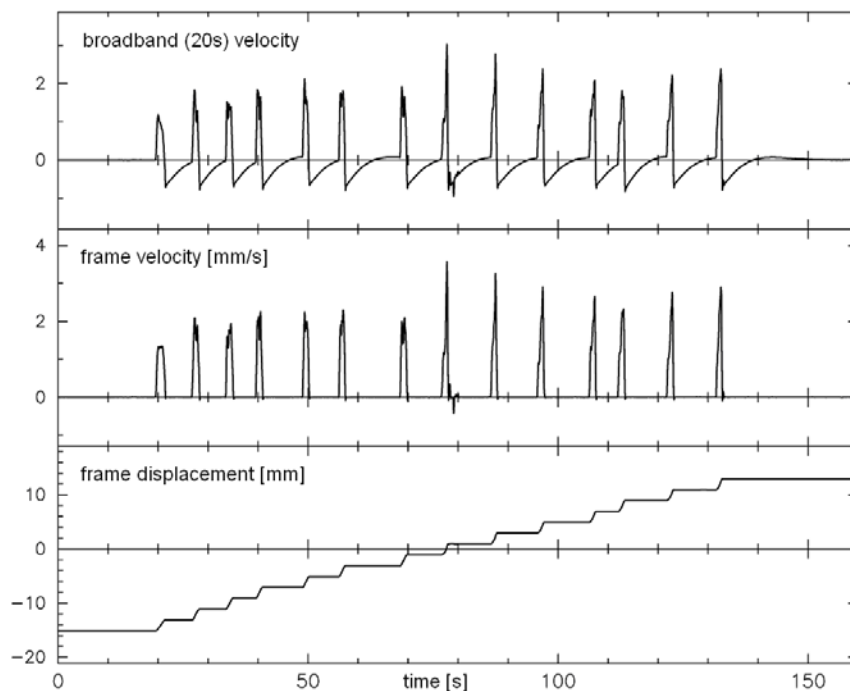
The precision of the method depends on avoiding two main sources of error:

1 - Restoring ground displacement from the seismic signal (a process of inverse filtration) is uncritical for broadband seismometers but requires a precise knowledge of the transfer function for short-period seismometers. Instruments with unstable parameters (such as electromagnetic seismometers) must be electrically calibrated while installed on the test table. However, once the response is known, the restitution of absolute ground motion is no problem even for a geophone with a free period of 0.1 s.

2 - The effect of tilt can only be removed from the displacement signal when the motion is sudden and short. The tilt is unknown during the motion, and since it is equivalent to an acceleration, it produces an unknown offset in the displacement trace that cannot be distinguished from a true displacement. The magnitude of this error signal can however be estimated from the apparent velocity observed when the true motion has ended. In contrast, static tilt before and after the motion produces linear trends in the velocity, which are easily removed.

The computational evaluation consists in the following major steps (Fig. 5.26):

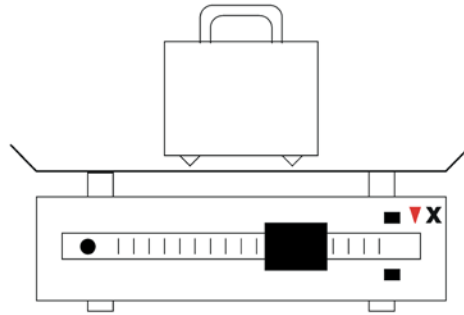
- 1) the trace is de-convolved with the velocity transfer function of the seismometer.
- 2) the trace is piecewise de-trended so that it is close to zero in the motion-free intervals.  
Interpolated trends are removed from the interval of motion.
- 3) the trace is integrated to represent displacement
- 4) the displacement steps are measured and compared to the actual motion.



**Fig. 5.26** Absolute mechanical calibration of an STS1-BB (20s) seismometer on the table of a milling machine, evaluated with DISPCAL. The table was manually moved in 14 steps of 2 mm each (one full turn of the dial at a time). Traces from top to bottom: recorded BB output signal; restored and de-trended frame velocity; restored frame displacement.



In principle, a single step-like displacement is all that is needed. However, the experiment takes so little time that it is convenient to produce a dozen or more equal steps, average the results, and do some error statistics. On a milling machine or lathe, it is recommended to install a mechanical device that stops the motion after each full turn of the spindle. On a balance, the table is repeatedly moved from stop to stop. The displacement may be measured with a micrometer dial or determined from the motion of the beam (Fig. 5.27).



**Fig. 5.27** Calibrating a vertical seismometer on a mechanical balance. When a mass of  $w_1$  grams at some point X near the end of the beam is in balance with  $w_2$  grams on the table or compensated with a corresponding shift of the sliding weight, then the motion of the table is by a factor  $w_1/w_2$  smaller than the motion at X.

From the mutual agreement between different experiments, and from the comparison with shake-table calibrations, the absolute accuracy is estimated to be better than 1%.

EX 5.2 by J. Bribach is an exercise for estimating seismometer parameters by step function.

### 5.6.10 Calibration with tilt

Accelerometers can be statically calibrated on a tilt table. A step table with an additional tilt platform can likewise be used. Starting from a horizontal position, the fraction of gravity coupled into the sensitive axis equals the sine of the tilt angle. (A tilt table is not required for accelerometers with an operating range exceeding  $\pm 1g$ ; these are simply turned upside down.). Force-balance seismometers normally have a mass-position output which is a slowly responding acceleration output. With some patience, this output can likewise be calibrated on a tilt table; the small static tilt range of sensitive broadband seismometers, however, may be inconvenient. The transducer constant of the calibration coil is then obtained by sending a direct current through it and comparing its effect with the tilt calibration. Finally, by exciting the coil with a sine wave whose acceleration equivalent is now known, the absolute calibration of the broadband output is obtained. The method is not explained in more detail here because we propose a simpler method. Anyway, seismometers of the homogeneous-triaxial type cannot be calibrated with static tilt because they do not have X, Y, Z mass-position signals.

The method that we propose (for horizontal components only; program **tiltcal**, section 5.8) is similar to what was described under 5.6.10, but this time we calibrate the seismometer with known steps of tilt, and evaluate the recorded output signal for acceleration rather than displacement.

This is simple: the difference between the drift rates of the de-convolved velocity trace before and after the step equals the tilt-induced acceleration. No baseline interpolation is required. If a tilt platform is not available, one can also tilt the seismometer with a lever under one of its feet, or by pulling out a strip of shim stock. In order to improve the signal-to-noise ratio, it is possible to use a tilt step that exceeds the static operating range of the seismometer. One then has to monitor the output signal and reverse the tilt before the output signal reaches the clipping level.

## 5.7 Testing for non-linear distortions

As we have seen in section 5.2.1, a linear system does not change the waveform of a sinewave. A system that does is said to produce nonlinear distortions. (Linear distortions are those resulting from the frequency-dependent response of a linear system; they affect only waveforms that are not sinusoidal.) A nonlinear system generates spurious signals with frequencies that are multiples (harmonics), sums and differences of the original frequencies. Other than in audio equipment where the high-frequency distortions are most annoying, in seismic recording the distortion (intermodulation) products with low frequencies are most obnoxious because they can cause large errors when signals are inversely filtered or integrated in an attempt to reconstruct ground displacement.

Testing a seismometer for nonlinear distortions, and reporting results properly, is a complex task and one must often be satisfied with a partial answer. Distortions can originate in the mechanical receiver, in the transducers, and in electronic circuits. While for a linear system a current through the calibration coil is unconditionally equivalent to a mechanical acceleration (see 5.2.7), with respect to nonlinear distortion it is not. An electrically linear sensor can still be nonlinear for a seismic input signal. An electrically nonlinear sensor is however unlikely to respond linearly to ground motion, so electrical tests are not useless; but such tests essentially probe the feedback circuit, not the transducers or the mechanical receiver. A serious test for nonlinear distortions therefore requires a nearly sinusoidal mechanical input from a shake table.

Fortunately it turns out that the shake table needs not be more linear than the seismometer if we use the right method. Two methods are available:

1 – The classical two-tone test. The shake table is excited with two superimposed sinewaves of nearly the same frequency, such as 1.00 and 1.02 Hz. Nonlinearity, either of the table or of the sensor, will generate a spurious output signal at the difference (beat) frequency  $\omega_B$ , here 0.02 Hz. It can be separated from the 1 Hz signals by low-pass filtration or Fourier analysis. The contributions from the table and from the seismometer can be separated from each other by repeating the experiment at different beat frequencies (0.05 Hz, 0.02 Hz, 0.01 Hz or even lower). Nonlinearity of the table motion causes an offset of the average table position; the amplitude of the equivalent acceleration at the beat frequency is proportional to  $\omega_B^2$ . Nonlinearity in the seismometer causes a spurious acceleration signal whose amplitude is independent of  $\omega_B$ . Thus, at sufficiently low beat frequencies, seismometer nonlinearity will always predominate. (Whether we see the beat signal or not depends, of course, on the noise level.) Other signal frequencies (2.00 and 2.02 Hz, 5.00 and 5.02 Hz etc.) should also be used because the distortion level depends strongly on the signal frequency.

2 – Linear modeling. The table is excited with a sinewave or a sweep signal. Its motion is measured with a displacement transducer (which may already be there as part of the control electronics) and recorded together with the output signal. With **calex** or an equivalent method, we can then compute a synthetic output signal and compare it to the observed one. The difference (residual, misfit) is composed of seismic noise, nonlinear distortions, and an error signal due to imperfect modeling. Nonlinear distortions can be distinguished from the other contributions because in this experiment their most prominent frequency is twice the input frequency. Although these distortions may not be harmful by themselves, we know that they are always associated with the low-frequency distortions for which the two-tone test was designed. If we want to see these directly, we can duplicate the two-tone test with linear modeling by using a two-tone input signal.

The combination of both methods – linear modeling followed by low-pass filtration of the residual – is especially suitable to detect low-frequency distortions (intermodulation). Modern broadband seismometers typically have mechanical intermodulation ratios around -100 dB in the same units. Electrical intermodulation ratios are typically around -130 dB in the same units. In other units, the dependence on the signal and beat frequencies is normally so strong that it would not be meaningful to quote a typical value. The test procedures outlined here are described in more detail in an USGS Open-File Report (Hutt et al. 2009, p. 21-24; <http://pubs.usgs.gov/of/2009/1295/>).

## 5.8 Free Software

All software mentioned in the text and printed in bold black letters can be downloaded from [HERE](#) or from the summary listing *Download Programs & Files* (see NMSOP-2 content **Overview** on the cover page). A still larger body of software for data analysis, instrument testing, generating synthetic signals, and tutorial purposes is found at

<http://www.geophys.uni-stuttgart.de/~erhard/downloads/>  
<http://www.software-for-seismometry.de/>

Test data are supplied where appropriate. Use the “software overview” on the given websites to select what you need, go to the appropriate software folder, and read the “program descriptions” for details.

A supplementary **calibrat** system, offered by J. Bribach, can also be downloaded via the summary listing. It comprises the programs **response**, **caliseis** and **seisfilt**. A short description of these programs is given in PD 5.1.

## Acknowledgments

This is a revised version (2009 and 2010) of Chapter 5 (Wielandt, 2002a) in the printed first edition of the IASPEI New Manual of Seismological Observatory Practice (NMSOP). For complete citation of this first edition see Bormann, P. (ed.) (2002) under References.

Three careful reviews by the Editor of the NMSOP, Peter Bormann, and suggestions by Axel Plešinger and Jens Havskov have significantly improved the clarity and completeness of this text. A shorter version (Wielandt, 2002b), with some advanced topics added, appeared in part A of the IASPEI International Handbook of Earthquake and Engineering Seismology, ed. W. H. Lee et al. (2002), ISBN-13: 978-0-12-440652-0, ISBN-10: 0-12-440652-1.

## References

- Agnew, D. C., Berger, J., Buland, R., Farrel, W. & Gilbert, F. (1976). International deployment of accelerometers - a network for very long period seismology. *EOS*, **57**(4), 180-188.
- Agnew, D. C., Berger, J., Farrell, W. E., Gilbert, J. F., Masters, G., and Miller, D. (1986). Project IDA: A decade in review. *EOS*, **67**(16), 203-212.
- Anderson, J. A., and Wood, H. O. (1925). Description and theory of the torsion seismometer. *Bull. Seism. Soc. Am.*, **15**(1), 72p.
- Beauduin, R., Lognonné, P., Montagner, J. P., Cacho, S., Karczewski, J. F., and Morand, M. (1996). The effects of the atmospheric pressure changes on seismic signals or How to improve the quality of a station. *Bull. Seism. Soc. Am.*, **86**, 6, 1760-1769.
- Benioff, H., and Press, F. (1958). Progress report on long period seismographs. *Geophy. J. R. astr. Soc.*, **1**(1), 208-215.
- Berger, J., Davis, P., Ekström, G. (2004). Ambient Earth noise: A survey of the Global Seismographic Network. *J. Geophys. Res.*, **109**, B11307, doi: 10.1029/2004JB003408.
- Berlage Jr., H. P. (1932). Seismometer. *Handbuch der Geophysik* (Ed. Gutenberg, B.) *Gebrüder Borntraeger Verlag*, Berlin, Vol. 4, Chapter 4, 299-526.
- Block, B., and Moore, R. D. (1970). Tidal to seismic frequency investigations with a quartz accelerometer of new geometry. *J. Geophys. Res.*, **75**(18), 4361-4375.
- Bormann, P. (ed.) (2002). *IASPEI New Manual of Seismological Observatory Practice*. ISBN 3-9808780-0-7, GeoForschungsZentrum Potsdam, Vol. 1 and 2, 1250 pp. Also: <http://nmsop.gfz-potsdam.de> or <http://www.iaspei.org/projects/NMSOP.html>; doi:10.2312/GFZ.NMSOP\_r1
- Bracewell, R. N. (1978). The Fourier transformation and its applications. *McGraw-Hill*, New York, 2<sup>nd</sup> edn.
- Buttkus, B. (Ed.) (1986). Ten years of the Gräfenberg array: defining the frontiers of broadband seismology. *Geologisches Jahrbuch E-35*, Hannover, 135 pp.
- Cooley, J. S., and Tukey, J. W. (1965). An algorithm for the calculation of complex Fourier Series. *Math. Comp.*, **19**, 297-301.
- Davis, P. and Berger, J. (2007). Calibration of the Global Seismographic Network using tides. *Seism. Res. Lett.*, **78**(4), 454-459; doi: 10.1785/gssrl.78.4.454
- de Quervain. A., and Piccard, A. (1924). Beschreibung des 21-Tonnen-Universal-Seismographen System de Quervain-Piccard. *Jahresbericht Schweiz. Erdbebendienst*.
- de Quervain. A., and Piccard, A. (1927). Description du séismographe universel de 21 tonnes système de Quervain-Piccard. *Publ. Bureau Centr. Séism. Int. Sér. A*, **4**(32).

- Dewey, J., and Byerly, P. (1969). The early history of seismometry (to 1900). *Bull. Seism. Soc. Am.*, **59**(1), 183-227.
- Ewing, J. A. (1884). Measuring earthquakes. *Nature*, **30**, 149-152, 174-177.
- Forbriger, Th. (2007). Reducing magnetic field induced noise in broad-band seismic recordings. *Geophys. J. Int.*, **169**, 240-258.
- Forbriger, Th. (2009). About the non-unique sensitivity of pendulum seismometers to translational, angular, and centripetal acceleration. In Lee et al. (eds) (2009), 1343-1351.
- Forbriger, Th., Widmer-Schmidrig, R., Wielandt, E., Hayman, M., and Ackerley, N. (2010). Magnetic field background variations can limit the sensitivity of seismic broadband sensors. *Geoph. J. Int.*, **183**, 303-312.
- Galitzin, B. B. (1914). Vorlesungen über Seismometrie. *Verlag Teubner*, Berlin 1914, VIII + 538 pp.
- Gal'perin, E. I. (1955). Azimutal'nij metod sejsmicheskikh nabludenij. Gostoptechizdat, 80 pp.
- Gal'perin, E. I. (1977). Polyarizationnyi metod seismicheskikh isledovanii. *Nedra Press*, Moscow.
- Harjes, H.-P., and Seidl, D. (1978). Digital recording and analysis of broad-band seismic data at the Gräfenberg (GRF-) array. *J. Geophys.*, **44**, 511-523.
- Havskov, J., and Alguacil, G. (2004). *Instrumentation in earthquake seismology*. Modern Advances in Geophysics, vol. 22, ISBN 1402029683, Springer Publishers, Berlin; 2<sup>nd</sup> edition (2006), 358 pp. plus CD.
- Holcomb, G. L. (1989). A direct method for calculating instrument noise levels in side-by-side seismometer evaluations. *U. S. Geological Survey Open-file report* **89-214**.
- Holcomb, G. L. (1990). A numerical study of some potential sources of error in side-by-side seismometer evaluations. *U. S. Geological Survey, Open-file report* **90-406**.
- Holcomb, G. L., and Hutt, C. R. (1992). An evaluation of installation methods for STS-1 seismometers. *U. S. Geological Survey, Albuquerque, NM, Open File Report* 92-302.
- Hutt, C. R., Evans, J. R., Followill, F., Nigbor, R. L., and Wielandt, E. (2009). Guidelines for standardized testing of broadband seismometers and accelerometer. *USGS Geological Survey, Albuquerque, New Mexico, Open File Report* 2009-1295, 62p; online publication: <http://pubs.usgs.gov/of/2009/1295/>
- IRIS (1985). The design goals for a new global seismographic network. *Incorporated Research Institutions for Seismology*, Washington, DC.
- Jones, R. V., and Richards, J. C. S. (1973). The design and some applications of sensitive capacitance micrometers. *J. Physics E: Scientific Instruments*, **6**, 589-600.
- Knothe, Ch. (1963). Verbesserte Auswertung tiefeiseismischer Beobachtungen durch Verwendung von Mehrkomponenten-Stationen. *Freiberger Forschungshefte*, D149, 53 pp.
- LaCoste, L. J. B. (1934). A new type long period seismograph. *Physics*, **5**, 178-180.
- Lehner, F. E. (1959). An ultra long-period seismograph galvanometer. *Bull. Seism. Soc. Am.*, **49**(4), 399-401.
- Melton, B. S. (1981a). Earthquake seismograph development: a history – part 1. *EOS*, **62**(21), 505-510.

- Melton, B. S. (1981b). ). Earthquake seismograph development: a history – part 2. *EOS*, **62**(25), 545-548.
- Melton, B. S., and Kirkpatrick, B. M. (1970). The symmetric triaxial seismometer - its design for application in long-period seismometry. *Bull. Seism. Soc. Am.*, **60**, 3, 717-739.
- Miller, W. F. (1963). The Caltech digital seismograph. *J. Geophys. Res.*, **68**(3), 841-847.
- Mitronovas, W., and Wielandt, E. (1975). High-precision phase calibration of long-period electromagnetic seismographs. *Bull. Seism. Soc. Am.*, **65**, 2, 441-424.
- Oliver, J., and Murphy, L. (1971). WWSSN: Seismology's global network of observing stations. *Science*, **174**, 254-261.
- Oppenheim, A. V., and Schafer, R. V. (1975) Digital signal processing. Prentice Hall, New Jersey, USA.
- Oppenheim, A. V., and Willsky, A. S. (1983). Signals and Systems. *Prentice Hall*, New Jersey, USA
- Oppenheim, A. V., Schafer, R. V., and Buck, J. R. (1999; 2009). Discrete-time signal processing. 2<sup>nd</sup> and 3<sup>rd</sup> edition, *Prentice Hall*, New Jersey, USA.
- Passcal Instrument Center (Anonymous) (2009 a). Broadband Vault Construction (Manual), <http://www.passcal.nmt.edu/content/broadband-vault-construction-manual>
- Passcal Instrument Center (Anonymous) (2009b). Seismic Vaults, <http://www.passcal.nmt.edu/content/instrumentation/field-procedures/seismic-vaults>
- Pavlis, G. L., and Vernon, F. L. (1994). Calibration of seismometers using ground noise. *Bull. Seism. Soc. Am.*, **84**, 4, 1243-1255.
- Peterson, J. (1993). Observations and modelling of seismic background noise. *U.S. Geol. Survey Open-File Report 93-322*, 95 pp.
- Peterson, J., Butler, H. M., Holcomb, L. G., and Hutt, C. R. (1976). The seismic research observatory. *Bull. Seism. Soc. Am.*, **66**(3), 2049-2068.
- Peterson, J., Hutt, C. R., and Holcomb, L. G. (1980). Test and calibration of the seismic research observatory. *U.S. Geol. Survey, Albuquerque, Open-File Report 80-187*.
- Plešinger, A., and Horalek, J. (1976). The seismic broad-band recording and data processing system FBS/DPS and its seismological applications. *J. Geophys.*, **42**, 201-217.
- Plešinger, A., Zmeškal, M., and Zednik, J. (1996). Automated preprocessing of digital seismograms: Principles and software. Version 2.2, Bergman, E. (Ed.), *Prague & Golden*.
- Press, F., Ewing, M., and Lehner, F. (1958). A long-period seismograph system. *Trans. Am. Geophys. Union*, **39**(1), 106-108.
- Riedesel, M. A., Moore, R. D., and Orcutt, J. A. (1990). Limits of sensitivity of inertial seismometers with velocity transducers and electronic amplifiers. *Bull. Seism. Soc. Am.*, **80**, 6, 1725 - 1752.
- Robinson, E. A., and Treitel, S. (1980). Geophysical signal analysis. *Prentice-Hall*, New York.
- Rodgers, P. W. (1968). The response of the horizontal pendulum seismometer to Rayleigh and Love waves, tilt, and free oscillations of the Earth. *Bull. Seism. Soc. Am.*, **58**, 1384-1406.

- Rodgers, P. W. (1969). A note on the response of the pendulum seismometer to plane wave rotation. *Bull. Seism. Soc. Am.*, **59**, 2101-2102.
- Rodgers, P. W. (1992). Frequency limits for seismometers as determined from signal-to-noise ratios. Part 1: The electromagnetic seismometer. *Bull. Seism. Soc. Am.*, **82**, 1071-1098.
- Rodgers, P. W. (1993). Maximizing the signal-to-noise ratio of the electromagnetic seismometer: The optimum coil resistance, amplifier characteristics, and circuit. *Bull. Seism. Soc. Am.*, **83**, 2, 561-582.
- Rodgers, P. W. (1994). Self-noise spectra for 34 common electromagnetic seismometer/pre-amplifier pairs. *Bull. Seism. Soc. Am.*, **84**, 1, 222-229.
- Rodgers, P. W., Martin, A. J., Robertson, M. C., Hsu, M. M., and Harris, D. B. (1995). Signal- coil calibration of electromagnetic seismometers. *Bull. Seism. Soc. Am.*, **85**, 3, 845-850.
- Savino, J. M., Murphy, A. J., Rynn, J. M. W., Tatham, R., Sykes, L. R., Choy, G. L., and McCamy, K. (1972). Results from the high-gain long-period seismograph experiment. *Geophys. J. R. astron. Soc.*, **31**(1). 179-203.
- Scherbaum, F. (1996). *Of poles and zeros; Fundamentals of Digital Seismometry*, Kluwer Academic Publishers, Boston, 256 pp.
- Scherbaum, F. (2007). *Of poles and zeros; Fundamentals of Digital Seismometry*. 3<sup>rd</sup> revised edition, *Springer Verlag*, Berlin und Heidelberg.
- Schuessler, H. W. (1981). A signal processing approach to simulation. *Frequenz*, **35**, 174-184.
- Sleeman, R., van Wettum, A., and Trampert, J. (2006). Three-channel correlation analysis: a new technique to measure instrumental noise of digitizers and seismic sensors. *Bull. Seism. Soc. Am.*, **96**(1), 258-271, doi:10.1785/0120050032.
- Tanimoto, T. (1999). Excitation of normal modes by atmospheric turbulence: source of long-period seismic noise. *Geophys. J. Int.*, **136**(2), 395-402.
- Trnkoczy, A. (1998). Guidelines for civil engineering works at remote seismic stations. Application Note 42, Kinematics Inc., 222 Vista Av., Pasadena, Ca. 91107; <http://www.pdfio.com/k-341167.html>
- Uhrhammer, R. A., and Karavas, W. (1997). Guidelines for installing broadband seismic instrumentation. *Technical Report*, Seismic station, University of California at Berkeley, <http://seismo.berkeley.edu/bdsn/instrumentation/guidelines.html>
- Uhrhammer, R. A., Karavas, W., and Romanovicz, B. (1998). Broadband Seismic Station Installation Guidelines, *Seism. Res. Lett.*, **69**, 15-26.
- Usher, M. J., Guralp, C., and Burch, R. F. (1978). The design of miniature wideband seismometers. *Geophys. J.*, **55**(3), 605-613.
- von Rebeur-Paschwitz, E. (1889). The earthquake of Tokyo, April 18, 1889. *Nature*, **40**, 294-295.
- Wielandt, E. (1975). Ein astasiertes Vertikalpendel mit tragender Blattfeder. *J. Geophys.*, **41**, 5, 545-547.
- Wielandt, E. (1983). Design principles of electronic inertial seismometers. In: *Earthquakes: Observations, theory and interpretation*, LXXXV Corso, Soc. Italiana di Fisica, Bologna, 354-365.



- Wielandt, E. (2002a). Seismic sensors and their calibration. In: P. Bormann (ed.) (2002). *IASPEI New Manual of Seismological Observatory Practice*, GeoForschungsZentrum Potsdam, Potsdam, Vol.1, Chapter 5, 46 pp. (electronic version 2009; DOI: 10.2312/GFZ.NMSOP\_r1\_ch5)
- Wielandt, E. (2002b). Seismometry. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*. Academic Press, Amsterdam, ISBN-10: 0-12-440652-1, 283-304.
- Wielandt, E., and Streckeisen, G. (1982). The leaf-spring seismometer: Design and performance. *Bull. Seism. Soc. Am.*, **72**, 2349-2367.
- Wielandt, E., and Steim, J. M. (1986). A digital very-broad-band seismograph. *Annales Geophysicae*, **4**, 227-232.
- Wieland, E., and Forbriger, T. (1999). Near-field seismic displacement and tilt associated with the explosive activity of Stromboli. *Annali die Geofisica*, **42**, 407-416.
- Willmore, P. L. (1959). The application of the Maxwell impedance bridge to the calibration of electromagnetic seismographs. *Bull. Seism. Soc. Am.*, **49**, 99-114.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.
- Zürn, W., and Widmer, R. (1995). On noise reduction in vertical seismic records below 2 mHz using local barometric pressure. *Geophys. Res. Lett.*, **22**(24), 3537-3540.



| Topic       | Common seismic sensors                                                                                                                                                     |
|-------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Compiled by | Erhard Wielandt (formerly Institute of Geophysics, University of Stuttgart, D-70184 Stuttgart); E-mail: <a href="mailto:e.wielandt@t-online.de">e.wielandt@t-online.de</a> |
| Version     | April 2011; DOI: 10.2312/GFZ.NMSOP-2_DS_5.1                                                                                                                                |

## Introduction

These data sheets describe some widely used broadband seismic sensors, and a few other sensors that are new or have an interesting principle of operation. We present them as examples of a format in which seismic sensors might be uniformly described; our choice does not imply any recommendation for or against a specific instrument.

Manufacturers' specifications, especially for complex characteristics such as sensitivity and dynamic range, are sometimes realistic, sometimes optimistic, and sometimes useless. Some widely used broadband seismometers have recently been tested for noise at the USGS Albuquerque Seismological Lab. We have used these data wherever possible (see Ringler and Hutt, 2010). But independent information is not always available, and then the information from the manufacturer's data sheet must somehow be translated into our scheme. We have done this to the best of our knowledge, but errors and even a subjective bias cannot be ruled out.

Noise specifications are among the most important for the seismologist and among the most difficult for the manufacturer. Seismometers are typically produced in a noisy industrial environment, so most manufacturers cannot easily test their products for self-noise. New instruments often show transient disturbances which may disappear within a few months in a permanent installation but interfere with noise tests immediately after production. In order to ascertain the noise specifications, lengthy tests of each instrument at a remote quiet site would be required, and a substantial portion of the production might have to go to scrap. Customers are not willing to pay for this, and consequently manufacturers do not guarantee the noise specifications. Nevertheless, depending on details of the production process and on the time allocated to testing, some manufacturers turn out a higher proportion of faultless instruments than others. The user should be aware of this problem, but we cannot extrapolate it from the past into the future, and cannot quantify it in our data sheets. Also, part of what appears to be self-noise is often of environmental origin, and depends on undocumented details of the site and the test setup.

The "category" information in the data sheets may require some explanation. The term "broadband" (BB) is commonly used for seismometers that have a flat response to velocity from short periods to at least 30 seconds. The nominal bandwidth is however not what really matters. Instruments should be named after the frequency band in which they deliver useful signals. We will therefore use the term "broadband" for instruments that can be used throughout the classical short-period and long-period bands, and "very broadband" or VBB for instruments suitable to record free oscillations of the Earth. "Symmetric triaxial" means that the sensitive axes of the mechanical sensors are 54.7 degrees from vertical (Galperin 1955 and 1977), such as in the STS2 and the Trilliums.

Below, the sensor data sheets are alphabetically ordered. Another overview over broadband seismometers is found at [http://sismic.iec.cat/orfeus/instrumentation/vault\\_sensors.html](http://sismic.iec.cat/orfeus/instrumentation/vault_sensors.html).

|                      |                                                                                                                                                     |
|----------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>BBVS-60</b>                                                                                                                                      |
| Manufacturer         | <b>Beijing GangZhen Mechanical 6 Electronic Technology Co. Ltd</b>                                                                                  |
| Address              | No. 20 Jinxing Rd, Daxing district, Beijing 102628, China                                                                                           |
| Phone (Fax)          | ++86-10-60212319 (-37)                                                                                                                              |
| E-mail               | <a href="mailto:geodevice@sina.com">geodevice@sina.com</a>                                                                                          |
| Homepage             | <a href="http://www.geodevice.cn/">http://www.geodevice.cn/</a>                                                                                     |
|                      |                                                                                                                                                     |
| Category             | <b>Force-balance broadband, three component</b>                                                                                                     |
| Flat response        | Velocity 16.7 mHz (60 s) to 50 Hz                                                                                                                   |
| Resolution: NLNM     | 16.7 mHz to 8 Hz                                                                                                                                    |
| within 10 dB of NLNM | 8 mHz to 20 Hz                                                                                                                                      |
| within 20 dB of NLNM | 5 mHz to 50 Hz                                                                                                                                      |
|                      |                                                                                                                                                     |
| Operating range      | $\pm 10$ mm/s                                                                                                                                       |
| Generator constant   | 2000 Vs/m                                                                                                                                           |
| Output               | differential, $\pm 10$ V per line                                                                                                                   |
|                      |                                                                                                                                                     |
| Weight               | 12 kg                                                                                                                                               |
| Size                 | 24 cm dia., 27 cm high                                                                                                                              |
| Power                | 9 to 18 volts, 1.1 watts                                                                                                                            |
| Manual control       | Mass lock                                                                                                                                           |
| Remote control       | centering                                                                                                                                           |
| Remote diagnostic    | Mass position                                                                                                                                       |
| Accessories          | Tools, thermal insulation                                                                                                                           |
| Typical installation | All purpose                                                                                                                                         |
| Data Sheet           | <a href="http://www.geodevice.cn/en/productInfo.aspx?n=20090921143217421562">http://www.geodevice.cn/en/productInfo.aspx?n=20090921143217421562</a> |
| Remarks              | This instrument uses an astatic leaf-spring suspension like the Streckeisen seismometers.                                                           |
|                      |                                                                                                                                                     |

|                      |                                                                                         |
|----------------------|-----------------------------------------------------------------------------------------|
| Type                 | <b>CMG-3ESP</b>                                                                         |
| Manufacturer         | <b>Guralp Systems Limited</b>                                                           |
| Address              | 3 Midas House, Calleva Park, Aldermaston,<br>Reading RG7 8EA, UK                        |
| Phone                | ++44 118 9819056                                                                        |
| E-mail               | <a href="mailto:sales@guralp.com">sales@guralp.com</a>                                  |
| Homepage             | <a href="http://www.guralp.com">www.guralp.com</a>                                      |
|                      |                                                                                         |
| Category             | <b>Force-balance BB, three-component</b>                                                |
| Flat response        | velocity 33 mHz (30 s) to 50 Hz, others available                                       |
| Resolution: NLNM     | 25 mHz to 16 Hz (manufacturer)                                                          |
| within 10 dB of NLNM |                                                                                         |
| within 20 dB of NLNM |                                                                                         |
|                      |                                                                                         |
| Operating range      | $\pm 10$ mm/s                                                                           |
| Generator constant   | 2000 Vs/m, others available                                                             |
| Output               | differential, $\pm 10$ V per line                                                       |
|                      |                                                                                         |
| Weight               | 9 kg                                                                                    |
| Size                 | 17 cm dia., 32 cm high                                                                  |
| Power                | 10 to 30 volts, 0.6 watts                                                               |
| Manual control       | mass lock                                                                               |
| Remote control       | centering, mass lock optional                                                           |
| Remote diagnostic    | mass position                                                                           |
| Accessories          | control and breakout box                                                                |
| Typical installation | all purpose. Submersible version is available                                           |
| Data sheet           | <a href="http://www.guralp.com/products/3ESP/">http://www.guralp.com/products/3ESP/</a> |
| Remarks              |                                                                                         |

|                      |                                                                                           |
|----------------------|-------------------------------------------------------------------------------------------|
| Type                 | <b>CMG-3ESP Compact</b>                                                                   |
| Manufacturer         | <b>Guralp Systems Limited</b>                                                             |
| Address              | 3 Midas House, Calleva Park, Aldermaston,<br>Reading RG7 8EA, UK                          |
| Phone                | ++44 118 9819056                                                                          |
| E-mail               | <a href="mailto:sales@guralp.com">sales@guralp.com</a>                                    |
| Homepage             | <a href="http://www.guralp.com">www.guralp.com</a>                                        |
|                      |                                                                                           |
| Category             | <b>Force-balance BB, three-component</b>                                                  |
| Flat response        | velocity 33 mHz (30 s) to 50 Hz (others available)                                        |
| Resolution: NLNM     | 40 mHz to 16 Hz (manufacturer)                                                            |
| within 10 dB of NLNM |                                                                                           |
| within 20 dB of NLNM |                                                                                           |
|                      |                                                                                           |
| Operating range      | $\pm 10$ mm/s                                                                             |
| Generator constant   | 2000 Vs /m (others available)                                                             |
| Output               | differential, $\pm 10$ V per line                                                         |
|                      |                                                                                           |
| Weight               | 8 kg                                                                                      |
| Size                 | 17 cm dia., 19 cm high                                                                    |
| Power                | 10 to 36 volts, 0.6 watts (low-power version available)                                   |
| Manual control       | -                                                                                         |
| Remote control       | mass lock, centering, calibration                                                         |
| Remote diagnostic    | mass position                                                                             |
| Accessories          | -                                                                                         |
| Typical installation | all purpose.                                                                              |
| Data sheet           | <a href="http://www.guralp.com/products/3ESPC/">http://www.guralp.com/products/3ESPC/</a> |
| Remarks              | Guralp offers this seismometer with a wide variety of responses.                          |

|                      |                                                                                                                                                                                                                  |
|----------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>CMG-3T, CMG3-TB (borehole)</b>                                                                                                                                                                                |
| Manufacturer         | <b>Guralp Systems Limited</b>                                                                                                                                                                                    |
| Address              | 3 Midas House, Calleva Park, Aldermaston,<br>Reading RG7 8EA, UK                                                                                                                                                 |
| Phone                | ++44 118 9819056                                                                                                                                                                                                 |
| E-mail               | <a href="mailto:sales@guralp.com">sales@guralp.com</a>                                                                                                                                                           |
| Homepage             | <a href="http://www.guralp.com">www.guralp.com</a>                                                                                                                                                               |
| Category             | <b>Force-balance VBB, three-component</b>                                                                                                                                                                        |
| Flat response        | velocity 8.33 mHz (120 s) to 50 Hz (others available)                                                                                                                                                            |
| Resolution: NLNM     | 30 mHz to 18 Hz*                                                                                                                                                                                                 |
| within 10 dB of NLNM | 5 mHz to 30 Hz*                                                                                                                                                                                                  |
| within 20 dB of NLNM | 3 mHz to 50 Hz*                                                                                                                                                                                                  |
| Operating range      | $\pm 13$ mm/s                                                                                                                                                                                                    |
| Generator constant   | 1500 Vs/m (others available)                                                                                                                                                                                     |
| Output               | differential, $\pm 10$ V per line                                                                                                                                                                                |
| Weight               | 14 kg                                                                                                                                                                                                            |
| Size                 | 17 cm dia., 34 cm high                                                                                                                                                                                           |
| Power                | 10 to 30 volts, 0.75 watts                                                                                                                                                                                       |
| Manual control       | -                                                                                                                                                                                                                |
| Remote control       | mass lock, centering, calibration                                                                                                                                                                                |
| Remote diagnostic    | mass position                                                                                                                                                                                                    |
| Accessories          | control and breakout box                                                                                                                                                                                         |
| Typical installation | observatory, vault, and field use                                                                                                                                                                                |
| Data sheet           | <a href="http://www.guralp.com/products/3T/">http://www.guralp.com/products/3T/</a>                                                                                                                              |
| Remarks              | Tight thermal shielding recommended. Until recently the CMG-3T was one of the two most widely used broadband seismometers (the other one being the STS-2). The noise data are from the borehole version CMG-3TB. |

\* based on Ringler and Hutt (2010). Resolution above 10 Hz is extrapolated; data from the borehole version.

|                      |                                                           |
|----------------------|-----------------------------------------------------------|
| Type                 | <b>CTS-1</b>                                              |
| Manufacturer         | <b>Wuhan Lique Seismic Observation Technology Co. Ltd</b> |
| Address              | No. 40 Hongshan Celu, Wuhan, Hubei, China                 |
| Phone and Fax        | ++86-27-87864644                                          |
| E-mail               | <a href="mailto:actseis@gmail.com">actseis@gmail.com</a>  |
| Homepage             | <a href="http://www.whlqtech.com">www.whlqtech.com</a>    |
|                      |                                                           |
| Category             | <b>Force-Balance VBB, three component</b>                 |
| Flat response        | Velocity 8.3 mHz (120 s) to 50 Hz                         |
| Resolution: NLNM     | 10 mHz (100 s) to 10 Hz                                   |
| within 10 dB of NLNM | 2 mHz (500 s) to 20 Hz                                    |
| within 20 dB of NLNM |                                                           |
|                      |                                                           |
| Operating range      | $\pm 10$ mm/s (velocity)                                  |
| Generator constant   | 2000 Vs/m (velocity)                                      |
| Output               | differential, $\pm 10$ V per line                         |
|                      |                                                           |
| Weight               | 12 kg                                                     |
| Size                 | 294 mm dia., 27 cm high                                   |
| Power                | 9 – 18 V, 2 W                                             |
| Manual control       |                                                           |
| Remote control       | Mass lock, centering                                      |
| Remote diagnostic    | Mass position (acceleration output)                       |
| Accessories          | RCU box                                                   |
| Typical installation | Observatory, vault                                        |
| Data Sheet           |                                                           |
| Remarks              |                                                           |

|                      |                                                                                                                                                                                                                                 |
|----------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>Episensor ES-T</b>                                                                                                                                                                                                           |
| Manufacturer         | <b>Kinematics Inc.</b>                                                                                                                                                                                                          |
| Address              | 222 Vista Avenue, Pasadena, CA 91107, USA                                                                                                                                                                                       |
| Phone                | ++1 626 795 2220                                                                                                                                                                                                                |
| E-mail               | <a href="mailto:sales@kmi.com">sales@kmi.com</a>                                                                                                                                                                                |
| Homepage             | <a href="http://www.kinematics.com">www.kinematics.com</a>                                                                                                                                                                      |
|                      |                                                                                                                                                                                                                                 |
| Category             | <b>Force-balance broadband accelerometer</b>                                                                                                                                                                                    |
| Flat response        | acceleration DC to 200 Hz                                                                                                                                                                                                       |
| Resolution: NLNM     | not applicable (this is a strong-motion instrument)                                                                                                                                                                             |
| within 10 dB of NLNM | n.a.                                                                                                                                                                                                                            |
| within 20 dB of NLNM | n.a.                                                                                                                                                                                                                            |
|                      |                                                                                                                                                                                                                                 |
| Operating range      | user selectable, $\pm 0.25$ g to $\pm 4$ g                                                                                                                                                                                      |
| Generator constant   | user selectable, order of $1 \text{ V s}^2 / \text{m}$                                                                                                                                                                          |
| Output               | $\pm 2.5$ volt single-ended to $\pm 20$ volt differential                                                                                                                                                                       |
|                      |                                                                                                                                                                                                                                 |
| Weight               | 2 kg                                                                                                                                                                                                                            |
| Size                 | 13 cm dia., 6 cm high                                                                                                                                                                                                           |
| Power                | $\pm 12$ V or single 12 V, 0.15 to 0.4 W (depending on option)                                                                                                                                                                  |
| Manual control       | Centering; set gain                                                                                                                                                                                                             |
| Remote control       | -                                                                                                                                                                                                                               |
| Remote diagnostic    | -                                                                                                                                                                                                                               |
| Accessories          | -                                                                                                                                                                                                                               |
| Typical installation | Anywhere: Strong-motion monitoring, seismic zonation                                                                                                                                                                            |
| Data sheet           | <a href="http://www.kinematics.com/p-87-EpiSensor-ES-T.aspx">http://www.kinematics.com/p-87-EpiSensor-ES-T.aspx</a>                                                                                                             |
| Remarks              | This is a modern and popular strong-motion sensor. Its large dynamic range overlaps considerably with that of high-sensitivity seismometers, making it possible to record microearthquakes and teleseisms with this instrument. |

|                      |                                                                                                                                                                     |
|----------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>FSS-3M, FSS-3DBH</b>                                                                                                                                             |
| Manufacturer         | <b>Beijing GangZhen Mechanical &amp; Electronic Technology Co. Ltd</b>                                                                                              |
| Address              | No. 20 Jinxing Rd, Daxing district, Beijing 102628, China                                                                                                           |
| Phone (Fax)          | ++86-10-60212319 (-37)                                                                                                                                              |
| E-mail               | <a href="mailto:geodevice@sina.com">geodevice@sina.com</a>                                                                                                          |
| Homepage             | <a href="http://www.geodevice.cn/">http://www.geodevice.cn/</a>                                                                                                     |
|                      |                                                                                                                                                                     |
| Category             | <b>Force-balance short-period, three component Surface (M) and borehole (DBH) packages</b>                                                                          |
| Flat response        | Velocity 0.5Hz to 70 Hz                                                                                                                                             |
| Resolution: NLNM     | 0.2 Hz to 6 Hz                                                                                                                                                      |
| within 10 dB of NLNM | 0.2 Hz to 10 Hz                                                                                                                                                     |
| within 20 dB of NLNM |                                                                                                                                                                     |
|                      |                                                                                                                                                                     |
| Operating range      | ±10 mm/s                                                                                                                                                            |
| Generator constant   | 2000 Vs/m                                                                                                                                                           |
| Output               | differential, ±10 V per line                                                                                                                                        |
|                      |                                                                                                                                                                     |
| Weight               | 12 kg (FSS-3M), 8 kg (FSS-3BH)                                                                                                                                      |
| Size                 | 240 mm dia., 270 mm high (FSS-3M)<br>10 cm dia., 80 cm high (FSS-3BH)                                                                                               |
| Power                | 9 to 18 volts, 0.6 watts                                                                                                                                            |
| Manual control       | No centering necessary. No mass lock?                                                                                                                               |
| Remote control       |                                                                                                                                                                     |
| Remote diagnostic    |                                                                                                                                                                     |
| Accessories          |                                                                                                                                                                     |
| Typical installation | All purpose. FSS-3BH is a borehole version.                                                                                                                         |
| Data Sheet           | <a href="http://www.geodevice.net/FSS-3DBH.asp">http://www.geodevice.net/FSS-3DBH.asp</a>                                                                           |
| Remarks              | Widely used in China. According to the data sheet, the displacement transducer is electromagnetic (must be an LVDT) – quite unusual for a short-period seismometer. |
|                      |                                                                                                                                                                     |



|                      |                                                                                                                                                                                                                                                                                                                                     |
|----------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>S-13, GS-13</b>                                                                                                                                                                                                                                                                                                                  |
| Manufacturer         | <b>Geotech Instruments LLC</b>                                                                                                                                                                                                                                                                                                      |
| Address              | 10755 Sanden Drive, Dallas, Texas 75238-1336, USA                                                                                                                                                                                                                                                                                   |
| Phone                | (001) 214 221 0000                                                                                                                                                                                                                                                                                                                  |
| E-mail               | <a href="mailto:info@geoinstr.com">info@geoinstr.com</a>                                                                                                                                                                                                                                                                            |
| Homepage             | <a href="http://www.geoinstr.com/">www.geoinstr.com/</a>                                                                                                                                                                                                                                                                            |
| Category             | <b>Electromagnetic single-component, short-period seismometer</b> , convertible between horizontal and vertical                                                                                                                                                                                                                     |
| Flat response        | velocity 1 Hz (adjustable) to >100 Hz                                                                                                                                                                                                                                                                                               |
| Resolution: NLNM     | 0.07 to 40 Hz with low-noise preamplifier (S-13)*<br>0.05 to 150 Hz with low-noise preamplifier (GS-13)*                                                                                                                                                                                                                            |
| within 10 dB of NLNM |                                                                                                                                                                                                                                                                                                                                     |
| within 20 dB of NLNM | above 0.04 Hz (GS-13)*                                                                                                                                                                                                                                                                                                              |
| Operating range      | large, limited by preamplifier                                                                                                                                                                                                                                                                                                      |
| Generator constant   | 630 Vs / m with 3600 Ohms coil (S-13)<br>2180 Vs / m with 3600 Ohms coil (GS-13)                                                                                                                                                                                                                                                    |
| Output               | differential, unlimited                                                                                                                                                                                                                                                                                                             |
| Weight               | 14 kg. Moving mass is 5 kg.                                                                                                                                                                                                                                                                                                         |
| Size                 | 17 cm dia., 38 cm high                                                                                                                                                                                                                                                                                                              |
| Power                | passive                                                                                                                                                                                                                                                                                                                             |
| Manual control       | mass lock, centering, conversion Z / H                                                                                                                                                                                                                                                                                              |
| Remote control       | calibration                                                                                                                                                                                                                                                                                                                         |
| Remote diagnostic    | -                                                                                                                                                                                                                                                                                                                                   |
| Accessories          | -                                                                                                                                                                                                                                                                                                                                   |
| Typical installation | observatory, short-period                                                                                                                                                                                                                                                                                                           |
| Data sheet           | <a href="http://www.geoinstr.com/ds-s13.pdf">http://www.geoinstr.com/ds-s13.pdf</a>                                                                                                                                                                                                                                                 |
| Remarks              | A widely used high-performance, short-period seismometer. For very quiet sites, model GS-13 is the best choice. It resolves high-frequency signals much better than any broadband-feedback seismometer. See publications by Riedesel et al. (1990) and by Rodgers (1993 and 1994) for information on preamplifier design and noise. |

\* based on circuit analysis with op-amp LT1007 as a preamplifier. Low-noise model may be inaccurate above 10 Hz.

|                                                                                                                  |                                                                                                     |
|------------------------------------------------------------------------------------------------------------------|-----------------------------------------------------------------------------------------------------|
| Type                                                                                                             | <b>OYO-Geospace GS-1</b>                                                                            |
| Manufacturer                                                                                                     | <b>OYO Geospace</b>                                                                                 |
| Address                                                                                                          | 7007 Pinemont, Houston, Texas 77040,USA                                                             |
| Phone                                                                                                            | ++1-713-986-4444 Fax: -8723                                                                         |
| E-mail                                                                                                           | <a href="mailto:sales@oyogeospace.com">sales@oyogeospace.com</a>                                    |
| Homepage                                                                                                         | <a href="http://www.oyogeospace.com/technologies">http://www.oyogeospace.com/technologies</a>       |
|                                                                                                                  |                                                                                                     |
| Category                                                                                                         | <b>Passive electromagnetic single-component short-period</b>                                        |
| Flat response                                                                                                    | Velocity 1 Hz to >100 Hz (2 Hz version is available)                                                |
| Resolution: NLNM                                                                                                 | 0.11 to 15 Hz with low-noise preamplifier*                                                          |
| within 10 dB of NLNM                                                                                             | 0.07 to 50 Hz with low-noise preamplifier*                                                          |
| within 20 dB of NLNM                                                                                             | 0.05 to 150 Hz with low-noise preamplifier*                                                         |
|                                                                                                                  |                                                                                                     |
| Operating range                                                                                                  | large, limited by preamplifier                                                                      |
| Generator constant                                                                                               | 275 Vs/m with 4550 Ohms coil                                                                        |
| output                                                                                                           | differential, unlimited                                                                             |
|                                                                                                                  |                                                                                                     |
| Weight                                                                                                           | 2 kg. Moving mass is 0.7 kg.                                                                        |
| Size                                                                                                             | 8 cm dia. , 16 cm high                                                                              |
| Power                                                                                                            | -                                                                                                   |
| Manual control                                                                                                   | No mass lock required. Short signal coil and tilt the instrument for transportation.                |
| Remote control                                                                                                   | -                                                                                                   |
| Remote diagnostic                                                                                                | -                                                                                                   |
| Accessories                                                                                                      | -                                                                                                   |
| Typical installation                                                                                             | temporary installations, field work                                                                 |
| Data sheet                                                                                                       | <a href="http://www.geospacelp.com/index.php?id=169">http://www.geospacelp.com/index.php?id=169</a> |
| Remarks                                                                                                          | A three-component package is available as "Seis-Monitor"                                            |
| * based on circuit analysis with op-amp LT1007 as a preamplifier. Low-noise model may be inaccurate above 10 Hz. |                                                                                                     |

|                      |                                                                        |
|----------------------|------------------------------------------------------------------------|
| Type                 | <b>JCZ-1</b>                                                           |
| Manufacturer         | <b>Wuhan Lique Seismic Observation Technology Co. Ltd</b>              |
| Address              | No. 40 Hongshan Celu, Wuhan, Hubei, China                              |
| Phone and Fax        | ++86-27-87864644                                                       |
| E-mail               | <a href="mailto:actseis@gmail.com">actseis@gmail.com</a>               |
| Homepage             | <a href="http://www.whlqtech.com">www.whlqtech.com</a>                 |
|                      |                                                                        |
| Category             | <b>Force-Balance VBB, three component</b>                              |
| Flat response        | Velocity 2.8 mHz (360 s) to 10 Hz                                      |
| Resolution: NLNM     | 5 mHz (200 s) to 10 Hz                                                 |
| within 10 dB of NLNM | 1 mHz (1000 s) to 20 Hz                                                |
| within 20 dB of NLNM |                                                                        |
|                      |                                                                        |
| Operating range      | $\pm 10$ mm/s (velocity), $\pm 1$ mm/s <sup>2</sup> (VLP acceleration) |
| Generator constant   | 2000 Vs/m (velocity), 20000 Vs <sup>2</sup> /m (VLP acceleration)      |
| Output               | differential, $\pm 10$ V per line                                      |
|                      |                                                                        |
| Weight               | 30 kg                                                                  |
| Size                 | 51 * 57 cm, 49 cm high                                                 |
| Power                | 9 – 18 V, 30 W                                                         |
| Manual control       |                                                                        |
| Remote control       | Mass lock, centering                                                   |
| Remote diagnostic    | Mass position (acceleration output)                                    |
| Accessories          | TCU and RCU boxes                                                      |
| Typical installation | Observatory                                                            |
| Data Sheet           | No English data sheet available at the time of writing                 |
| Remarks              | Inner temperature stabilized with a thermostat                         |
|                      |                                                                        |

|                                     |                                                                                                                                                               |
|-------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                                | <b>KS-1</b>                                                                                                                                                   |
| Manufacturer                        | <b>Geotech Instruments LLC</b>                                                                                                                                |
| Address                             | 10755 Sanden Drive, Dallas, Texas 75238-1336, USA                                                                                                             |
| Phone                               | (001) 214 221 0000                                                                                                                                            |
| E-mail                              | <a href="mailto:info@geoinstr.com">info@geoinstr.com</a>                                                                                                      |
| Homepage                            | <a href="http://www.geoinstr.com/">www.geoinstr.com/</a>                                                                                                      |
|                                     |                                                                                                                                                               |
| Category                            | <b>Force-balance VBB, three-component</b>                                                                                                                     |
| Flat response                       | velocity 3 mHz (330 s) to 5 Hz                                                                                                                                |
| Resolution: NLNM                    | 10 mHz to 5 Hz*                                                                                                                                               |
| within 10 dB of NLNM                | 4 mHz to 9 Hz*                                                                                                                                                |
| within 20 dB of NLNM                | 0.3 mHz to 12 Hz*                                                                                                                                             |
|                                     |                                                                                                                                                               |
| Operating range                     | $\pm 8$ mm/s                                                                                                                                                  |
| Generator constant                  | 2400 Vs/m                                                                                                                                                     |
| Output                              | differential, $\pm 10$ V per line                                                                                                                             |
|                                     |                                                                                                                                                               |
| Weight                              | 43 kg                                                                                                                                                         |
| Size                                | 27 cm dia., 34 cm high                                                                                                                                        |
| Power                               | 24 V, 2.4 W                                                                                                                                                   |
| Manual control                      | -                                                                                                                                                             |
| Remote control                      | mass lock, centering, calibration                                                                                                                             |
| Remote diagnostic                   | mass position                                                                                                                                                 |
| Accessories                         | -                                                                                                                                                             |
| Typical use                         | Observatory                                                                                                                                                   |
| Data sheet                          | <a href="http://www.geoinstr.com/ds-ks1.pdf">http://www.geoinstr.com/ds-ks1.pdf</a>                                                                           |
| Remarks                             | This is the same system as the KS54000 borehole seismometer but packaged in two separate units (mechanical sensors and electronics) for surface installation. |
| * based on Ringler and Hutt (2010). |                                                                                                                                                               |

|                      |                                                                                                                                                              |
|----------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>KS-2000</b>                                                                                                                                               |
| Manufacturer         | <b>Geotech Instruments LLC</b>                                                                                                                               |
| Address              | 10755 Sanden Drive, Dallas, Texas 75238-1336, USA                                                                                                            |
| Phone                | (001) 214 221 0000                                                                                                                                           |
| E-mail               | <a href="mailto:info@geoinstr.com">info@geoinstr.com</a>                                                                                                     |
| Homepage             | <a href="http://www.geoinstr.com/">www.geoinstr.com/</a>                                                                                                     |
| Category             | <b>Force-balance, three-component, BB seismometer</b>                                                                                                        |
| Flat response        | velocity 83 mHz to 50 Hz                                                                                                                                     |
| Resolution: NLNM     | 50 mHz to 15 Hz*                                                                                                                                             |
| within 10 dB of NLNM | 35 mHz to 50 Hz*                                                                                                                                             |
| within 20 dB of NLNM | 15 mHz to >100 Hz*                                                                                                                                           |
| Operating range      | $\pm 10$ mm/s                                                                                                                                                |
| Generator constant   | 2000 Vs/m                                                                                                                                                    |
| Output               | differential, $\pm 10$ V per line                                                                                                                            |
| Weight               | 11 kg                                                                                                                                                        |
| Size                 | 19 cm dia., 29 cm high                                                                                                                                       |
| Power                | 9 to 36 V, no current specified                                                                                                                              |
| Manual control       | mass lock, centering (standard)                                                                                                                              |
| Remote control       | mass lock, centering (M and borehole models), calibration                                                                                                    |
| Remote diagnostic    | mass position                                                                                                                                                |
| Accessories          |                                                                                                                                                              |
| Typical installation | Vault, field deployment                                                                                                                                      |
| Data sheet           | <a href="http://www.geoinstr.com/ds-ks2000m.pdf">http://www.geoinstr.com/ds-ks2000m.pdf</a>                                                                  |
| Remarks              | Based on the KS-54000 but smaller and about 20 dB noisier at low frequencies. Standard broadband response. Borehole and short-period versions are available. |

\* based on Ringler and Hutt (2010). Resolution above 10 Hz is extrapolated.

|                      |                                                                                                     |
|----------------------|-----------------------------------------------------------------------------------------------------|
| Type                 | <b>KS-54000</b>                                                                                     |
| Manufacturer         | <b>Geotech Instruments LLC</b>                                                                      |
| Address              | 10755 Sanden Drive, Dallas, Texas 75238-1336, USA                                                   |
| Phone                | (001) 214 221 0000                                                                                  |
| E-mail               | <a href="mailto:info@geoinstr.com">info@geoinstr.com</a>                                            |
| Homepage             | <a href="http://www.geoinstr.com/">www.geoinstr.com/</a>                                            |
|                      |                                                                                                     |
| Category             | <b>Force-balance, three-component, VBB borehole seismometer</b>                                     |
| Flat response        | velocity 0.003 Hz (330 s) to 5 Hz                                                                   |
| Resolution: NLNM     | 10 mHz to 5 Hz*                                                                                     |
| within 10 dB of NLNM | 4 mHz to 9 Hz*                                                                                      |
| within 20 dB of NLNM | 0.3 mHz to 12 Hz*                                                                                   |
|                      |                                                                                                     |
| Operating range      | $\pm 1$ mm/s                                                                                        |
| Generator constant   | 20000 Vs/m                                                                                          |
| Output               | differential, $\pm 10$ V per line                                                                   |
|                      |                                                                                                     |
| Weight               | 66 kg                                                                                               |
| Size                 | 14 cm dia., 249 cm high                                                                             |
| Power                | 24 V, 2.4 W                                                                                         |
| Manual control       | -                                                                                                   |
| Remote control       | mass lock, module leveling / centering, calibration                                                 |
| Remote diagnostic    | mass position                                                                                       |
| Accessories          | ancillary equipment for borehole deployment                                                         |
| Typical installation | borehole up to 300 m deep                                                                           |
| Data sheet           | <a href="http://www.geoinstr.com/ds-ks54000.pdf">http://www.geoinstr.com/ds-ks54000.pdf</a>         |
| Remarks              | The standard borehole seismometer of the global network.<br>Accepts up to 10 degrees borehole tilt. |

\* based on Ringler and Hutt (2010).

|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|----------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>Le-3d lite</b>                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| Manufacturer         | <b>Lennartz Electronic</b>                                                                                                                                                                                                                                                                                                                                                                                                                        |
| Address              | Bismarckstrasse 136, D-72072 Tuebingen, Germany                                                                                                                                                                                                                                                                                                                                                                                                   |
| Phone                | ++49 7071 93550                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| E-mail               | <a href="mailto:info@lennartz-electronic.de">info@lennartz-electronic.de</a>                                                                                                                                                                                                                                                                                                                                                                      |
| Homepage             | <a href="http://www.lennartz-electronic.de">www.lennartz-electronic.de</a>                                                                                                                                                                                                                                                                                                                                                                        |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| Category             | <b>Active short-period, three-component</b>                                                                                                                                                                                                                                                                                                                                                                                                       |
| Flat response        | velocity 1 to 80 Hz                                                                                                                                                                                                                                                                                                                                                                                                                               |
| Resolution: NLNM     | ---                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| within 10 dB of NLNM | 0.2 to 0.5 Hz (microseismic peak)                                                                                                                                                                                                                                                                                                                                                                                                                 |
| within 20 dB of NLNM | 0.15 to 20 Hz                                                                                                                                                                                                                                                                                                                                                                                                                                     |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| Operating range      | $\pm 13$ mm/s below 4.5 Hz, decreasing at higher freq.                                                                                                                                                                                                                                                                                                                                                                                            |
| Generator constant   | 400 Vs/m                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| Output               | $\pm 5$ V single-ended                                                                                                                                                                                                                                                                                                                                                                                                                            |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| Weight               | 1.8 kg                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| Size                 | 95 mm dia., 65 mm high                                                                                                                                                                                                                                                                                                                                                                                                                            |
| Power                | 12 V, 0.1 W                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| Manual control       | no mass lock required                                                                                                                                                                                                                                                                                                                                                                                                                             |
| Remote control       | calibration                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| Remote diagnostic    | -                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| Accessories          | -                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| Typical installation | monitoring networks, field deployment                                                                                                                                                                                                                                                                                                                                                                                                             |
| Data sheet           | <a href="http://www.lennartz-electronic.de/MamboV4.5.2/index.php?option=com_content&amp;task=view&amp;id=72&amp;Itemid=55">http://www.lennartz-electronic.de/MamboV4.5.2/index.php?option=com_content&amp;task=view&amp;id=72&amp;Itemid=55</a>                                                                                                                                                                                                   |
| Remarks              | A very small and rugged three-component, short-period, active seismometer. The sensor is a commercial 4.5 Hz geophone whose response is electronically extended by a negative damping resistance (a form of negative feedback). Not as sensitive as larger seismometers but good enough where ground noise is not extremely low. A single-component version Le-1d is also available. Higher performance is offered by the Le-3d/5s and Le-3d/20s. |

|                      |                                                                                                                                                                                                                                                                                                                            |
|----------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>Le-3d-5s</b>                                                                                                                                                                                                                                                                                                            |
| Manufacturer         | <b>Lennartz Electronic</b>                                                                                                                                                                                                                                                                                                 |
| Address              | Bismarckstrasse 136, D-72072 Tuebingen, Germany                                                                                                                                                                                                                                                                            |
| Phone                | ++49 7071 93550                                                                                                                                                                                                                                                                                                            |
| E-mail               | <a href="mailto:info@lennartz-electronic.de">info@lennartz-electronic.de</a>                                                                                                                                                                                                                                               |
| Homepage             | <a href="http://www.lennartz-electronic.de">www.lennartz-electronic.de</a>                                                                                                                                                                                                                                                 |
|                      |                                                                                                                                                                                                                                                                                                                            |
| Category             | <b>Active medium-period, three-component</b>                                                                                                                                                                                                                                                                               |
| Flat response        | velocity 0.2 to 40 Hz                                                                                                                                                                                                                                                                                                      |
| Resolution: NLNM     | 0.15 to 1 Hz (microseismic peak)                                                                                                                                                                                                                                                                                           |
| within 10 dB of NLNM | 0.13 to 15 Hz                                                                                                                                                                                                                                                                                                              |
| within 20 dB of NLNM | 0.1 to 40 Hz                                                                                                                                                                                                                                                                                                               |
|                      |                                                                                                                                                                                                                                                                                                                            |
| Operating range      | $\pm 13$ mm/s below 4.5 Hz, decreasing at higher freq.                                                                                                                                                                                                                                                                     |
| Generator constant   | 400 Vs /m                                                                                                                                                                                                                                                                                                                  |
| Output               | $\pm 5$ V single-ended                                                                                                                                                                                                                                                                                                     |
|                      |                                                                                                                                                                                                                                                                                                                            |
| Weight               | 6.5 kg                                                                                                                                                                                                                                                                                                                     |
| Size                 | 195 mm dia., 165 mm high                                                                                                                                                                                                                                                                                                   |
| Power                | 12 V, 0.12 W                                                                                                                                                                                                                                                                                                               |
| Manual control       | no mass lock required                                                                                                                                                                                                                                                                                                      |
| Remote control       | calibration                                                                                                                                                                                                                                                                                                                |
| Remote diagnostic    | -                                                                                                                                                                                                                                                                                                                          |
| Accessories          | -                                                                                                                                                                                                                                                                                                                          |
| Typical installation | monitoring networks, field deployment                                                                                                                                                                                                                                                                                      |
| Data sheet           | <a href="http://www.lennartz-electronic.de/MamboV4.5.2/index.php?option=com_content&amp;task=view&amp;id=72&amp;Itemid=55">http://www.lennartz-electronic.de/MamboV4.5.2/index.php?option=com_content&amp;task=view&amp;id=72&amp;Itemid=55</a>                                                                            |
| Remarks              | A small and rugged three-component, medium-period, active seismometer. The sensor is a commercial 2 Hz geophone whose response is electronically extended by a negative damping resistance (a form of negative feedback). Not as sensitive as larger seismometers but good enough where ground noise is not extremely low. |



|                      |                                                                                                                                                                                                                                                                                                                                                                                |
|----------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>Mark L4-3D (out of production)</b>                                                                                                                                                                                                                                                                                                                                          |
| Manufacturer         | <b>Mark Products (disappeared)</b>                                                                                                                                                                                                                                                                                                                                             |
| Address              |                                                                                                                                                                                                                                                                                                                                                                                |
| Phone                |                                                                                                                                                                                                                                                                                                                                                                                |
| E-mail               |                                                                                                                                                                                                                                                                                                                                                                                |
| Homepage             |                                                                                                                                                                                                                                                                                                                                                                                |
| Category             | <b>Passive electromagnetic three-component short-period</b>                                                                                                                                                                                                                                                                                                                    |
| Flat response        | velocity 1 Hz to >100 Hz (2 Hz version was available)                                                                                                                                                                                                                                                                                                                          |
| Resolution: NLNM     | 0.12 to 12 Hz with low-noise preamplifier*                                                                                                                                                                                                                                                                                                                                     |
| within 10 dB of NLNM | 0.07 to 35 Hz with low-noise preamplifier*                                                                                                                                                                                                                                                                                                                                     |
| within 20 dB of NLNM | 0.05 to 100 Hz with low-noise preamplifier*                                                                                                                                                                                                                                                                                                                                    |
| Operating range      | large, limited by preamplifier                                                                                                                                                                                                                                                                                                                                                 |
| Generator constant   | 275 Vs/m with 5000 Ohms coil                                                                                                                                                                                                                                                                                                                                                   |
| output               | differential, unlimited                                                                                                                                                                                                                                                                                                                                                        |
| Weight               | 12 kg. Moving mass is 1 kg per component.                                                                                                                                                                                                                                                                                                                                      |
| Size                 | 24 cm dia. , 24 cm high                                                                                                                                                                                                                                                                                                                                                        |
| Power                | -                                                                                                                                                                                                                                                                                                                                                                              |
| Manual control       | no mass lock required. Short signal coils and tilt the instrument for transportation.                                                                                                                                                                                                                                                                                          |
| Remote control       | -                                                                                                                                                                                                                                                                                                                                                                              |
| Remote diagnostic    | -                                                                                                                                                                                                                                                                                                                                                                              |
| Accessories          | -                                                                                                                                                                                                                                                                                                                                                                              |
| Typical installation | temporary installations, field work                                                                                                                                                                                                                                                                                                                                            |
| Data sheet           | <a href="http://touro.ligo-la.caltech.edu/~ospjeld/HEPI_Documentation_Website/HEPI_Main_Page/pdfs/L4.pdf">http://touro.ligo-la.caltech.edu/~ospjeld/HEPI_Documentation_Website/HEPI_Main_Page/pdfs/L4.pdf</a>                                                                                                                                                                  |
| Remarks              | A still widely used short-period seismometer for field work. See publications by Riedesel et al. (BSSA <b>80</b> ,6) and by Rodgers (BSSA <b>83</b> ,2 and <b>84</b> ,1) for information on preamplifier design and noise.<br>The Mark L4-3D is out of production but OYO-Geospace offers a comparable three-component package, the “Seis-Monitor” based on the GS-1 geophone. |

\* based on circuit analysis with op-amp LT1007 as a preamplifier. Low-noise model may be inaccurate above 10 Hz.

|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|----------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>PMD BB 603</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| Manufacturer         | <b>Precision Measurement Devices</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| Address              | 105F W. Dudleytown Rd., Bloomfield, CT 06002, USA                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| Phone                | ++1 860 242 8177                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| E-mail               | <a href="mailto:sales@pmdsci.com">sales@pmdsci.com</a>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| Homepage             | <a href="http://www.pmdsci.com/">http://www.pmdsci.com/</a>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| Category             | <b>Molecular-electronic BB three-component, with force feedback</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| Flat response        | velocity 16.7 mHz (60 s) to 50 Hz                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| Resolution: NLNM     | 0.05 Hz (20 s) to 5 Hz (manufacturer)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| within 10 dB of NLNM |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| within 20 dB of NLNM |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| Operating range      | $\pm 10$ mm/s at low frequencies                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| Generator constant   | 2000 Vs/m (other values optional)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| Output               | differential, $\pm 10$ V per line                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| Weight               | 11 kg                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| Size                 | 20 cm dia., 22 cm high                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| Power                | 9 – 30 volt, 35 mW; low-power version (15 mW) available                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| Manual control       | No mass lock or centering required                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| Remote control       | Calibration, gain setting over serial port (optional)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| Remote diagnostic    | -                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| Accessories          | -                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| Typical installation | battery-powered stations, harsh environments.<br>borehole and OBS versions available.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| Data sheet           | <a href="http://www.pmdsci.com/pdfs/BB603.pdf">http://www.pmdsci.com/pdfs/BB603.pdf</a>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| Remarks              | An unconventional sensor. The horizontals are essentially circular-tube tiltmeters where the flow of an electrolyte through a platinum mesh is measured using the MET effect (molecular-electronic transfer, also used in the “solion” transducer). The vertical sensor supports a column of electrolyte in a vertical tube with a membrane and a spring. These sensors have extremely low power consumption, are very rugged, need not be locked or leveled; they simply start working after power-on. Their performance in other respects is however inferior to force-balance sensors. |

|                      |                                                                                                                 |
|----------------------|-----------------------------------------------------------------------------------------------------------------|
| Type                 | <b>REFTEK 151-120 “Observer”</b>                                                                                |
| Manufacturer         | <b>Refraction Technology, Inc.</b>                                                                              |
| Address              | 1600, 10th Street, Suite A, Plano TX 75074, USA                                                                 |
| Phone                | ++1 214 440 1265                                                                                                |
| E-mail               | <a href="mailto:Info@reftek.com">Info@reftek.com</a>                                                            |
| Homepage             | <a href="http://www.reftek.com/">http://www.reftek.com/</a>                                                     |
|                      |                                                                                                                 |
| Category             | <b>Force-balance VBB, three components</b>                                                                      |
| Flat response        | Velocity 8.3 mHz (120 s) to 50 Hz                                                                               |
| Resolution: NLNM     | 8 mHz to 15 Hz*                                                                                                 |
| within 10 dB of NLNM | 3 mHz to 30 Hz*                                                                                                 |
| within 20 dB of NLNM | 0.4 mHz to 50 Hz*                                                                                               |
|                      |                                                                                                                 |
| Operating range      | ±10 mm/s                                                                                                        |
| Generator constant   | 2000 Vs/m                                                                                                       |
| Output               | differential, ± 10 V per line                                                                                   |
|                      |                                                                                                                 |
| Weight               | 12 kg                                                                                                           |
| Size                 | 24 cm dia., 27 cm high                                                                                          |
| Power                | 9 to 18 volts, 1.3 watts                                                                                        |
| Manual control       | mass lock                                                                                                       |
| Remote control       | centering                                                                                                       |
| Remote diagnostic    | mass position                                                                                                   |
| Accessories          |                                                                                                                 |
| Typical installation | observatory, vault                                                                                              |
| Data Sheet           | <a href="http://www.reftek.com/pdf/151-120.pdf">http://www.reftek.com/pdf/151-120.pdf</a>                       |
| Remarks              | A very good VBB sensor, in the free-mode band comparable to KS-1 and CMG-3TB according to Ringler & Hutt, 2010. |

\* based on Ringler and Hutt (2010). Resolution above 10 Hz is extrapolated.

|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|----------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>STS-1 VBB (out of production, see remarks below)</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| Manufacturer         | <b>Streckeisen AG</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| Address              | Dättlikoner Str. 5, CH-8422 Pfungen, Switzerland                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| Phone                | ++41 52 315 2161                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| E-mail               | none                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| Homepage             | none                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| Category             | <b>Force-balance VBB, separate Z and H components</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| Flat response        | velocity 2.67 mHz (360 sec) to 10 Hz                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| Resolution: NLNM     | originally <0.3 mHz to 3 Hz, later versions up to 8 Hz*                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| within 10 dB of NLNM | 0.1 mHz to 13 Hz*                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| within 20 dB of NLNM | <0.1 mHz to 20 Hz*                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| Operating range      | ± 8 mm/s                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| Generator constant   | 2400 Vs/m                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| Output               | differential, ±10 V per line                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| Weight               | 4 kg (vert.); 5.5 kg (hor.)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| Size                 | 12 * 17 * 18 cm (vert.); 20 cm dia. and 16 cm high (hor.)<br>Shields require a space of 50 * 50 cm, 60 cm high.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| Power                | ± 15 volts, 3.5 watts per component                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| Manual control       | mass lock (pins and screws inserted by hand)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| Remote control       | mass centering, calibration                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| Remote diagnostic    | mass position                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| Accessories          | feedback electronics are separate. Magnetic shield (for vert. comp. only), aluminum shield, glass base plate, vacuum glass bell included. Monitor (breakout) box optional.                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| Typical installation | observatory                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| Data sheet           | <a href="http://www.software-for-seismometry.de/textfiles/Seismometry/">www.software-for-seismometry.de/textfiles/Seismometry/</a>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| Remarks              | Operated under partial vacuum; no electronics inside the sensor. This seismometer made the very-broad-band velocity response popular, and is the only seismometer to resolve minimum ground noise throughout the long-period seismic band. (Some say it defines the low-noise model.)<br>The STS1 is no longer manufactured by Streckeisen AG but Metrozet LLC (USA) have developed a modern, remotely controllable replacement for the feedback electronics and a modified mechanical sensor. The new system has the excellent low-frequency resolution of the original STS-1 but is easier to install and maintain. |

\* based on Ringler and Hutt (2010). Resolution above 10 Hz is extrapolated.

|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |                                |
|----------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--------------------------------|
| Type                 | <b>STS-2</b> (out of production; replaced by STS 2.5)                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |                                |
| Manufacturer         | <b>Streckeisen AG</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |                                |
| Address              | Dättlikoner Str. 5, CH-8422 Pfungen, Switzerland                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |                                |
| Phone                | ++41 52 315 2161                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |                                |
| E-mail               | none                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |                                |
| Homepage             | none                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |                                |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |                                |
| Category             | <b>Force-balance VBB, symmetric-triaxial</b>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |                                |
| Flat response        | velocity 8.33 mHz (120 sec) to 50 Hz                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |                                |
| Resolution: NLNM     | 30 mHz to 8 Hz <sup>1</sup>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   | 3 mHz to 20 Hz <sup>3</sup>    |
| within 10 dB of NLNM | <1 mHz to 20 Hz <sup>1</sup>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  | 0.3 mHz to 35 Hz <sup>3</sup>  |
| within 20 dB of NLNM | <0.1 mHz to 50 Hz <sup>1</sup>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                | <0.1 mHz to 50 Hz <sup>3</sup> |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |                                |
| Operating range      | ±13 mm/s                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |                                |
| Generator constant   | 1500 Vs/m (high-gain version: 20 000 Vs/m)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |                                |
| Output               | differential, ±10 V per line                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |                                |
|                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |                                |
| Weight               | 13 kg                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |                                |
| Size                 | 23 cm dia., 26 cm high                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |                                |
| Power                | 9 to 28 volts, 1.5 watts, low-power version was available                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |                                |
| Manual control       | mass lock                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |                                |
| Remote control       | centering, short-period mode, calibration, select UVW positions or UVW broadband signals                                                                                                                                                                                                                                                                                                                                                                                                                                      |                                |
| Remote diagnostic    | mass positions or UVW signals (analog)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |                                |
| Accessories          | A “Host Box” with power conditioning and two parallel signal connectors included                                                                                                                                                                                                                                                                                                                                                                                                                                              |                                |
| Typical installation | observatory and vault. Needs additional protection in humid or dirty environments.                                                                                                                                                                                                                                                                                                                                                                                                                                            |                                |
| Data sheet           | <a href="http://www.passcal.nmt.edu/content/instrumentation/sensors/broadband-sensors/sts-2-bb-sensor">http://www.passcal.nmt.edu/content/instrumentation/sensors/broadband-sensors/sts-2-bb-sensor</a><br>Manual: <a href="http://www.passcal.nmt.edu/webfm_send/488">http://www.passcal.nmt.edu/webfm_send/488</a>                                                                                                                                                                                                          |                                |
| Remarks              | <p>Tightly-fitting thermal isolation required for optimum performance. Until recently, the STS-2 was one of the two most widely used VBB instruments (the other being the CMG3). Three versions (generations) of the internal electronics exist. Production ended in 2010.</p> <p>The successor model STS 2.5 has the same housing but completely redesigned mechanical sensors and electronics, and can be remotely controlled with a laptop computer. Its performance is similar to that of the third-generation STS-2.</p> |                                |

<sup>1</sup> first generation, <sup>3</sup> third-generation, high-gain STS-2, based on Ringler and Hutt 2010. Resolution above 10 Hz is extrapolated.

|                      |                                                                                                                                                                                                                                                                                                                                                         |
|----------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>STS 2.5</b>                                                                                                                                                                                                                                                                                                                                          |
| Manufacturer         | <b>Streckeisen GmbH; exclusively sold by Quanterra, USA</b>                                                                                                                                                                                                                                                                                             |
| Address              | St: Dättlikoner Str. 5, CH-8422 Pfungen, Switzerland<br>Q: 325 Ayer Rd, Harvard, Ma. 01451, USA                                                                                                                                                                                                                                                         |
| Phone                | ++41 52 315 6700 (Streckeisen)<br>++1 978 772 4774 (Quanterra)                                                                                                                                                                                                                                                                                          |
| E-mail               | <a href="mailto:bkemp@quanterra.com">bkemp@quanterra.com</a>                                                                                                                                                                                                                                                                                            |
| Homepage             | <a href="http://www.quanterra.com">www.quanterra.com</a> ;<br><a href="http://www.kinematics.com/p-163-Home.aspx">www.kinematics.com/p-163-Home.aspx</a>                                                                                                                                                                                                |
| Category             | <b>Force-balance VBB, symmetric-triaxial</b>                                                                                                                                                                                                                                                                                                            |
| Flat response        | velocity 8.33 mHz (120 sec) to 50 Hz                                                                                                                                                                                                                                                                                                                    |
| Resolution: NLNM     | 3 mHz to 20 Hz <sup>3</sup>                                                                                                                                                                                                                                                                                                                             |
| within 10 dB of NLNM | 0.3 mHz to 35 Hz <sup>3</sup>                                                                                                                                                                                                                                                                                                                           |
| within 20 dB of NLNM | 0.1 mHz to 50 Hz <sup>3</sup>                                                                                                                                                                                                                                                                                                                           |
| Operating range      | ±13 mm/s                                                                                                                                                                                                                                                                                                                                                |
| Generator constant   | 1500 Vs/m                                                                                                                                                                                                                                                                                                                                               |
| Output               | differential, ±10 volts per line                                                                                                                                                                                                                                                                                                                        |
| Weight               | 12 kg                                                                                                                                                                                                                                                                                                                                                   |
| Size                 | 23 cm dia., 26 cm high                                                                                                                                                                                                                                                                                                                                  |
| Power                | 9 to 28 volts, 0.5 watts                                                                                                                                                                                                                                                                                                                                |
| Manual control       | -                                                                                                                                                                                                                                                                                                                                                       |
| Remote control       | mass lock, centering, calibration, short-period mode                                                                                                                                                                                                                                                                                                    |
| Remote diagnostic    | mass position, state-of health parameters including temperature, internal humidity, tilt, serial number, supply voltage                                                                                                                                                                                                                                 |
| Accessories          | a "Host Box" with power conditioning and LED status indicators is included.                                                                                                                                                                                                                                                                             |
| Typical installation | Observatory, vault. Needs additional protection in humid or dirty environments.                                                                                                                                                                                                                                                                         |
| Data Sheet           | <a href="http://www.kinematics.com/uploads/PDFs/STS-2_5.pdf">http://www.kinematics.com/uploads/PDFs/STS-2_5.pdf</a>                                                                                                                                                                                                                                     |
| Remarks              | Tightly-fitting thermal isolation required for optimum performance. The STS 2.5 has the same housing as the STS-2 but completely redesigned mechanical sensors and electronics. Remote control with a laptop computer. The STS 2.5 has not yet been extensively tested but appears to perform equally or slightly better than a third-generation STS-2. |

<sup>3</sup> self-noise data are from a third-generation, high-gain STS2, based on Ringler and Hutt (2010). Resolution above 10 Hz is extrapolated.

|                      |                                                                                                                           |
|----------------------|---------------------------------------------------------------------------------------------------------------------------|
| Type                 | <b>Trillium 120 P / PA</b>                                                                                                |
| Manufacturer         | <b>Nanometrics Inc.</b>                                                                                                   |
| Address              | 250 Herzberg Road, Kanata, Ontario, Canada K2K 2A1                                                                        |
| Phone                | ++1 613 592 6776                                                                                                          |
| E-mail               | <a href="mailto:info@nanometrics.ca">info@nanometrics.ca</a>                                                              |
| Homepage             | <a href="http://www.nanometrics.ca/">http://www.nanometrics.ca/</a>                                                       |
| Category             | <b>Force-balance VBB, symmetric-triaxial</b>                                                                              |
| Flat response        | velocity 83 mHz (120 s) to 145 Hz                                                                                         |
| Resolution: NLNM     | 25 mHz to 12 Hz*                                                                                                          |
| within 10 dB of NLNM | 4 mHz to 20 Hz*                                                                                                           |
| within 20 dB of NLNM | <0.3 mHz to 50 Hz*                                                                                                        |
| Operating range      | $\pm 16$ mm/s                                                                                                             |
| Generator constant   | 1200 Vs/m                                                                                                                 |
| Output               | differential, $\pm 10$ V per line                                                                                         |
| Weight               | 7.5 kg                                                                                                                    |
| Size                 | 21 cm dia., 22 cm high                                                                                                    |
| Power                | 9 to 36 volt, 0.62 watt                                                                                                   |
| Manual control       | centering (120P). No mass lock required. Motorized mass centering (120PA).                                                |
| Remote control       | centering, UVW/XYZ mode, short/long period mode, firmware updates                                                         |
| Remote diagnostic    | mass positions, temperature, serial number, status                                                                        |
| Accessories          | -                                                                                                                         |
| Typical installation | all purpose                                                                                                               |
| Data sheet           | <a href="http://www.nanometrics.ca/products/trillium-120-ppa">http://www.nanometrics.ca/products/trillium-120-ppa</a>     |
| Remarks              | Large temperature range without re-centering ( $\pm 45^\circ$ ).<br>A posthole version with remote leveling is available. |

\* based on Ringler and Hutt (2010) and the manufacturer

|                                                         |                                                                                                                                                                                                                         |
|---------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Type                                                    | <b>Trillium 240</b>                                                                                                                                                                                                     |
| Manufacturer                                            | <b>Nanometrics Inc.</b>                                                                                                                                                                                                 |
| Address                                                 | 250 Herzberg Road, Kanata, Ontario, Canada K2K 2A1                                                                                                                                                                      |
| Phone                                                   | ++1 613 592 6776                                                                                                                                                                                                        |
| E-mail                                                  | <a href="mailto:info@nanometrics.ca">info@nanometrics.ca</a>                                                                                                                                                            |
| Homepage                                                | <a href="http://www.nanometrics.ca/">http://www.nanometrics.ca/</a>                                                                                                                                                     |
|                                                         |                                                                                                                                                                                                                         |
| Category                                                | <b>Force-balance VBB, symmetric-triaxial</b>                                                                                                                                                                            |
| Flat response                                           | velocity 41.7 mHz (240 s) to 35 Hz                                                                                                                                                                                      |
| Resolution: NLNM                                        | 10 mHz to 10 Hz*                                                                                                                                                                                                        |
| within 10 dB of NLNM                                    | 0.3 mHz to 20 Hz*                                                                                                                                                                                                       |
| within 20 dB of NLNM                                    | 0.1 mHz to 35 Hz*                                                                                                                                                                                                       |
|                                                         |                                                                                                                                                                                                                         |
| Operating range                                         | $\pm 16$ mm/s                                                                                                                                                                                                           |
| Generator constant                                      | 1200 Vs/m                                                                                                                                                                                                               |
| Output                                                  | differential, $\pm 10$ volts per line                                                                                                                                                                                   |
|                                                         |                                                                                                                                                                                                                         |
| Weight                                                  | 14 kg                                                                                                                                                                                                                   |
| Size                                                    | 25 cm dia., 29 cm high                                                                                                                                                                                                  |
| Power                                                   | 9 to 36 volt, 0.65 watt                                                                                                                                                                                                 |
| Manual control                                          | none: no mass lock required                                                                                                                                                                                             |
| Remote control                                          | integrated web browser with many control options                                                                                                                                                                        |
| Remote diagnostic                                       | Mass positions, temperature, serial number, status, calibration parameters                                                                                                                                              |
| Accessories                                             | prefabricated heat shield                                                                                                                                                                                               |
| Typical installation                                    | observatory, vault.                                                                                                                                                                                                     |
| Data sheet                                              | <a href="http://www.nanometrics.ca/products/trillium-240">http://www.nanometrics.ca/products/trillium-240</a>                                                                                                           |
| Remarks                                                 | An excellent very-broad-band seismometer, comparable to the STS2 in its performance. May be slightly better than the STS2 at the lowest frequencies when magnetically shielded. Has a modern digital control interface. |
| * based on Ringler and Hutt (2010) and the manufacturer |                                                                                                                                                                                                                         |



|                                                         |                                                                                                                       |
|---------------------------------------------------------|-----------------------------------------------------------------------------------------------------------------------|
| Type                                                    | <b>Trillium Compact</b>                                                                                               |
| Manufacturer                                            | <b>Nanometrics Inc.</b>                                                                                               |
| Address                                                 | 250 Herzberg Road, Kanata, Ontario, Canada K2K 2A1                                                                    |
| Phone                                                   | ++1 613 592 6776                                                                                                      |
| E-mail                                                  | <a href="mailto:info@nanometrics.ca">info@nanometrics.ca</a>                                                          |
| Homepage                                                | <a href="http://www.nanometrics.ca/">http://www.nanometrics.ca/</a>                                                   |
| Category                                                | <b>Miniature Force-balance BB, symmetric-triaxial</b>                                                                 |
| Flat response                                           | velocity 8.3 mHz (120 s) to 100 Hz                                                                                    |
| Resolution: NLNM                                        | 50 mHz to 8 Hz*                                                                                                       |
| within 10 dB of NLNM                                    | 45 mHz to 20 Hz*                                                                                                      |
| within 20 dB of NLNM                                    | 30 mHz to 50 Hz*                                                                                                      |
| Operating range                                         | ±26 mm/s                                                                                                              |
| Generator constant                                      | 750 Vs/m                                                                                                              |
| Output                                                  | differential, ±10 V per line                                                                                          |
| Weight                                                  | comparable to a 1 Hz geophone                                                                                         |
| Size                                                    | comparable to a 1 Hz geophone                                                                                         |
| Power                                                   | 9 to 36 volt, 0.16 watt                                                                                               |
| Manual control                                          | none: no mass lock or centering required                                                                              |
| Remote control                                          | RS-232 compatible integrated web server, UVW selection                                                                |
| Remote diagnostic                                       | mass positions, temperature, serial number, poles and zeros                                                           |
| Accessories                                             | -                                                                                                                     |
| Typical installation                                    | temporary installations, field work                                                                                   |
| Data sheet                                              | <a href="http://www.nanometrics.ca/products/trillium-compact">http://www.nanometrics.ca/products/trillium-compact</a> |
| Remarks                                                 | OBS and all-terrain versions with internal self-leveling gimbal are available.                                        |
| * based on Ringler and Hutt (2010) and the manufacturer |                                                                                                                       |

## Acknowledgment

The data for the Chinese seismometers of type BBVS-60, CTS-1, FSS-3M, FSS-3DHB, and JCZ-1 have kindly been provided by Prof. Liu Ruifeng from the China Earthquake Network Center (CENC).

## References

Gal'perin, E. I. (1955). Azimutal'nij metod sejsmiceskich nabludenij. Gostoptechizdat, 80 pp.

- Gal'perin, E. I. (1977). Polyarizationnyi metod seismicheskikh isledovaniy. *Nedra Press*, Moscow.
- Riedesel, M. A., Moore, R. D., and Orcutt, J. A. (1990). Limits of sensitivity of inertial seismometers with velocity transducers and electronic amplifiers. *Bull. Seism. Soc. Am.*, **80**, 6, 1725 - 1752.
- Ringler, A. T., and Hutt, C. R. (2010). Self-Noise Models of Seismic Instruments. *Seism. Res. Lett.*, **81**, 972-983.
- Rodgers, P. W. (1993). Maximizing the signal-to-noise ratio of the electromagnetic seismometer: The optimum coil resistance, amplifier characteristics, and circuit. *Bull. Seism. Soc. Am.*, **83**, 2, 561-582.
- Rodgers, P. W. (1994). Self-noise spectra for 34 common electromagnetic seismometer/pre-amplifier pairs. *Bull. Seism. Soc. Am.*, **84**, 1, 222-229.

| Title   | Plotting seismograph response<br>(BODE-diagram)                                                                                                                                                          |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Jens Bribach, GFZ German Research Centre for Geosciences, Dept. 2, Telegrafenberg, D-14473 Potsdam, Germany; Fax: +49 331 288 1266; E-mail: <a href="mailto:brib@gfz-potsdam.de">brib@gfz-potsdam.de</a> |
| Version | May 2001; DOI: 10.2312/GFZ.NMSOP-2_EX_5.1                                                                                                                                                                |

## 1 Aim

The exercise aims at making you familiar with the easy way of construction of a BODE-diagram which displays the transfer function of a given device as a plot of logarithmic amplitude  $A$  and of linear phase shift  $\phi$  versus logarithmic frequency  $f$  (or period  $1/f$ ). Its advantage is that response curves are approximated by straight lines (see IS 5.2). The main features are:

- any Pole in the transfer function generates an amplitude decay proportional to frequency  $f$  (20 dB per decade or 6 dB per octave) and a phase shift  $\phi$  of  $-90^\circ$ ;
- any Zero causes a slope of 1:1 too and a phase shift of  $+90^\circ$ ;
- corner frequencies (e.g., of filters) correspond to the point of intersection of two straight lines.

All stages of a signal-transfer chain can thus be constructed component-wise, one after the other. It is recommended to decompose all functions into parts of 1<sup>st</sup> or 2<sup>nd</sup> order. One gets the complete transfer function by multiplying these individual functions. In both the logarithmic amplitude scale and the linear phase scale this means adding the related individual curves.

## 2 Tasks

**Task 1:** Plot the BODE-diagrams (amplitude only) of the following seismograph components:

### Seismometer

|                     |                             |
|---------------------|-----------------------------|
| Transducer Constant | $G_S = 15.915 \text{ Vs/m}$ |
| Natural Period      | $T_S = 5 \text{ s}$         |
| Attenuation         | $D_S = 0.707$               |

### HIGH Pass HP1 (1<sup>st</sup> order)

|                  |                            |
|------------------|----------------------------|
| Magnification    | $A_{H1} = 3$               |
| Corner Frequency | $f_{H1} = 0.01 \text{ Hz}$ |

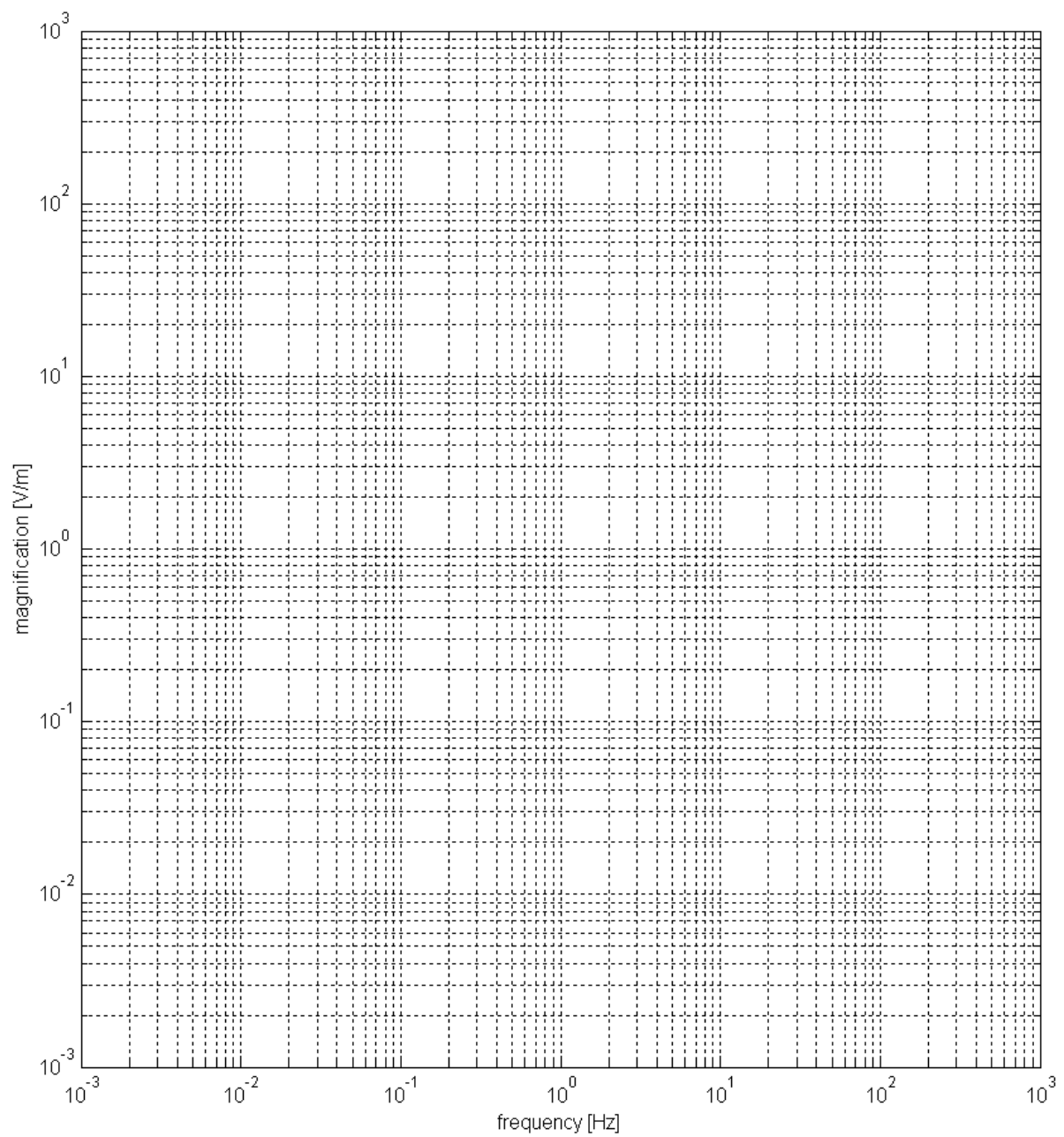
### LOW Pass LP1 (1<sup>st</sup> order)

|                  |                           |
|------------------|---------------------------|
| Magnification    | $A_{L1} = 5$              |
| Corner Frequency | $f_{L1} = 0.2 \text{ Hz}$ |

### LOW Pass LP2 (2<sup>nd</sup> order)

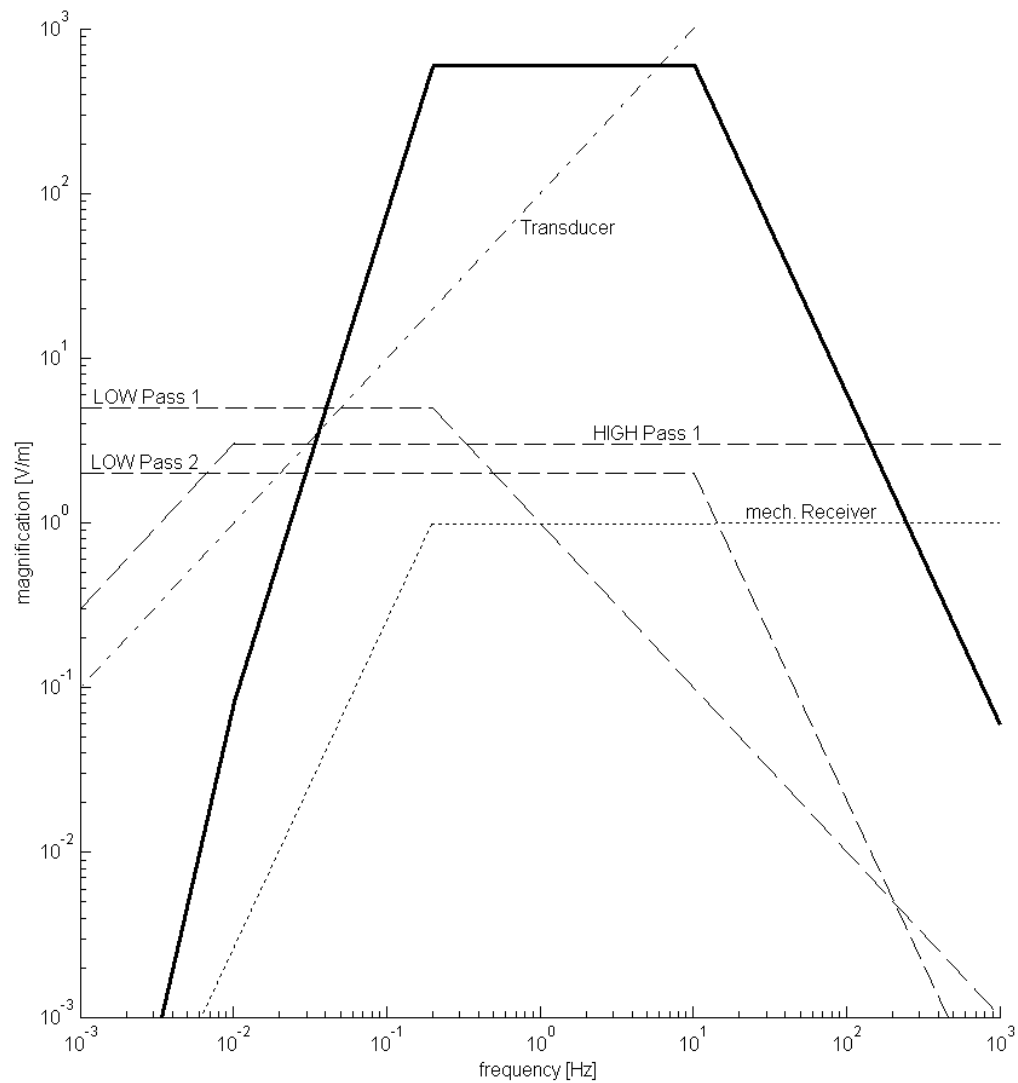
|                  |                          |
|------------------|--------------------------|
| Magnification    | $A_{L2} = 2$             |
| Corner Frequency | $f_{L2} = 10 \text{ Hz}$ |
| Attenuation      | $D_{L2} = 0.707$         |

**Task 2:** Plot the overall amplitude response of the system approximated by straight lines on double logarithmic paper (see Figure 1).

**Figure 1**

### 3 Solution

The solution to this exercise is given in Figure 2 below.



**Figure 2** Overall BODE-diagram (solid curve) for the seismograph amplitude response. It results from the logarithmic addition of the BODE-diagrams of all individual components given in Task 1.

| Title   | Estimating seismometer parameters by step function (STEP)                                                                                                                                                |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Jens Bribach, GFZ German Research Centre for Geosciences, Dept. 2, Telegrafenberg, D-14473 Potsdam, Germany; Fax: +49 331 288 1266; E-mail: <a href="mailto:brib@gfz-potsdam.de">brib@gfz-potsdam.de</a> |
| Version | May 2001; DOI: 10.2312/GFZ.NMSOP-2_EX_5.2                                                                                                                                                                |

## 1 Aim

To determine the response of a seismometer system in the time domain to a *STEP function* input. Applying a step impulse to a seismometer allows to derive the main seismometer parameters by analysing the generated time series. In the absence of expensive calibration equipment (e.g., shake table) or in the case of sealed seismometers this simple method is very suitable and can also be used under field conditions.

## 2 Procedures and relationships

### 2.1 Applying STEPs to the seismometer

Applying steps is the oldest calibration method in seismology. Teupser (1962) describes three main types:

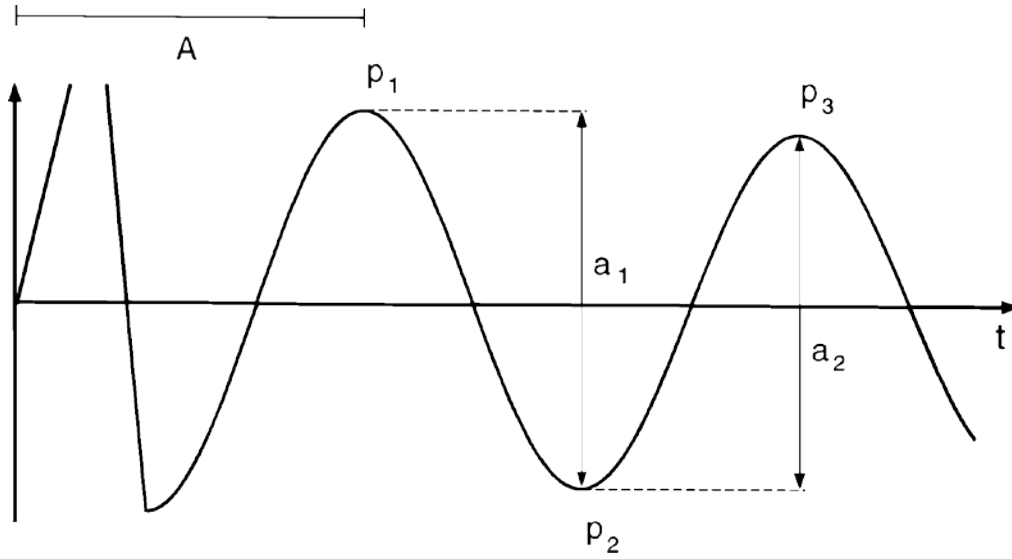
- a) pulling a thin block (thickness max. 0.01 mm) off the seismometer bottom;
- b) applying a heavy weight upon the seismometer platform;
- c) applying a constant current to the coil of an electrodynamical system (if available; for driving current see EX 5.3: Seismometer calibration by harmonic drive).

Because a) is the roughest method one should use it for field or for portable seismometers only and never for sensitive station sensors. In case a) and b) the seismometer mass will return to the former position after deflection, in case c) the seismometer mass will move to an offset position which will depend on the applied current. To ensure linearity the mass deflection - or the seismometer displacement - should not exceed several 100 micrometers.

### 2.2 Evaluating STEP-transition time series

#### 2.2.1 All types of seismometers ( $D_S < 0.5$ )

Figure 1 shows the time series of a low-damped seismometer ( $D_S = 0.1$ ). The time section A represents the time from the moment of step input up to the transition to a real harmonic movement of the mass. The moment of step causes odd signals. Mechanical application of a step impulse generates additional vibrations because of hitting effects. An electrical step can induce an electrical pulse if the calibration coil and the signal coil are mounted to the same core (the so-called transformer effect). Therefore the analysis of the generated time series should start only beyond section A with:



**Figure 1** Response of a low-damped seismometer to a step pulse.

**First step:** Measuring of the period and damping of the time series.

The period  $T$  should be measured by averaging over as many cycles as possible (10 or more) to get an accuracy better than 99%.

**Note!** The measured period is larger than the natural period because of the seismometer damping.

The damping  $D$  is calculated from the equation

$$D = \frac{1}{\sqrt{\left(\frac{(N-1)\pi}{\ln(a_1/a_N)}\right)^2 + 1}} \quad (1)$$

with  $a_1$  as the double amplitude between the first two oscillation peaks ( $p_1$  and  $p_2$ ) and  $a_N$  as the double amplitude between the peaks ( $p_N$  and  $p_{N+1}$ ).  $N$  should be selected so as to get an  $a_N \approx 0.2 \dots 0.4 a_1$ .

**Second step:** Estimation of the natural period  $T_S$  of the seismometer.

If possible switch off all external attenuators (e.g., resistors) to decrease the measuring error. For example: with a damping  $D = 0.2$  the measured period is  $T = 1.02 T_S$ .

The natural period of the seismometer is

$$T_S = T \sqrt{1 - D^2}. \quad (2)$$

### 2.2.2 Electro-dynamical system (moving coil)

The electro-dynamical constant (or generator constant)  $G_S$  of a moving coil system can also be estimated by step transition via its relation to damping  $D$ . The complete seismometer damping  $D_S$  is

$$D_S = D_{S0} + D_G \quad (3)$$

with  $D_{S0}$  as the natural damping of the seismometer (mostly mechanical effects), and with  $D_G$  as the moving coil damping. The latter is caused by an external resistor  $R_a$  shorting the coil (electromagnetic force), i.e.:

$$D_G = \frac{G_S^2 T_S}{4\pi m_S (R_a + R_S)} \quad (4)$$

Except for  $R_a$  and  $T_S$  all other parameters in Equation (4) are documented by the manufacturer and will not change over time. While for pendulum seismometers usually the parameters

- |                      |                         |
|----------------------|-------------------------|
| - $K_S$ [ $kg.m^2$ ] | inertial moment         |
| - $l_0$ [ $m$ ]      | reduced pendulum length |

are given instead of  $m_S$  one gets for geophone systems

- |                                   |                  |
|-----------------------------------|------------------|
| - $m_S$ [ $kg$ ] = $K_S l_0^{-2}$ | seismic mass and |
| - $R_S$ [ $Ohms$ ]                | coil resistance. |

**Note!** When measuring coil resistance don't forget to lock the seismometer.

The evaluation again starts as above with: measuring of the period and damping of the time series as the first step and the estimation of the seismometer's natural period  $T_S$  as the second step.

This is followed by:

**Third step:** Estimation of the seismometer's natural damping  $D_{S0}$ .

The external damping resistor must be removed (open circuit). Then we get, similarly to (1),

$$D_{S0} = \frac{1}{\sqrt{\left(\frac{(N-1)\pi}{\ln(a_1 / a_N)}\right)^2 + 1}} \quad (5)$$

**Fourth step:** External damping.

The external resistor must be set to a value that causes a damping down to 20 - 50% per period. Then we measure the neighbouring amplitudes  $a_1$  (between  $p_1$  and  $p_2$ ) and  $a_2$  (between  $p_2$  and  $p_3$ ;  $p_2$  is used twice to reduce measuring error) and get



$$G_S \quad [\text{Vs/m}] \quad = \quad \sqrt{(D_S - D_{S0}) \frac{4\pi}{T_S} m_S (R_a + R_S)}.$$

(6)

This constant can also be used when calibrating a system by harmonic drive.

**Note!** For pendulum seismometers there are different notations for this constant:

- 1) force/current  $[\text{N/A}] = [\text{Vs/m}]$  and
- 2) torque/current  $[\text{Nm/A}] = [\text{Vs}]$

They are related via the reduced pendulum length  $l_0$  as follows:

$$G_{S1} [V_S] = G_{S2} [V_S / m] \cdot l_0 [m]. \quad (7)$$

### 3 Data: Application to a specific seismometer

Below a typical seismometer parameter list is given.

#### Mechanical constants:

|                            |          |                          |
|----------------------------|----------|--------------------------|
| Natural period             | $T_S$    | ..... s                  |
| Open damping (Attenuation) | $D_{S0}$ | .....                    |
| Reduced pendulum length    | $l_0$    | 0.0785 m                 |
| Inertial moment            | $K_S$    | 0.0201 kg m <sup>2</sup> |
| Seismic mass               | $m_S$    | ..... kg                 |

#### Transducer constants 1 (signal coil):

|                           |          |                         |
|---------------------------|----------|-------------------------|
| Coil resistance           | $R_{S1}$ | 6030 $\Omega$           |
| Electrodynamical constant | $G_{S1}$ | ..... V <sub>S</sub> /m |

#### Transducer constants 2 (calibration coil):

|                           |          |                         |
|---------------------------|----------|-------------------------|
| Coil resistance           | $R_{S2}$ | 835 $\Omega$            |
| Electrodynamical constant | $G_{S2}$ | ..... V <sub>S</sub> /m |

### 4 Tasks

#### Task 1:

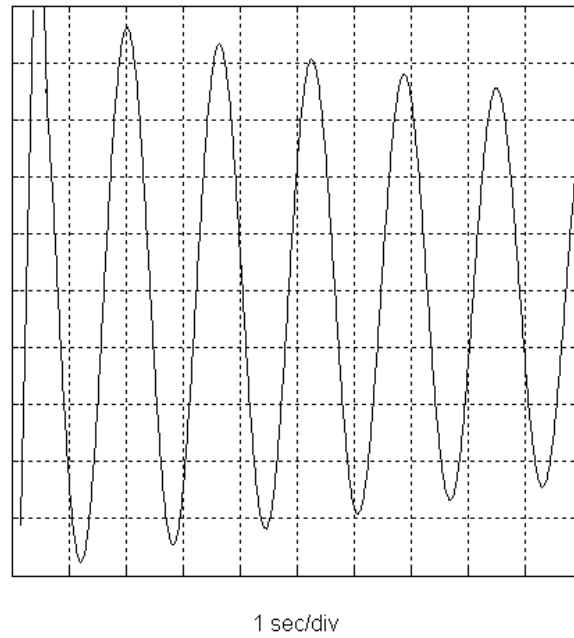
Mark those seismometer parameters which are absolutely necessary for calculating the seismometer response curve (BODE-diagram).

#### Task 2:

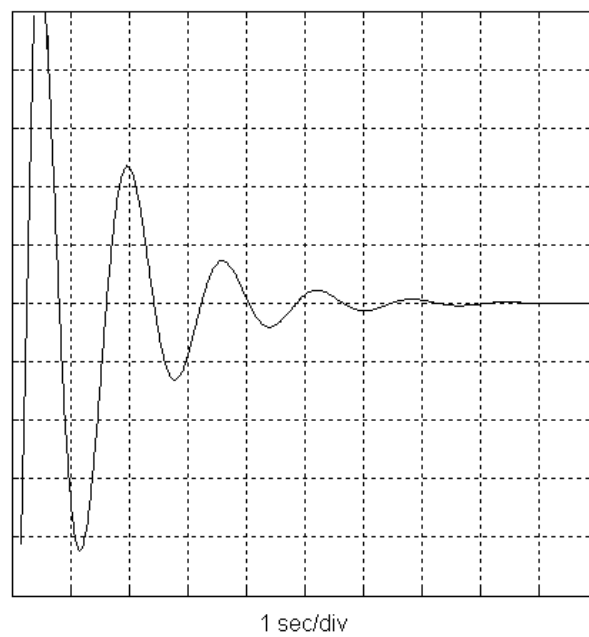
Complete the list above by analysing the related time series plots in Figure 2a - c.

#### Task 3:

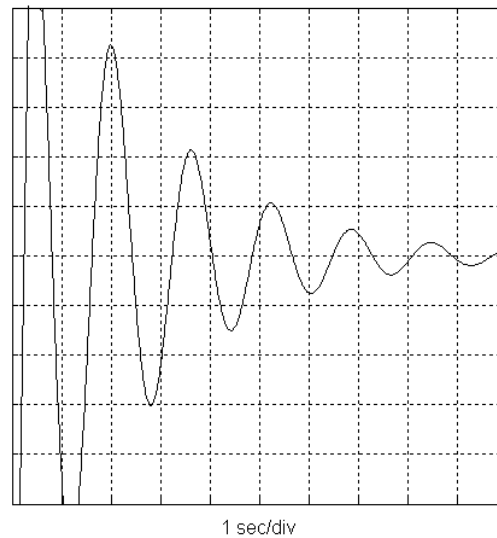
Calculate the current  $I_C$  through the calibration coil which is necessary to deflect the seismometer mass by 1  $\mu\text{m}$  at a frequency  $f = 1 \text{ Hz}$  (see EX 5.3: Seismometer calibration by harmonic drive).



**Figure 2a** Seismometer step response: open circuit.



**Figure 2b** Seismometer step response: signal coil with external resistor  $R_a = 67 \text{ k}\Omega$ .



**Figure 2c** Seismometer step response: calibration coil with external resistor  $R_a = 1 \text{ k}\Omega$ .

## 5 Solutions

### Task 1:

Seismometer parameters absolutely necessary for calculating the seismometer response curve are marked by an asterisk (\*) in the listing below (see Task 2). Additionally required are the *seismometer damping*, consisting of the *open damping* plus the *external* (here: *electrodynamical*) damping.

### Task 2:

Completed list of seismometer parameter:

#### Mechanical constant:

|                            |          |   |                          |     |
|----------------------------|----------|---|--------------------------|-----|
| Natural Period             | $T_{S0}$ | = | 1.617 s                  | (*) |
| Open damping (attenuation) | $D_{S0}$ | = | 0.0102                   |     |
| Reduced Pendulum Length    | $l_0$    | = | 0.0785 m                 |     |
| Inertial Moment            | $K_S$    | = | 0.0201 kg m <sup>2</sup> |     |
| Seismic Mass               | $m_S$    | = | 3.262 kg                 |     |

#### Transducer constants 1 (signal coil):

|                           |          |   |               |     |
|---------------------------|----------|---|---------------|-----|
| Coil Resistance           | $R_{S1}$ | = | 6030 $\Omega$ |     |
| Electrodynamical Constant | $G_{S1}$ | = | 571.1 Vs/m    | (*) |

#### Transducer constants 2 (calibration coil):

|                           |          |   |              |  |
|---------------------------|----------|---|--------------|--|
| Coil Resistance           | $R_{S2}$ | = | 835 $\Omega$ |  |
| Electrodynamical Constant | $G_{S2}$ | = | 67.97 Vs/m   |  |

### Task 3:

In order to deflect the seismometer mass for  $1\text{ }\mu\text{m}$ , a current of  $I_C = 0.12\text{ mA}$  has to be driven through the calibration coil.

## References

Teupser, Ch. (1962). Die Eichung und Prüfung von elektromagnetischen Seismographen, *Freiberger Forschungshefte C130*, 103 pp.

| Title   | Seismometer calibration by harmonic drive                                                                                                                                                                                                                                                                                       |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Jens Bribach</b> (after a manuscript by Christian Teupser <sup>†</sup> )<br>Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences,<br>Dept. 2, Physics of the Earth, Telegrafenberg, D-14473 Potsdam, Germany;<br>Fax: +49 331 288 1266; E-mail: <a href="mailto:brib@gfz-potsdam.de">brib@gfz-potsdam.de</a> |
| Version | October 2009; DOI: 10.2312/GFZ.NMSOP-2_EX_5.3                                                                                                                                                                                                                                                                                   |

If the seismometer possesses an auxiliary magnet and coil assembly, the calibration can be carried out with the aid of an electric current. According to Eq. (5.25) in Chapter 5 and related discussion a current  $i_s$  acts in the same way as a ground acceleration

$$\frac{d^2 x_e}{dt^2} = \frac{G_{s2} l_0^2}{K_s} i_s . \quad (1)$$

where  $G_{s2}$  is the electrodynamic constant of the auxiliary coil (given in [Vs/m]). For other constants see EX 5.2 *Estimating seismometer parameters by STEP function*. It corresponds to a harmonic drive of frequency  $f$  with an equivalent ground displacement

$$x_e = \frac{G_{s2} l_0^2}{4\pi^2 f^2 K_s} i_s . \quad (2)$$

For a translational seismometer, for example a geophone, with seismic mass  $m_s$ , the equivalent seismic displacement is

$$x_e = \frac{G_{s2}}{4\pi^2 f^2 m_s} i_s . \quad (3)$$

Since the output voltage of a geophone with an electromagnetic transducer is

$$E_s = G_{s1} \frac{dz}{dt} , \quad (4)$$

where  $z$  is the displacement of the seismic mass,  $G_{s1}$  is the electrodynamic constant of the signal coil and  $f_s$  the natural frequency, one obtains for a harmonic excitation

$$E_s = \frac{G_{s1} G_{s2} f}{2\pi m_s \sqrt{(f^2 - f_s^2)^2 + 4D_s^2 f^2 f_s^2}} . \quad (5)$$

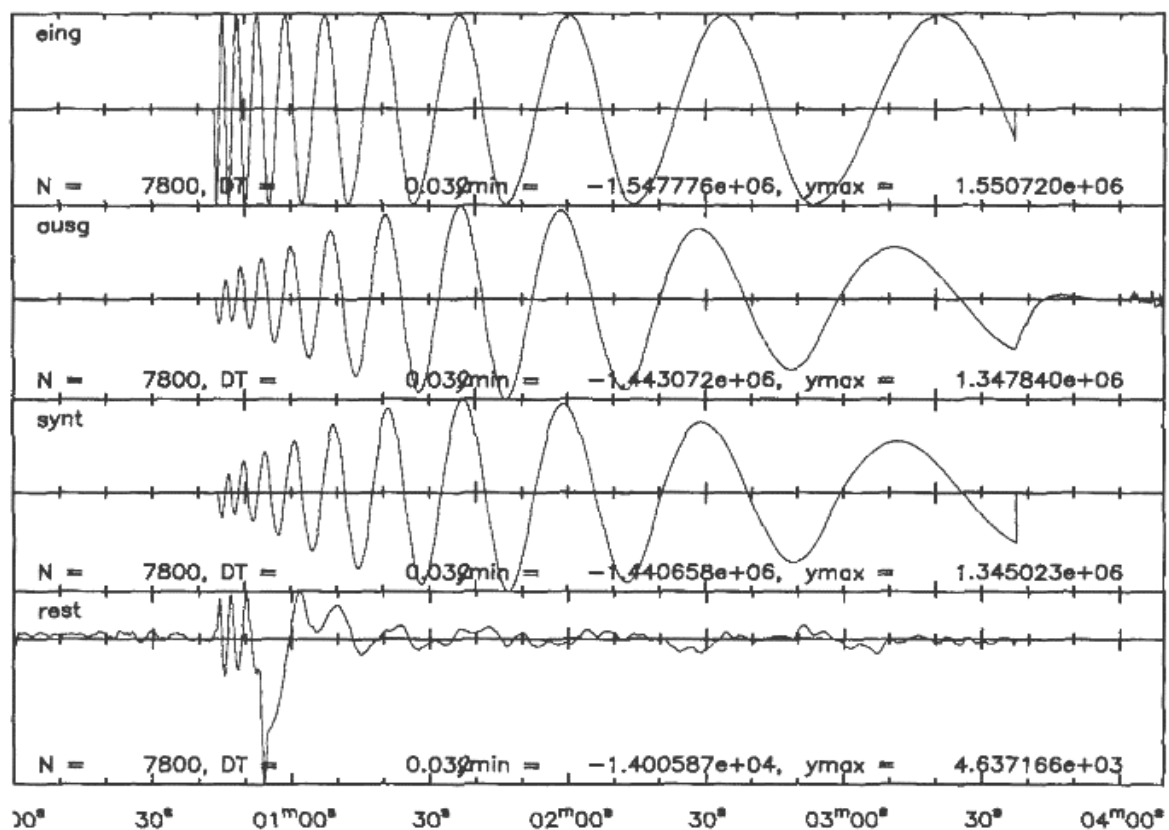
Changing the frequency of the exciting current the output voltage attains a maximum at  $f = f_s$ . This can be used to determine the natural frequency and the damping using an oscilloscope.

| Title   | Seismometer calibration with program CALEX                                                                                                                                   |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Erhard Wielandt (formerly Institute of Geophysics, University of Stuttgart, D - 70184 Stuttgart); E-mail: <a href="mailto:e.wielandt@t-online.de">e.wielandt@t-online.de</a> |
| Version | October 2010; DOI: 10.2312/GFZ.NMSOP-2_EX_5.4                                                                                                                                |

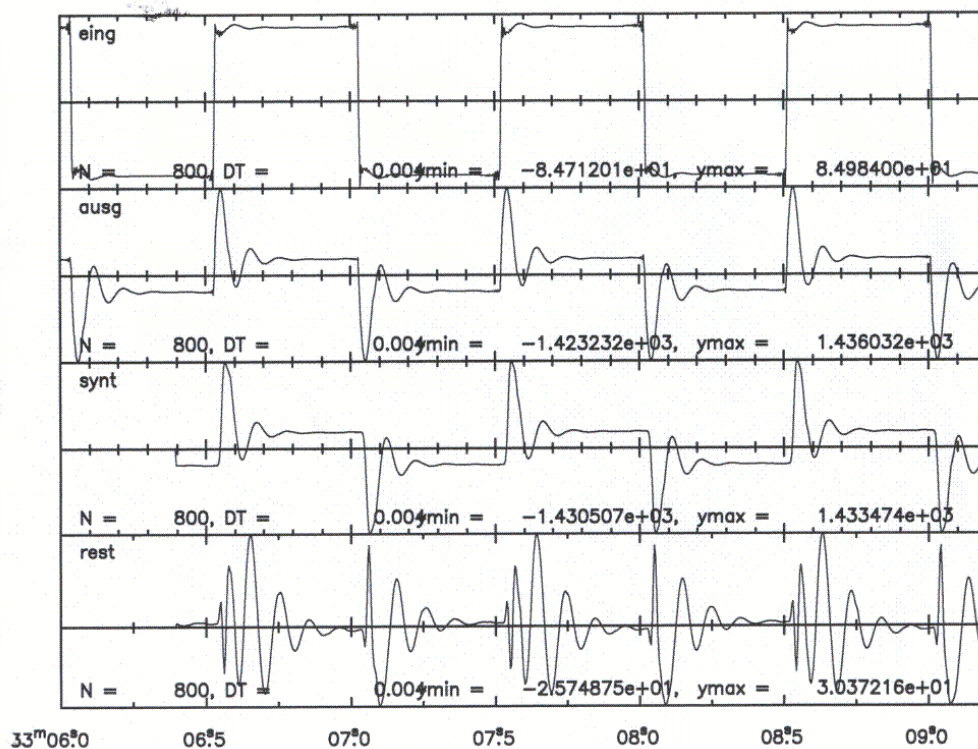
## 1 Software and Data

It is recommended to download the latest versions of the **calex** and **winplot** software from either [www.software-for-seismometry.de](http://www.software-for-seismometry.de) or by right mouse click on these program names in the NMSOP-2 content overview folder *Download Programs & Files*.

Two pairs of input ("eing") and output ("ausg") signals are available. They are shown in Figures 1 and 2; one set is from a broadband seismometer and the other from a short-period geophone. File names may differ depending on the version.



**Figure 1** The input signal into the calibration coil (eing) is a "sweep", a sine-wave whose frequency is automatically tuned from about 2 s to 50 s. It is used for calibration of a 20-sec STS1 seismometer. The second and third traces show the output (ausg) signal and the best fitting synthetics (synt), respectively. The lowermost trace is the residual signal (ausg - synt = rest).



**Figure 2** The input signal is a square wave. It is used for calibration of an undamped 10 Hz geophone in a half-bridge. Note that the input signal appears in the output by coupling through the coil resistance. This would not be the case for a seismic input signal and therefore must not be interpreted as part of the transfer function.

## 2 Tasks

- 1 - Plot the signals on the screen and get an idea of the time scale, of the free period and damping of the sensor, and of the type of response (high-pass, band-pass, low-pass?). You will also need an estimate of polarity and gain between input and output. Compare what you see in the plot to your knowledge of the general properties of seismometer transfer functions.
- 2 - Set up the **calex.par** file for each experiment, as specified in the program description. A sample file is listed there.
- 3 - Run **calex** to determine the exact instrumental constants. Inspect the residual signal and determine its magnitude relative to the output signal (the misfit). Is the misfit caused by improper parametrization of the transfer function, by seismic or environmental noise, or by nonlinear behavior of the sensor? In the latter case, can you guess what the problem might be?
- 4 - Run **calex** again with deliberately offset start parameters, to see if their choice (within a reasonable range) is critical. You may also restrict the analysis to a smaller time window within the record and see if you get different results.

### 3 Solutions

Please read the Program Description for the **calex** and **winplot** routines before continuing. If you use an older program version, it may be necessary to copy 'eing1' onto 'eing' (input signal) and 'ausg1' onto 'ausg' (output signal).

Tasks 1 and 2, first signal: On a Windows PC, you may use the 'winplot' routine for plotting the seismograms. To display the two signals, you have to prepare a small parameter file named winplot.par containing parameters and file names (as specified in the program description of winplot). Under LINUX, use your own plotting routine.

An electromagnetic seismometer, or a broadband-velocity sensor, acts as a band-pass filter for ground accelerations or calibration signals. It gives a maximum response when the signal period equals the free period of the system. By inspecting of the ausg signal, you will recognize that this is in fact a band-pass response, and the free period is around 20 s. The damping is more difficult to estimate, but since the resonance is not sharp, damping must be considerable. Knowing that  $1/\sqrt{2}$  or about 0.7 is a standard value for seismometers, you should start the inversion with this value. The gain between eing and ausg is around unity and the signals have the same phase at the resonance, so the gain parameter in calex.par may be set to 1.

Second signal: copy the data files as above if necessary, and inspect the 'ausg' file. Use the "sub" parameter to remove the resistive component of the output signal (see program description). The free period is obviously around 0.1 s. The damping may be estimated using formula 5.32 (section 5.6.5) of the NMSOP; it is similar to the previous case. The gain between eing and ausg is again near unity and positive.

Task 3, first signal: you should approximately get 19.7 s for the free period, 0.72 for the damping, 1.36 for the gain, 10 ms for the delay, and a rms residual below 0.003. Change the second parameter in 'winplot.par' to 4 and add the file names 'synt' and 'rest'. (New versions of **calex** do this automatically.) Note the transient disturbance in the 'rest' signal at the end of the first minute, which was caused by a person entering the room. Its effect on the result is quite small.

Second signal: the period is 0.102 s, the damping 0.65, the gain is 1.17, and about 48% of the input signal are present in the output signal. The rms residual is again below 0.003. Note the asymmetry in the residual between upgoing and downgoing steps. What you see is mass-position dependent, nonlinear behaviour of the geophone; this is not a bad geophone, but it's no force-balance sensor. You may also notice the small wiggles before and after each step of the input and output signals. (Zoom into a time window with 'seipl02' for better resolution.) The wiggles are not present in the analog signals but are generated by the decimation filters of the digital recorder. Since these filters affect both signals, they don't appear in the transfer function.

Task 4: the results should be nearly independent of the start parameters and of the data window as long as the essential information is preserved in the window.



| Title   | Interpreting Poles and Zeros from SEED headers                                                                                                                               |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Erhard Wielandt (formerly Institute of Geophysics, University of Stuttgart, D - 70184 Stuttgart); E-mail: <a href="mailto:e.wielandt@t-online.de">e.wielandt@t-online.de</a> |
| Version | October 2010; DOI: 10.2312/GFZ.NMSOP-2_EX_5.5                                                                                                                                |

## 1 Aim

The complex transfer function (or the related complex frequency response) of the analog part of a seismograph is a rational function of frequency. Such functions can be specified by corner frequencies and damping constants, by polynomial coefficients, or by their poles and zeros. The latter method is chosen in the IRIS SEED data volumes. For each data channel of each station, the data header contains a list of poles and zeros of the transfer function together with some auxiliary information. IRIS supplies a software library 'evalresp' for extracting and interpreting these parameters. The exercise aims at making you familiar with interpreting poles and zeros in terms of the amplitude response versus frequency.

## 2 Task

Interpret one or more of the annexed SEED headers with respect to the analog part of the seismograph. Sketch the amplitude response for one of the stations as a Bode-diagram on double logarithmic paper. (The digital part is usually of minor interest since it is supposed to have a flat amplitude response and zero phase delay.) Does the header describe a very broadband, broadband or narrowband system? Note that the answer does not only depend on the mathematical form of the response but also on the definition of the input signal - displacement, velocity or acceleration. A broadband seismograph is supposed to have a broadband response to velocity but a broadband accelerometer has a broadband response to acceleration. Be careful with the units - some headers refer to Hertz rather than radians/sec. Check also whether the poles and zeros refer to the Laplace transform or Fourier transform. Can you guess which type of sensor is used? Are the constants nominal or were they determined from an individual calibration?

A little computer program **polzero** in BASIC can be downloaded from either [www.software-for-seismometry.de](http://www.software-for-seismometry.de) or by right mouse click on this program name in the NMSOP-2 content overview folder *Download Programs & Files*. It will do for you the numerical conversions and plot the amplitude response. Use this program to analyze some more of the SEED headers. The stations are:

KIP (Kipapa, Hawaii)  
 KONO (Kongsberg, Norway)  
 KMI (Kunming, China)  
 PFO (Pinion Flat Observatory, California)  
 XAN (Xi'an, China)

## 3 Annex

SEED headers for stations KIP, KONO, KMI, PFO and XAN

## KIP

```

:-----:
RESP.G.KIP..LHE
:-----:
#          << IRIS SEED Reader, Release 4.16 >>
#
#          ===== CHANNEL RESPONSE DATA =====
B050F03    Station:      KIP
B050F16    Network:      G
B052F03    Location:     ??
B052F04    Channel:      LHE
B052F22    Start date:   1988,147
B052F23    End date:     No Ending Time
#
#          +-----+
#          |                                     |
#          |                                     |
#          |                                     |
#          +-----+
#
B053F03    Transfer function type:      B [Analog (Hz)]
B053F04    Stage sequence number:       1
B053F05    Response in units lookup:     M/S - Velocity
B053F06    Response out units lookup:    V - Volts
B053F07    A0 normalization factor:     25.0743
B053F08    Normalization frequency:     0.01
B053F09    Number of zeroes:            2
B053F14    Number of poles:             4
#
#          Complex zeroes:
#          i real      imag      real_error  imag_error
B053F10-13  0  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13  1  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
#
#          Complex poles:
#          i real      imag      real_error  imag_error
B053F15-18  0 -1.964190E-03  1.964190E-03  0.000000E+00  0.000000E+00
B053F15-18  1 -1.964190E-03 -1.964190E-03  0.000000E+00  0.000000E+00
B053F15-18  2 -3.117500E+00  3.909120E+00  0.000000E+00  0.000000E+00
B053F15-18  3 -3.117500E+00 -3.909120E+00  0.000000E+00  0.000000E+00
#
#          +-----+
#          |                                     |
#          |                                     |
#          |                                     |
#          +-----+
#
B058F03    Stage sequence number:       1
B058F04    Gain:                      2.398000E+03
B058F05    Frequency of gain:          1.000000E-02 HZ
B058F06    Number of calibrations:      0
#
#          +-----+
#          |                                     |
#          |                                     |
#          |                                     |
#          +-----+
#
B053F03    Transfer function type:      B [Analog (Hz)]
B053F04    Stage sequence number:       2
B053F05    Response in units lookup:     V - Volts
B053F06    Response out units lookup:    V - Volts
B053F07    A0 normalization factor:     15593.8
B053F08    Normalization frequency:     0.01
B053F09    Number of zeroes:            0
B053F14    Number of poles:             6
#
#          Complex zeroes:
#          i real      imag      real_error  imag_error
#
#          Complex poles:
#          i real      imag      real_error  imag_error
B053F15-18  0 -4.832580E+00  1.273240E+00  0.000000E+00  0.000000E+00
B053F15-18  1 -4.832580E+00 -1.273240E+00  0.000000E+00  0.000000E+00
B053F15-18  2 -3.538230E+00  3.529300E+00  0.000000E+00  0.000000E+00
B053F15-18  3 -3.538230E+00 -3.529300E+00  0.000000E+00  0.000000E+00
B053F15-18  4 -1.295000E+00  4.829390E+00  0.000000E+00  0.000000E+00
B053F15-18  5 -1.295000E+00 -4.829390E+00  0.000000E+00  0.000000E+00
#

```

## KONO

```

:.....:
RESP.IU.KONO.10.LHE
:.....:
#          << IRIS SEED Reader, Release 4.16 >>
#
#          ===== CHANNEL RESPONSE DATA =====
B050F03      Station:      KONO
B050F16      Network:     IU
B052F03      Location:    10
B052F04      Channel:     LHE
B052F22      Start date:  1999,040,13
B052F23      End date:    No Ending Time
#
#          +-----+
#          +               | Response (Poles & Zeros), KONO ch LHE |
#          +-----+
#
B053F03      Transfer function type:      A [Laplace Transform (Rad/sec)]
B053F04      Stage sequence number:      1
B053F05      Response in units lookup:    M/S - Velocity in Meters Per Second
B053F06      Response out units lookup:   V - Volts
B053F07      A0 normalization factor:    7.1367E+07
B053F08      Normalization frequency:    0.1
B053F09      Number of zeroes:           2
B053F14      Number of poles:            5
#          Complex zeroes:
#          i real      imag      real_error      imag_error
B053F10-13    0  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13    1  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
#          Complex poles:
#          i real      imag      real_error      imag_error
B053F15-18    0 -3.701000E-02  3.701000E-02  0.000000E+00  0.000000E+00
B053F15-18    1 -3.701000E-02 -3.701000E-02  0.000000E+00  0.000000E+00
B053F15-18    2 -1.979000E+02  1.979000E+02  0.000000E+00  0.000000E+00
B053F15-18    3 -1.979000E+02 -1.979000E+02  0.000000E+00  0.000000E+00
B053F15-18    4 -9.111000E+02  0.000000E+00  0.000000E+00  0.000000E+00
#
#          +-----+
#          +               | Channel Gain, KONO ch LHE |
#          +-----+
#
B058F03      Stage sequence number:      1
B058F04      Gain:                     2.026400E+04
B058F05      Frequency of gain:         2.000000E-02 HZ
B058F06      Number of calibrations:     0
#
#          +-----+
#          +               | Response (Coefficients), KONO ch LHE |
#          +-----+
#
B054F03      Transfer function type:      D
B054F04      Stage sequence number:      2
B054F05      Response in units lookup:    V - Volts
B054F06      Response out units lookup:   COUNTS - Digital Counts
B054F07      Number of numerators:       0
B054F10      Number of denominators:     0
#
#          +-----+
#          +               | Decimation, KONO ch LHE |
#          +-----+
#
B057F03      Stage sequence number:      2
B057F04      Input sample rate:          5.120000E+03
B057F05      Decimation factor:          1
B057F06      Decimation offset:          0
B057F07      Estimated delay (seconds):  0.000000E+00

```

## KMI

```

#
# .....
RESP.CD.KMI..LHZ
# .....
# << IRIS SEED Reader, Release 4.16 >>
#
# ===== CHANNEL RESPONSE DATA =====
B050F03      Station:      KMI
B050F16      Network:     CD
B052F03      Location:    ??
B052F04      Channel:     LHZ
B052F22      Start date:  1986,159
B052F23      End date:    1996,108
#
# -----+-----+-----+-----+
#          +          | Response (Poles & Zeros), KMI ch LHZ |
#          +          +-----+-----+
#
B053F03      Transfer function type:      A [Laplace Transform (Rad/sec)]
B053F04      Stage sequence number:      1
B053F05      Response in units lookup:    M - Earth Displacement in Meters
B053F06      Response out units lookup:   COUNTS - Digital Counts
B053F07      A0 normalization factor:    0.000492889
B053F08      Normalization frequency:    0.04
B053F09      Number of zeroes:           4
B053F14      Number of poles:            10
#
#      Complex zeroes:
#      i      real      imag      real_error      imag_error
B053F10-13   0  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13   1  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13   2  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13   3  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
#
#      Complex poles:
#      i      real      imag      real_error      imag_error
B053F15-18   0 -2.221000E-01  2.221000E-01  0.000000E+00  0.000000E+00
B053F15-18   1 -2.221000E-01 -2.221000E-01  0.000000E+00  0.000000E+00
B053F15-18   2 -7.405000E-03  7.405000E-03  0.000000E+00  0.000000E+00
B053F15-18   3 -7.405000E-03 -7.405000E-03  0.000000E+00  0.000000E+00
B053F15-18   4 -2.023000E-01  5.420000E-02  0.000000E+00  0.000000E+00
B053F15-18   5 -2.023000E-01 -5.420000E-02  0.000000E+00  0.000000E+00
B053F15-18   6 -1.481000E-01  1.481000E-01  0.000000E+00  0.000000E+00
B053F15-18   7 -1.481000E-01 -1.481000E-01  0.000000E+00  0.000000E+00
B053F15-18   8 -5.420000E-02  2.023000E-01  0.000000E+00  0.000000E+00
B053F15-18   9 -5.420000E-02 -2.023000E-01  0.000000E+00  0.000000E+00
#
#          +          +-----+-----+
#          +          | Channel Sensitivity, KMI ch LHZ |
#          +          +-----+-----+
#
B058F03      Stage sequence number:      0
B058F04      Sensitivity:                 1.800000E+09
B058F05      Frequency of sensitivity:    4.000000E-02 HZ
B058F06      Number of calibrations:      0
#

```

## PFO

```

:.....:
RESP.TS.PFO..LHZ
:.....:
#      << IRIS SEED Reader, Release 4.16 >>
#
#      ===== CHANNEL RESPONSE DATA =====
B050F03      Station:      PFO
B050F16      Network:     TS
B052F03      Location:    ??
B052F04      Channel:     LHZ
B052F22      Start date:  1990,304
B052F23      End date:    No Ending Time
#
#      =====
#      +-----+-----+-----+-----+
#      +               | Response (Poles & Zeros),   PFO ch LHZ |
#      +-----+-----+-----+-----+
#
B053F03      Transfer function type:      A [Laplace Transform (Rad/sec)]
B053F04      Stage sequence number:      1
B053F05      Response in units lookup:    M/S - Velocity in Meters Per Second
B053F06      Response out units lookup:   V - Volts
B053F07      A0 normalization factor:    3948.58
B053F08      Normalization frequency:    0.02
B053F09      Number of zeroes:           2
B053F14      Number of poles:            4
#      Complex zeroes:
#      i real      imag      real_error      imag_error
B053F10-13    0  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13    1  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
#      Complex poles:
#      i real      imag      real_error      imag_error
B053F15-18    0 -1.234000E-02  1.234000E-02  0.000000E+00  0.000000E+00
B053F15-18    1 -1.234000E-02 -1.234000E-02  0.000000E+00  0.000000E+00
B053F15-18    2 -3.918000E+01  4.912000E+01  0.000000E+00  0.000000E+00
B053F15-18    3 -3.918000E+01 -4.912000E+01  0.000000E+00  0.000000E+00
#
#      +-----+-----+-----+-----+
#      +               | Channel Gain,      PFO ch LHZ |
#      +-----+-----+-----+-----+
#
B058F03      Stage sequence number:      1
B058F04      Gain:                      2.122720E+03
B058F05      Frequency of gain:          2.000000E-02 HZ
B058F06      Number of calibrations:      0
#
#      +-----+-----+-----+-----+
#      +               | Response (Coefficients),   PFO ch LHZ |
#      +-----+-----+-----+-----+
#
B054F03      Transfer function type:      D
B054F04      Stage sequence number:      2
B054F05      Response in units lookup:    V - Volts
B054F06      Response out units lookup:   COUNTS - Digital Counts
B054F07      Number of numerators:       43
B054F10      Number of denominators:     0
#      Numerator coefficients:
#      i, coefficient, error
B054F08-09    0 -3.557280E-09 -7.114550E-11
B054F08-09    1  3.273000E-06  6.546030E-08
B054F08-09    2 -3.791030E-04 -7.582060E-06
B054F08-09    3 -2.870530E-03 -5.741070E-05
B054F08-09    4 -2.949110E-03 -5.898210E-05
B054F08-09    5  3.191820E-03  6.383630E-05
B054F08-09    6 -2.121360E-03 -4.242730E-05
B054F08-09    7 -5.931070E-04 -1.186210E-05
B054F08-09    8  4.816940E-03  9.633870E-05

```

## XAN

```

:.....:
RESP.IC.XAN..LHE
:.....:
#      << IRIS SEED Reader, Release 4.16 >>
#
#      ----- CHANNEL RESPONSE DATA -----
B050F03      Station:      XAN
B050F16      Network:      IC
B052F03      Location:      ??
B052F04      Channel:      LHE
B052F22      Start date:   1992,334
B052F23      End date:     1995,149
#      =====
#      +-----+-----+-----+-----+
#      +               | Response (Poles & Zeros),   XAN ch LHE |
#      +-----+-----+-----+-----+
#
B053F03      Transfer function type:      A [Laplace Transform (Rad/sec)]
B053F04      Stage sequence number:      1
B053F05      Response in units lookup:    M/S - Velocity in Meters Per Second
B053F06      Response out units lookup:   V - Volts
B053F07      A0 normalization factor:    5.96806E+07
B053F08      Normalization frequency:    0.02
B053F09      Number of zeroes:           2
B053F14      Number of poles:            5
#      Complex zeroes:
#      i      real      imag      real_error      imag_error
B053F10-13   0  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13   1  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
#      Complex poles:
#      i      real      imag      real_error      imag_error
B053F15-18   0 -3.564700E-02 -3.687900E-02  0.000000E+00  0.000000E+00
B053F15-18   1 -3.564700E-02  3.687900E-02  0.000000E+00  0.000000E+00
B053F15-18   2 -2.513300E+02  0.000000E+00  0.000000E+00  0.000000E+00
B053F15-18   3 -1.310400E+02 -4.672900E+02  0.000000E+00  0.000000E+00
B053F15-18   4 -1.310400E+02  4.672900E+02  0.000000E+00  0.000000E+00
#
#      +-----+-----+-----+-----+
#      +               | Channel Gain,      XAN ch LHE |
#      +-----+-----+-----+-----+
#
B058F03      Stage sequence number:      1
B058F04      Gain:                      1.500000E+03
B058F05      Frequency of gain:          2.000000E-02 HZ
B058F06      Number of calibrations:     0
#
#      +-----+-----+-----+-----+
#      +               | Response (Coefficients),   XAN ch LHE |
#      +-----+-----+-----+-----+

```

### 3 Solutions

|      |                                                                                                                                                                                                                                                                                                                                                                              |
|------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| KIP  | velocity very broadband, lower corner 360 s, upper corner 0.2 s<br>Obviously an older STS1-VBB seismometer. No extra filters.<br>Nominal parameters.                                                                                                                                                                                                                         |
| KONO | velocity broadband, lower corner 120 s, upper corner 44.5 Hz<br>Must be an STS2 or a CMG3-T. Nominal parameters. Additional low-pass Filter at 145 Hz.                                                                                                                                                                                                                       |
| KMI  | narrowband LP as a displacement sensor, but better characterized as a long-period acceleration sensor. Response is flat to acceleration from 30 s to 600 s. The sensor must be an old STS1 (20 s). A 6 <sup>th</sup> -order Butterworth low-pass filter limits the bandwidth at 30 s; this would today be done with digital filters in the recorder. Parameters are nominal. |
| PFO  | velocity very broadband, lower corner 360 s, upper corner 0.1 s.<br>A modern STS1-VBB. No extra filters. Nominal parameters.                                                                                                                                                                                                                                                 |
| XAN  | velocity broadband, lower corner 120 s, upper corner 44 Hz.<br>Probably an STS2 or a CMG3-T seismometer. Additional low-pass filter at 77 Hz. Parameters were probably measured.                                                                                                                                                                                             |

| Title   | The response of a WWSSN-LP seismograph                                                                                                                                       |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Erhard Wielandt (formerly Institute of Geophysics, University of Stuttgart, D - 70184 Stuttgart); E-mail: <a href="mailto:e.wielandt@t-online.de">e.wielandt@t-online.de</a> |
| Version | May 2010; DOI: 10.2312/GFZ.NMSOP-2_EX_5.6                                                                                                                                    |

## 1 Introduction

The long-period seismograph of the now obsolete WWSSN (Worldwide Standardized Seismograph Network) consisted of a long-period electrodynamic seismometer normally tuned to a free period of 15 sec, and a long-period mirror-galvanometer with a free period around 90 sec. (In order to avoid confusion with the frequency variable  $s = j\omega$  of the Laplace transformation, we use the non-standard abbreviation „sec“ for seconds in the present subsection.) The WWSSN seismograms were recorded on photographic paper rotating on a drum. The simple design of this system gives us an opportunity to write down several equivalent forms of the transfer function explicitly. Since input and output signals can be measured as displacements, it is natural to describe the system by its displacement response. The absolute value of this response is the frequency-dependent magnification. We assume the numerical damping to be 0.6 for the seismometer and 0.9 for the galvanometer. For simplicity, we ignore the small difference between the free periods and damping constants of the physical subunits and those that appear in the transfer function of the coupled system. We further assume that the effective generator constant of the electromagnetic velocity transducer in the seismometer is 200 V per m/s, and the photographic trace is deflected by 0.3935 mm per microvolt.

## 2 Tasks

- **Write down** the transfer function for displacement as a function of the Laplace variable  $s$ , using symbolic algebra (that is, representing free period and damping by mathematical symbols, not their numerical values)
- **Write down** the formula for the “magnification curve” (amplitude response as a function of the angular frequency), using symbolic algebra
- **Represent** the magnification curve as a “Bode plot”, that is, approximate it by asymptotic straight lines in a double-logarithmic plot (as in exercise EX 5.1)
- **Determine** the poles and zeros of the transfer function in symbolic algebra, and evaluate them numerically
- **Sketch** the position of poles and zeros in the complex  $s$  plane
- **Write down** the transfer function as the ratio of two polynomials with numerical coefficients



### 3 Solution

As shown in Chapter 5, section 5.2.8, Eq.(5.23), the transfer function of an electromagnetic seismometer (input: displacement, output: voltage) is

$$H_s(s) = Es^3 / (s^2 + 2s\omega_s h_s + \omega_s^2) \quad (1)$$

where  $\omega_s = 2\pi/T_s$  is the angular eigenfrequency and  $h_s$  the numerical damping. (see EX 5.2 for a practical determination of these parameters.) The factor  $E$  is the generator constant of the electromagnetic transducer, for which we assume a value of 200 Vs/m.

The galvanometer is a second-order low-pass filter and has the transfer function

$$H_g(s) = \gamma\omega_g^2 / (s^2 + 2s\omega_g h_g + \omega_g^2) \quad (2)$$

Here  $\gamma$  is the responsivity (in meters per volt) of the galvanometer with the given coupling network and optical path. We use a value of 393.5 m/V, which gives the desired overall magnification. The overall transfer function  $H_d$  of the seismograph is obtained in our simplified treatment as the product of the factors given in Eqs. (1) and (2):

$$H_d(s) = \frac{Cs^3}{(s^2 + 2s\omega_s h_s + \omega_s^2)(s^2 + 2s\omega_g h_g + \omega_g^2)} \quad (3)$$

The numerical values of the constants are  $C = E\gamma\omega_g^2 = 383.6/\text{sec}$ ,  $2\omega_s h_s = 0.5027/\text{sec}$ ,  $\omega_s^2 = 0.1755/\text{sec}^2$ ,  $2\omega_g h_g = 0.1257/\text{sec}$ , and  $\omega_g^2 = 0.00487/\text{sec}^2$ .

As the input and output signals are displacements, the absolute value  $|H_d(s)|$  of the transfer function is simply the frequency-dependent magnification of the seismograph. The gain factor  $C$  has the physical dimension  $\text{sec}^{-1}$ , so  $H_d(s)$  is in fact a dimensionless quantity.  $C$  itself is however not the magnification of the seismograph. To obtain the magnification at the angular frequency  $\omega$ , we have to evaluate  $M(\omega) = |H_d(j\omega)|$ :

$$M(\omega) = \frac{C\omega^3}{\sqrt{(\omega_s^2 - \omega^2)^2 + 4\omega^2\omega_s^2 h_s^2} \sqrt{(\omega_g^2 - \omega^2)^2 + 4\omega^2\omega_g^2 h_g^2}} \quad (4)$$

Eq. (3) is a factorized form of the transfer function in which we still recognize the sub-units of the system. We may of course insert the numerical constants and expand the denominator into a fourth-order polynomial

$$H_d(s) = 383.6s^3 / (s^4 + 0.6283s^3 + 0.2435s^2 + 0.0245s + 0.000855) \quad (5)$$

but the only advantage of this form would be its shortness.

The poles and zeros of the transfer function are most easily determined from Eq. (3). We read immediately that a triple zero is present at  $s = 0$ . Each factor  $s^2 + 2s\omega_0 h + \omega_0^2$  in the denominator has the zeros

$$s_0 = \omega_0(-h \pm j\sqrt{1-h^2}) \quad \text{for } h < 1$$

$$s_0 = \omega_0(-h \pm \sqrt{h^2 - 1}) \quad \text{for } h \geq 1$$

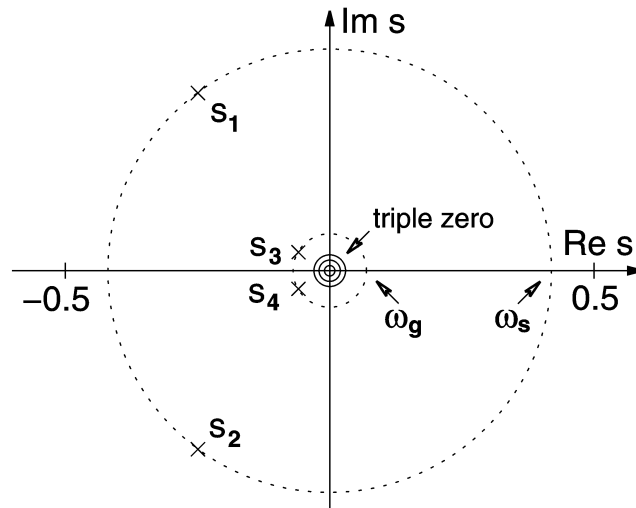
so the poles of  $H_d(s)$  in the complex  $s$  plane are (see Figure 1):

$$s_1 = \omega_s(-h_s + j\sqrt{1-h_s^2}) = -0.2513 + 0.3351j \quad [\text{sec}^{-1}]$$

$$s_2 = \omega_s(-h_s - j\sqrt{1-h_s^2}) = -0.2513 - 0.3351j \quad [\text{sec}^{-1}]$$

$$s_3 = \omega_g(-h_g + j\sqrt{1-h_g^2}) = -0.0628 + 0.0304j \quad [\text{sec}^{-1}]$$

$$s_4 = \omega_g(-h_g - j\sqrt{1-h_g^2}) = -0.0628 - 0.0304j \quad [\text{sec}^{-1}]$$



**Figure 1** Position of the poles of the WWSSN-LP system in the complex  $s$  plane.

In order to reconstruct  $H_d(s)$  from its poles and zeros and the gain factor, we write

$$H_d(s) = \frac{Cs^3}{(s-s_1)(s-s_2)(s-s_3)(s-s_4)}. \quad (5)$$

It is now convenient to pair-wise expand the factors of the denominator into second-order polynomials:

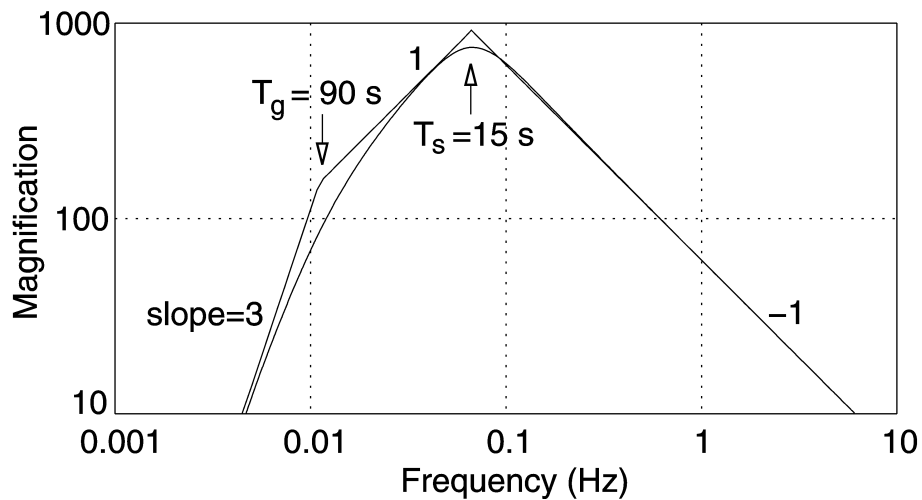
$$H_d(s) = \frac{Cs^3}{(s^2 - s(s_1 + s_2) + s_1s_2)(s^2 - s(s_3 + s_4) + s_3s_4)}. \quad (6)$$

This makes all coefficients real because  $s_2 = s_1^*$  and  $s_4 = s_3^*$ . Since  $s_1 + s_2 = -2\omega_s h_s$ ,  $s_1s_2 = \omega_s^2$ ,  $s_3 + s_4 = -2\omega_g h_g$ , and  $s_3s_4 = \omega_g^2$ , Eq. (6) is in fact the same as Eq. (3). We may of course also reconstruct  $H_d(s)$  from the numerical values of the poles and zeros. Dropping the physical units, we obtain

$$H_d(s) = \frac{383.6s^3}{(s^2 + 0.5027s + 0.1755)(s^2 + 0.1257s + 0.00487)} \quad (7)$$

in agreement with Eq.(4).

Figure 2 shows the corresponding amplitude response of the WWSSN seismograph as a function of frequency. The maximum magnification is 750 near a period of 15 sec. The slopes of the asymptotes are at each frequency determined by the dominant powers of  $s$  in the numerator and denominator of the transfer function. Generally, the low-frequency asymptote has the slope  $m$  (the number of zeros, here = 3) and the high-frequency asymptote has the slope  $m-n$  (where  $n$  is the number of poles, here = 4). What happens in between depends on the position of the poles in the complex  $s$  plane. Generally, a pair of poles  $s_1, s_2$  corresponds to a second-order corner of the amplitude response with  $\omega_0^2 = s_1s_2$  and  $2\omega_0h = -s_1 - s_2$ . A single pole at  $s_0$  is associated with a first-order corner with  $\omega_0 = s_0$ . The poles and zeros however do not indicate whether the respective subsystem is a low-pass, high-pass, or band-pass filter. This does not matter; the corners bend the amplitude response downward in each case. In the WWSSN-LP system, the low-frequency corner at 90 sec corresponding to the pole pair  $s_1, s_2$  reduces the slope of the amplitude response from 3 to 1, and the corner at 15 sec corresponding to the pole pair  $s_3, s_4$  reduces it further from 1 to -1.



**Figure 2** Amplitude response of the WWSSN-LP system with asymptotes (Bode plot).

Looking at the transfer function  $H_s$  in Eq. (1) of the electromagnetic seismometer alone, we see that the low-frequency asymptote has the slope 3 because of the triple zero in the numerator. The pole pair  $s_1, s_2$  corresponds to a second-order corner in the amplitude response at  $\omega_s$  which reduces the slope to 1. The resulting response is shown in a normalized form in the upper right panel of Fig. 5.3 in Chapter 5. As stated [there](#) in section 5.2.6 under point 3, this case of  $n < m$  can only be an approximation in a limited bandwidth. In modern seismograph systems, the upper limit of the bandwidth is usually set by an analog or digital cut-off (anti-alias) filter.

As we have shown in section 5.2.8, the classification of a subsystem as a high-pass, band-pass or low-pass filter may be a matter of definition rather than hardware; it depends on the type of ground motion (displacement, velocity, or acceleration) to which it relates. We also notice that interchanging  $\omega_s, h_s$  with  $\omega_g, h_g$  will change the gain factor  $C$  in the numerator of Eq. (4) from  $E\gamma\omega_g^2$  to  $E\gamma\omega_s^2$  and thus the gain, but will leave the denominator and therefore the shape of the response unchanged. While the transfer function is insensitive to arbitrary factorization, the hardware may be quite sensitive, and certain engineering rules must be observed when a given transfer function is realized in hardware. For example, it would have been difficult to realize a WWSSN seismograph with a 15 sec galvanometer and a 90 sec seismometer; the restoring force of a Lacoste-type suspension cannot be made small enough without becoming unstable.

| Topic   | Strainmeters                                                                                                                                                                    |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Walter Zürn, Black Forest Observatory, Universities Karlsruhe/Stuttgart, Heubach 206, D - 77709 Wolfach; E-mail: <a href="mailto:walter.zuern@kit.edu">walter.zuern@kit.edu</a> |
| Version | September 2009; DOI: 10.2312/GFZ.NMSOP-2_IS_5.1                                                                                                                                 |

|                                            | page |
|--------------------------------------------|------|
| <b>1 Introduction</b>                      | 1    |
| <b>2 Types of strainmeters</b>             | 1    |
| 2.1 Linear strainmeters                    | 1    |
| 2.2 Volumetric strainmeters                | 3    |
| <b>3 Properties</b>                        | 3    |
| 3.1 Sensitivity                            | 3    |
| 3.2 Frequency response                     | 4    |
| 3.3 Calibration                            | 4    |
| 3.4 Direction sensitivity                  | 5    |
| 3.5 Local arrays of linear strainmeters    | 5    |
| 3.6 Local phase velocity                   | 5    |
| 3.7 Effects of local heterogeneity         | 6    |
| <b>4 Some results</b>                      | 6    |
| <b>5 Strain- vs. inertial seismometers</b> | 8    |
| <b>6 References</b>                        | 10   |

## 1 Introduction

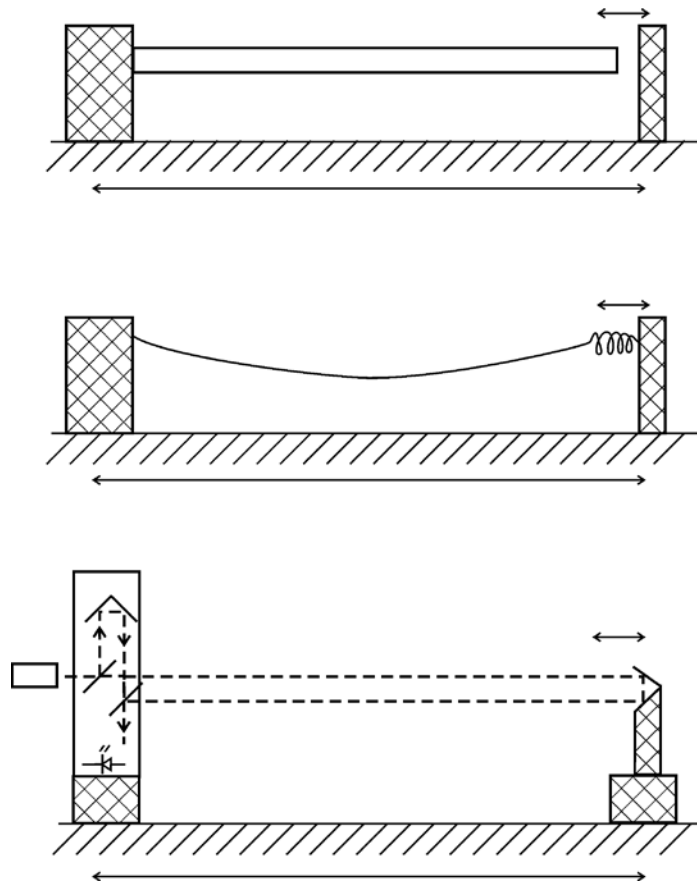
In contrast to inertial seismometers, which respond to ground acceleration and thus to the second derivative of the displacement with respect to time, strain-seismometers, commonly termed extensometers or strainmeters, respond to the spatial derivatives of the displacement field of the incoming seismic wave or in other words, to a combination of the components of the wave's strain tensor. For this reason, strainmeters are inherently instruments which are sensitive to phenomena with "zero" (i.e., very low) frequency and thus particularly suitable for studying crustal deformations due to solid Earth tides and normal mode oscillations of the Earth. More precisely, linear strainmeters record the changes of the distance between the two points at which the instrument is fixed to the ground, while volumetric strainmeters (dilatometers) record the changes of a volume standard which is imbedded in the ground. An excellent, thorough and comprehensive review of strainmeters with an extensive bibliography was written by Agnew (1986) and is still up to date. Strainmeters were introduced into seismology already by Milne (1888).

## 2 Types of strainmeters

### 2.1 Linear strainmeters

This is the most frequently deployed type of strainmeter. The changes  $dL(t)$  with time  $t$  of a fixed distance  $L$  between two points of the Earth are measured with the help of some length

standard. Only a handful of strainmeters ever measured non-horizontal strains and all these were vertical, by far the majority was measuring horizontal strain. Agnew (1986) distinguishes rod strainmeters, wire strainmeters and laser strainmeters (see Figure 1).



**Figure 1** Schematics of the most frequent strainmeter designs. From top to bottom: rod-, (tensioned) wire- and laser strainmeters. The top two must be equipped with displacement transducers at the right. The bottom sketch indicates the laser, two beam splitters, one corner cube reflector and some sensor (symbol of photodiode) able to detect the change in the interference fringes due to the relative motion of the corner cube reflector on the pedestal to the right.

The length standard should be very stable against all kinds of environmental variables, especially temperature, air pressure and humidity. Because of these requirements, rods are mostly made of quartz, invar or superinvar and wires from invar or carbon fiber. Long rods somehow must be supported without friction, while wires must be tensioned. The length changes are detected by displacement or velocity transducers very similar to the ones used in modern inertial seismometers (see 5.3.7 and 5.3.8). One example of an Invar-rod strainmeter and its installation is described by Fix and Sherwin (1970). A frequently used type of wire strainmeter is described by King and Bilham (1976) and an installation in Widmer et al. (1992). Very short rod strainmeters can be placed in borehole packages, which then must be cemented to the borehole wall. An instrument of this type is described by Gladwin (1984). Laser strainmeters use the wavelength of light as a length standard and an unequal-arm

Michelson interferometer for detection of strains. The interference fringes between the light beams along the long arm (the measuring distance  $L$ ) and a short reference arm are observed with different methods, details of which can be found in the references given by Agnew (1986). It is clear that simple fringe counting necessitates  $L$  to be very large to obtain high enough sensitivity: the wavelength of light is of the order of 500 nm, which is 10 times the amplitude of the Earth tide if  $L$  equals 1 m. Therefore the fringe counting laser strainmeters at Pinion Flat Observatory in California are more than 700 m long (Agnew, 1986; Wyatt et al., 1982, Agnew et al. 1989). Two other laser strainmeters with refined methods to determine the length changes using the fringes are described by Levine and Hall (1972) and by Goult et al. (1974). The smallness of the expected signals with respect to the local noise from changes in the environmental conditions necessitates either installation deep underground in mines or boreholes or possibilities for anchoring the mounts to points deep in the ground (Wyatt et al., 1982). A typical installation depth with hope for success is larger than about 30 m below surface. Shielding the instruments is also mandatory and not as easy as for the much smaller inertial seismometers.

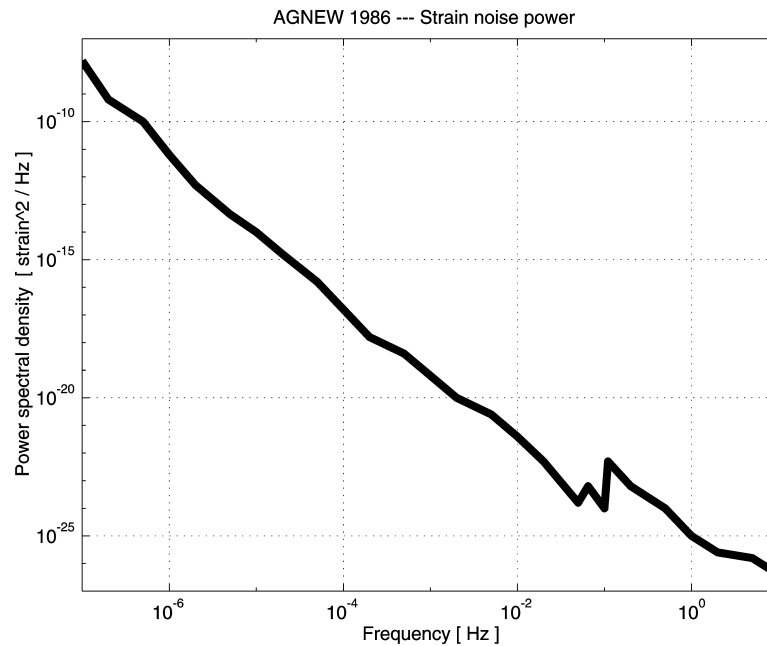
## 2.2 Volumetric strainmeters

Volumetric strainmeters or dilatometers measure the change  $dV(t)$  of a certain volume  $V$ . A borehole instrument of this kind was constructed by Sacks et al. (1971) and widely deployed, especially in Japan. The volume changes are sensed by a liquid-filled tube which is cemented into the borehole. The deformation of the volume causes the liquid to expand or contract bellows, the movement of which is transmitted by a lever arm to displacement transducers. In passing it should be mentioned that water wells drilled into confined aquifers act as sensors for the strain tensor in their vicinity, because applied strains force water in and out of the well, if it is open to the aquifer. Either the level of the water surface or water pressure at a constant depth are measured (e.g. Kümpel, 1992).

## 3 Properties

### 3.1 Sensitivity

A typical strain amplitude  $dL/L$  (no dimension!) of the solid Earth tide is 50 nano ( $50 \cdot 10^{-9}$ ), while Widmer et al. (1992) reported 10 pico ( $10^{-11}$ ) for the fundamental toroidal mode  ${}_0T_2$  of the Earth (see Fig. 2.22) excited by the 1989 Macquarie-Ridge event with a moment magnitude of 8.2. Note that these numbers correspond to one wavelength of (red) light in 10 m and 50 km, respectively. It is obvious that the necessary resolution of the transducer depends critically on the dimension of the strainmeter, i. e. the longer  $L$ , the less resolution is needed to achieve a certain resolution in strain. Agnew (1986) shows a power spectrum of Earth's strain noise (Figure 2) and describes the sources of the noise as follows: above 0.5 Hz body wave energy, machinery and wind-blown vegetation, between 0.05 and 0.5 Hz marine microseisms with high temporal variability and from 1 mHz to 0.05 Hz atmospheric pressure changes deforming the ground (wind turbulence, infrasound) (see 4.3). Below 1 mHz the sources are hard to identify, possibilities being thermoelastic deformations, pore pressure and groundwater changes (Evans and Wyatt, 1984), or air pressure changes. Instrumental effects play an important role at the lowest frequencies (drift) and are not easily ruled out (Zadro and Braitenberg, 1999).



**Figure 2** Power spectral density in (strain squared)/Hz as published by Agnew (1986). This is from the 730 m-NW laser horizontal strainmeter at Pinion Flat, California. This instrument is installed at the surface, but referenced with "optical anchors" to a depth of 30 m (Wyatt et al. 1982). See discussion of noise sources in text.

### 3.2 Frequency response

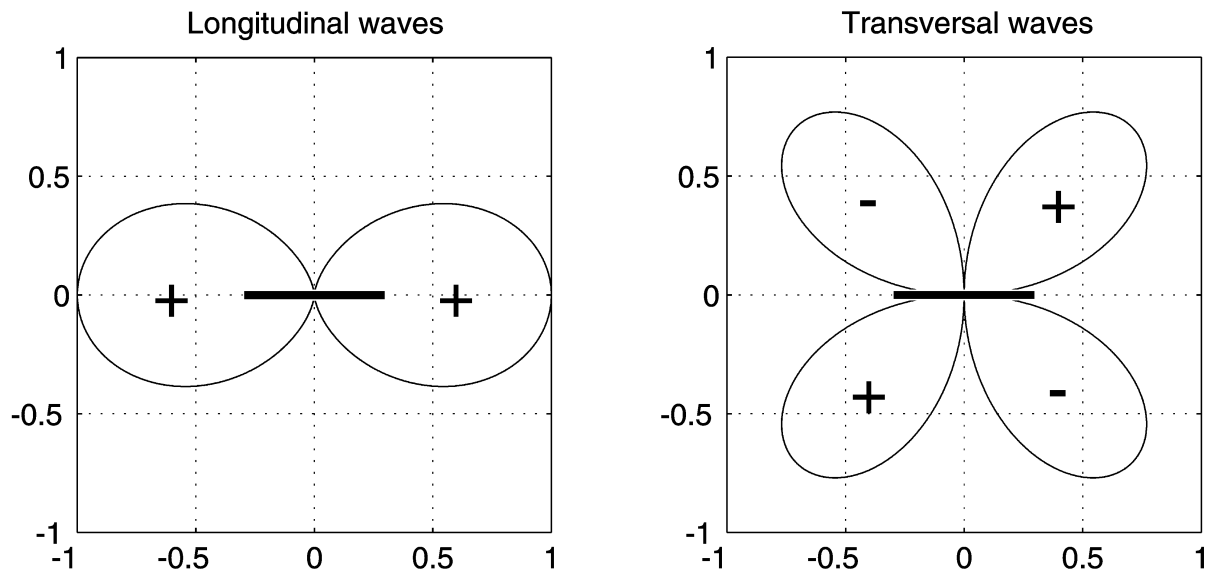
Lomnitz (1997) discusses the amplification of strainmeters for the case, when the seismic wavelength becomes comparable to the length  $L$ . However, the wavelengths of seismic waves are normally much larger than  $L$ , therefore the dependence of the amplification on the wavelength can be neglected in all but a few exceptional cases. So, basically strainmeters are extremely broadband instruments whose range extends from zero frequency to frequencies of 1 Hz and higher. However, the upper limit of this range depends very critically on the design of the individual instrument since all devices possess parasitic resonances at high frequencies.

### 3.3 Calibration

The method for absolute calibration of a strainmeter depends on the individual instrument. In-situ calibration is highly recommended, that is calibration of the installed device is preferable to calibration calculation from components calibrated in the laboratory. Small displacements of that end of a linear strainmeter, which is fixed to the rock simulates ground motion. Uncertainties arise from the definition of the effective length  $L$ , because basically the piers are part of the instruments and have some extent in length. Greater difficulties arise for instruments which have to be cemented into a borehole (Sacks-Evertson dilatometers, tensor strainmeters), because the strains in the ground are transferred to the actual sensor through the borehole wall, the casing and the cement. In these cases a very rough calibration can be obtained with the help of Earth tide strains, which are theoretically at least known to the order of magnitude. However, very local heterogeneities may complicate this method (King et al., 1976).



### 3.4 Direction sensitivity



**Figure 3** Relative direction sensitivity of linear strainmeters to apparent longitudinal (left) and apparent transversal (right) elastic waves. The strainmeter is indicated by the thick horizontal bar in the center of diagrams. Note that in both cases the strainmeter response is identical for opposite arrival directions.

Dilatometers naturally have isotropic direction sensitivity, the signal does not depend on the direction of arrival of the seismic wave. Figure 3 shows the directional sensitivity patterns of a horizontal strainmeter for longitudinal (left panel) and transversal (right panel) elastic waves. The thick solid bar in the center of each panel indicates the strainmeter. If  $p$  is the angle between the arrival direction of the wave and the direction  $L$ , these functions are described by  $\cos(p) \cdot \cos(p)$  for longitudinal, by  $\sin(2p)$  for transversal waves. This means, that when looking into a certain arrival direction, the sign of the output does not depend on whether the wave comes from the front or from behind. Near the free horizontal surface, vertical strain is proportional to the areal strain (also direction independent), therefore vertical strainmeters provide less information than horizontal strainmeters.

### 3.5 Local arrays of linear strainmeters

If three (or more) horizontal linear strainmeters with different azimuth are deployed at one site, any horizontal component of the strain tensor associated with the incoming wave can be determined from a particular combination of the calibrated signals. Widmer et al. (1992) use this property to demonstrate, that toroidal free oscillation peaks in shear strain spectra do not exist in the areal strain spectrum, a theoretically required result.

### 3.6 Local phase velocity

Assuming a plane elastic wave is recorded by a linear strainmeter and an inertial seismometer at the same station, then one can in principle derive the phase velocity of the wave at this

location. This is due to the fact that the inertial seismometer's output amplitude is proportional to the second derivative of displacement with time (frequency squared), while the strainmeter output is proportional to the spatial derivative (wavenumber).

Depending on the components one can derive equations which relate the frequency-dependent phase velocity with the amplitude spectra of both instruments (Mikumo and Aki, 1964). However, this attractive method has not found many applications, because of local deviations of the deformation field from a simple plane wave (Sacks et al., 1976; King et al., 1976).

### 3.7 Effects of local heterogeneity

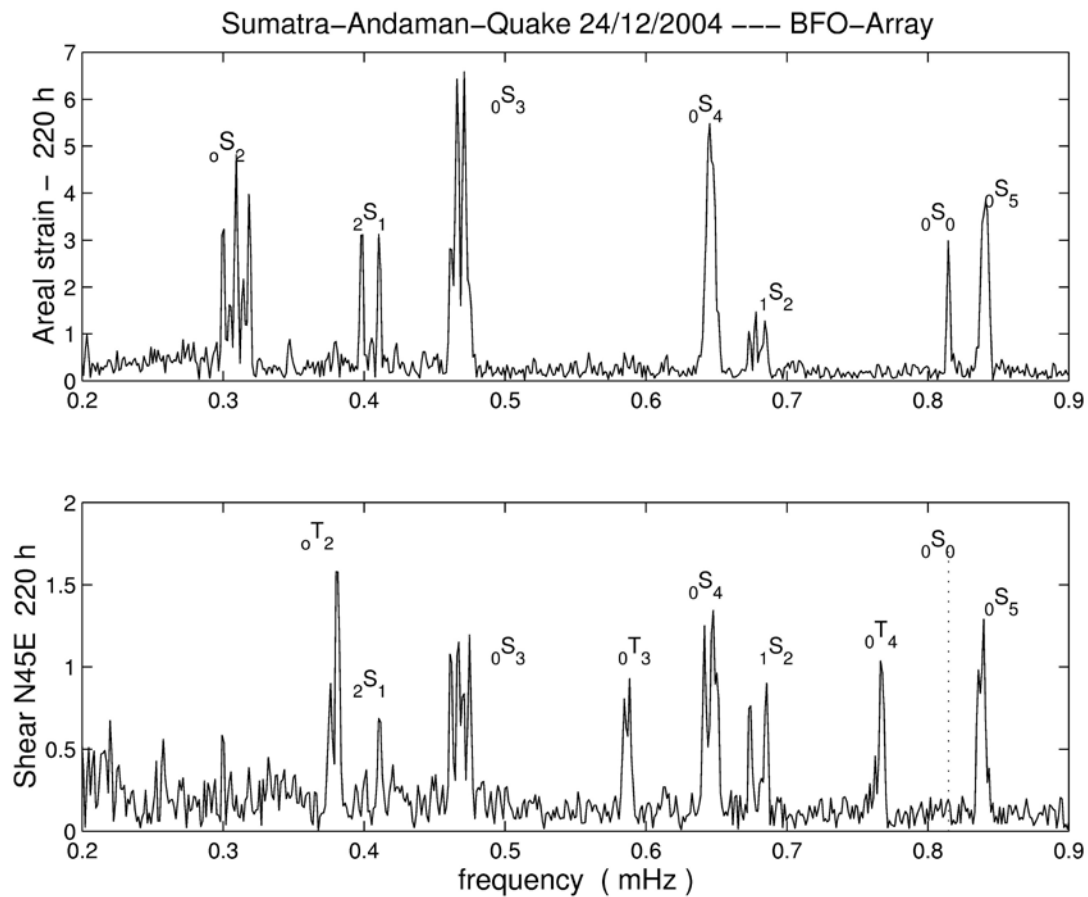
It has been already mentioned twice that the interpretation of results from strainmeters is plagued by the distortion of the strainfield of the arriving seismic waves by local heterogeneities. In tidal research this effect is well known to affect amplitudes and phases and in that field the terms: cavity, topographical and geological effects are used. Arrival times and frequencies are not affected, but the local displacement field could differ appreciably from that of a theoretical plane wave even if the approaching wave was plane (see also Wielandt 1993). The scale of these effects is of the order of magnitude of the signal and with purposeful installation one can obtain apparent mechanical amplification up to a factor of 50. Beavan et al. (1979) have shown this with a 1 m-invar wire strainmeter at BFO for earthquakes and tides. Those local effects can be minimized by installing strainmeters far from any local heterogeneities in the long direction, in an area without topography and in/on large homogeneous rock units. However, even with a lot of care these effects cannot be avoided completely. The ground around the instrument and its heterogeneities must be considered a part of the instrument (largely unknown) if amplitudes, phases and waveforms are interpreted and these unknown properties could be a function of time. King et al. (1976) and Sacks et al. (1976) deal with this problem for seismological applications. Beaumont and Berger (1974) suggest using the geological effect on tides for earthquake prediction (see below). For inertial seismometers these effects play a role only at periods much longer than ten seconds, because at higher frequencies the inertial effect (proportional to squared frequency) overwhelms the other contributions to the extent that they are negligible (King et al. 1976). One example, where the locally produced tilts were needed to explain the observations with broadband inertial seismometers was encountered in the near-field of explosions at Stromboli, Italy by Wielandt and Forbriger (1999). Strainseismometers are subject to these effects at all frequencies because they measure in fact differences in the displacement fields. The longer the baseline length  $L$  of a strainmeter, the more one can hope that local effects are averaged out, at least small scale effects. Gombert and Agnew (1996) discuss some results from PFO in this context.

## 4 Some results

The following list is not meant to be comprehensive. It should simply present the spectrum of research possibilities involving strainmeters.

- Strainmeters were successful in recording the Earth's free vibrations and long-period surface waves from the beginning. One famous example is the record of the Isabella, California, quartz-rod strainmeter of the Great Chilean quake 1960 (Ben-Menahem and Singh, 1981, Fig. 5.29). Widmer et al. (1992) and Zürn et al. (2000) show shear

strain spectra from 10 m Invar wire strainmeters at BFO where the fundamental toroidal mode of the Earth,  ${}_0T_2$  with a period of 44 minutes, stands clearly above the noise floor for the Macquarie 1989 and Balleny Island 1998 events, respectively. The disastrous N-Sumatra–Andaman Islands earthquake on December 26, 2004, excited the seismic free oscillations of the Earth to the largest amplitudes since 1964. Several strainmeters were recording those with unprecedented quality. Park et al. (2008) analyzed spectra from the records of two orthogonal laser extensometers in the Gran Sasso observatory, Italy with very clear observation of the lowest degree fundamental toroidal modes. The strainmeter array at BFO also provided very high quality records. Fig. 4 shows spectra of areal and shear strains computed from the records (Widmer et al. 1992) of three linear strainmeters with the fundamental spheroidal and toroidal modes. The “breathing mode”  ${}_0S_0$  stands out clearly in the spectrum of areal strain, its initial amplitude was estimated to  $8.8 \times 10^{-12}$ .



**Figure 4** Spectra of areal (top) and shear (bottom) strains calculated from the records of the three linear strainmeters at BFO. Toroidal modes are missing in the areal strain and enhanced with respect to spheroidal modes in the shear strains. Many of the multiplets show splitting due to symmetry breaking by rotation and ellipticity. The “breathing mode”  ${}_0S_0$  is missing in the shear strain, its frequency is indicated by the dotted line. A barometer correction was applied to the linear strain data.

- Coseismic steps consistent with source theory were repeatedly observed with the laser strainmeters at PFO for earthquakes in California (Wyatt, 1988; Agnew and Wyatt, 1989).
- Very clear postseismic strain signals lasting many days were recorded at PFO, California by the laser strainmeters, the borehole tensor strainmeter (Gladwin, 1984) and several tiltmeters (Wyatt et al. 1994) for the 1992 Landers earthquake sequence. The authors conclude that possibly different processes contribute to the observed signals and discuss those.
- Linde et al. (1993) were able to derive a detailed picture of the mechanism of an eruption of Hekla volcano, Iceland from the records of several Sacks-Evertson borehole dilatometers installed between 14 and 45 km away from the summit.
- Slow and silent earthquakes have repeatedly been reported from records by borehole dilatometers in California and Japan (e. g. Linde et al. 1996).
- Earth tides are continuously probing the Earth with periods of 12 and 24 hours. They can be used to study the response of the rocks. Agnew (1981) tried to find out about nonlinear behaviour of the rocks using data from the laser strainmeters at PFO. He concludes, that in the absence of evidence for nonlinearity from the tides, seismologists are justified in treating the Earth as a linear system. This kind of study is limited by the strain effects of nonlinear ocean tides.
- Beaumont and Berger (1974) suggested from experiments with Finite-Element models, that the earthquake preparation process should modify rock properties near the fault (i.e., by dilatancy) and thus the amplitudes of the tidal strains observed near the fault. Several groups made attempts to make such observations with no success up to date: Linde et al. (1992) looked at borehole dilatometer and tensor strainmeter records in the vicinity of the Loma Prieta, California, quake in 1992 and Omura et al. (2001) investigated super-invar bar-strainmeter data around the 1995 Kobe earthquake from a mine at a distance of 25 km from the epicenter (see also Westerhaus and Zschau, 2001, for a short summary of other attempts) . Latynina and Rizaeva (1976) report tidal strain amplitude variations observed with quartz rod strainmeters before an earthquake, but are not certain about the significance of this result.
- Secular crustal deformation rates have been always a major observation goal for strainmeters. Basically they are able to see this signal, but because of the high and non-stationary noise at ultra-low frequencies, the interpretation in this spectral band is extremely difficult. The work with the very long laser strainmeters at Pinion Flat (PFO), in combination with other instruments and methods, is the most careful one (see articles by Agnew, Wyatt and colleagues) ever performed in this direction. PFO is located between the San Andreas and San Jacinto faults in Southern California and only 10 to 15 km away from both.

## 5 Strain- vs. inertial seismometers

In practice inertial seismometers by far outnumber strainmeters. It is also a fact that experimental seismological research is based mainly on the records from inertial seismometers with very few contributions from the few strainmeters. More than 100 borehole tensor strainmeters (type: Gladwin, 1984) and long-baseline laser strainmeters will be installed (starting 2004) at the Plate Boundary Observatory in the Western United States in order to detect strain signals close to the active faults. Although this presents a large boost in the total number of strainmeters it will not yet change the general imbalance in the use of strainmeters as compared to inertial seismometers.

There are several reasons for this imbalance:

- Inertial seismometers, short-period and broadband, are commercially available, while highly sensitive strainmeters are not. The costs to produce a competitive laser strainmeter are very high, but a Cambridge type wire strainmeter can be produced very cheaply, compared to the cost of a modern broadband seismometer.
- Short-period and most broadband seismometers are very easy to set up. STS-1 seismometers need more care if highest quality is requested. Strainseismometers require much more work for their installation. Borehole seismometer and borehole strainmeter installation probably is comparable.
- By far the most seismological routine work, especially at the regional scale with local networks, is performed analyzing body waves with periods of a few seconds to frequencies of several tens of Hz. At these frequencies most strainmeters, due to their relatively large dimension, suffer from parasitic resonances of some kind depending on the individual design. Possible exceptions are the Sacks-Evertson dilatometer and the borehole tensor strainmeters because they are more compact.
- Strainmeters, in contrast to short-period seismometers, are extremely noisy if installed near the surface of the Earth due to environmental variations of temperature and air pressure and their effects on the instrument itself and the ground around it. Therefore high quality can only be obtained in boreholes, mines and tunnels or by anchoring them to points at depth. This leads to added installation costs, especially if boreholes have to be drilled for the installation. Basically the cost of the borehole has to be added to the cost of the instrument.

It is noted here that the users of global digital broadband data know the differences in quality between vertical and horizontal components as a function of the depth of installation. At long periods the horizontals are very sensitive to tilts (see 5.3.3). Both tilt and strain are local spatial derivatives of the displacement field and show similar local effects in terms of noise and distortions. Therefore the fairest comparison for strainmeters would be to the long-period horizontal inertial seismometers.

For a given input wave amplitude, the amplitude of the output signal is proportional to  $1/\lambda = f/c$  for a strainmeter and  $\sim f^2$  for an inertial seismometer (with  $\lambda$  - wavelength,  $f$  - frequency,  $c$  - phase velocity). Accordingly, when considering waves with equal  $c$ , strainmeters have more and more advantage the lower the frequency gets (for both types of sensors the noise power (see Figure 2) rises strongly with decreasing frequency). Most of the research cited above belongs to "zero-frequency seismology". Low-frequency research work makes sense especially in the near-fields of earthquake faults and active volcanoes (creep events, slow and

silent earthquakes, pre-, co- and postseismic strain transients, de- and inflation periods, etc.) . However, it is prudent not to rely on a single instrument because noise at very long periods is non-stationary and any changes in the coupling of the instrument to the ground or in the materials of the instrument itself will appear as a signal.

## References

- Agnew, D. C. (1981). Nonlinearity in rock: evidence from Earth tides. *J. Geophys. Res.*, **86**, B5, 3969 - 3978.
- Agnew, D. C. (1986). Strainmeters and tiltmeters. *Rev. Geophys.*, **24**, 579 - 624.
- Agnew, D. C., Wyatt, F. K. (1989). The 1987 superstition hills earthquake sequence: strains and tilts at Pinion Flat Observatory. *Bull. Seism. Soc. Am.*, **79**, 2, 480 - 492.
- Beaumont, C., and Berger, J. (1974). Earthquake prediction: modification of Earth tide tilts and strains by dilatancy. *Geophys. J.R. astr. Soc.*, **39**, 111 - 118.
- Beavan, J., Bilham, R., Emter, D., King, G. C. P. (1979). Observation of strain enhancement across a fissure. *Dt. Geod. Komm., Reihe B*, **231**, 47 - 58.
- Ben-Menahem, A., and Singh, S. J. (1981). Seismic waves and sources. *Springer-Verlag*, New York, Heidelberg, Berlin, 1108 pp.
- Evans, K., and Wyatt, F. (1984). Water table effects on the measurement of Earth strain. *Tectonophysics*, **108**, 323 - 337
- Fix, J. E., and Sherwin, J. R. (1970). A high-sensitivity strain-inertial seismograph installation. *Bull. Seism. Soc. Am.*, **60**, 1803 - 1822.
- Gladwin, M. T. (1984). High precision multi-component borehole deformation monitoring. *Rev. Sci. Instrum.*, **55**, 2011 - 2016.
- Gomberg, J., and Agnew, D. C. (1996). The Accuracy of Seismic Estimates of Dynamic Strains: An Evaluation Using Strainmeter and Seismometer Data from Pinion Flat Observatory, California. *Bull. Seism. Soc. Am.*, **86**, 212 - 220.
- Goult, N. R., King, G. C. P., and Wallard, A. J. (1974). Iodine stabilized laser strainmeter. *Geophys. J. R. astr. Soc.*, **39**, 269 - 282.
- King, G. C. P., and Bilham, R. G. (1976). A geophysical wire strainmeter. *Bull. Seism. Soc. Am.*, **66**, 2039-2047.
- King, G. C. P., Zürn, W., Evans, R., and Emter, D. (1976). Site corrections for long period seismometers, tiltmeters and strainmeters. *Geophys. J. R. astr. Soc.*, **44**, 405 - 411.
- Kümpel, H.-J. (1992). About the potential of wells to reflect stress variations within the inhomogeneous crust. *Tectonophysics*, **211**, 317 - 336.
- Latynina, L. A., and Rizaeva, S. D. (1976). On tidal-strain variations before earthquakes. *Tectonophysics*, **31**, 121 - 127.
- Levine, J., and Hall, J. L. (1972). Design and operation of a methane absorption stabilized laser strainmeter. *J. Geophys. Res.*, **77**, 2595 - 2609.
- Linde, A. T., Gladwin, M. T., and Johnston, M. J. S. (1992). The Loma Prieta Earthquake, 1989 and the Earth strain tidal amplitudes: an unsuccessful search for associated changes. *Geophys. Res. Lett.*, **19**, 317 - 320.
- Linde, A. T., Agustsson, K., Sacks, I. S., and Stefansson, R. (1993). Mechanism of the 1991 eruption of Hekla from continuous borehole strain monitoring. *Nature*, **365**, 737-740.

- Linde, A. T., Gladwin, M., Johnston, M. J. S., Gwyther, R. L., and Bilham, R. (1996). A slow earthquake sequence on the San Andreas fault. *Nature*, **383**, 65 - 68.
- Lomnitz, C. (1997). Frequency response of a strainmeter. *Bull. Seism. Soc. Am.*, **87**, 1078 - 1080.
- Mikumo, T., and Aki, K. (1964). Determination of local phase velocity by intercomparison of seismograms from strain and pendulum instruments. *J. Geophys. Res.*, **69**, 721-731.
- Milne, J. (1888). The relative motion of neighbouring points of ground. *Trans. Seism. Soc. Jap.*, **12**, 63 - 66.
- Omura, M., Otsuka, S., Fujimori, K., and Yamamoto, T. (2001). Tidal strains observed at a station crossing the Otsuki Fault, Kobe, Japan, before the 1995 Hyogoken-Nanbu earthquake. *J. Geodetic Soc. Japan*, **47**, 441 - 447.
- Park, J., Amoroso, A., Crescentini, L., and Boschi, E. (2008). Long-period toroidal earth free oscillations from the great Sumatra-Andaman earthquake observed by paired laser extensometers in Gran Sasso, Italy. *Geophys. J. Int.* **173**, 887-905.
- Sacks, I. S., Suyehiro, S., Evertson, D. W., and Yamagishi, Y. (1971). Sacks-Evertson strainmeter, its installation in Japan and some preliminary results concerning strain steps. *Papers in meteorology and geophysics*, **22**, 195 - 208.
- Sacks, I. S., Snoke, J. A., Evans, R., King, G., and Beavan, J. (1976). Single-site phase velocity measurement. *Geophys. J. R. astr. Soc.*, **46**, 253 - 258.
- Westerhaus, M., and Zschau, J. (2001). No clear evidence for temporal variations of tidal tilt prior to the 1999 Izmit and Düzce earthquakes in NW-Anatolia. *J. Geodetic Soc. Japan*, **47**, 448 - 455.
- Wielandt, E. (1993). Propagation and structural interpretation of non-plane waves. *Geophys. J. Int.*, **113**, 45 - 53.
- Wielandt, E., and Forbriger, T. (1999). Near-field seismic displacement and tilt associated with the explosive activity at Stromboli. *Annali Geofisica*, **42**, 407 - 416.
- Widmer, R., Zürn, W., and Masters, G. (1992). Observation of low-order toroidal modes from the 1989 Macquarie Rise event. *Geophys. J. Int.*, **111**, 226 - 236.
- Wyatt, F. (1988). Measurements of coseismic deformation in Southern California. *J. Geophys. Res.*, **93**, 7923 - 7942.
- Wyatt, F., Beckstrom, K., and Berger, J. (1982). The optical anchor - a geophysical strainmeter. *Bull. Seism. Soc. Am.*, **72**, 1707 - 1716.
- Wyatt, F., Agnew, D. C., and Gladwin, M. (1994). Continuous measurements of crustal deformation for the 1992 Landers Earthquake Sequence. *Bull. Seism. Soc. Am.*, **84**, 768 - 779.
- Zadro, M., and Braitenberg, C. (1999). Measurements and interpretations of tilt-strain gauges in seismically active areas. *Earth Science Reviews*, **47**, 151 - 187.
- Zürn, W., Laske, G., Widmer-Schmidrig, R., and Gilbert, F. (2000). Observation of Coriolis coupled modes below 1 mHz. *Geophys. J. Int.*, **143**, 113 - 118.

| Topic   | Constructing response curves: Introduction to the BODE-diagram                                                                                                                                             |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Jens Bribach, GFZ German Research Centre for Geosciences, Dept. 2: Physics of the Earth, Telegrafenberg, D-14473 Potsdam, Germany;<br>E-mail: <a href="mailto:brib@gfz-potsdam.de">brib@gfz-potsdam.de</a> |
| Version | May 2001; DOI: 10.2312/GFZ.NMSOP-2_IS_5.2                                                                                                                                                                  |

## 1 The BODE diagram

*True ground motion* is of major interest in seismology. On the other hand any measuring device will alter the incoming signal as well as any amplifier and any output device. Working in the frequency domain, the quotient of input signal and output signal is called the *response*. This response is complex, and to get a more meaningful result it can be split into two terms: amplitude response and phase response (see 5.2.3). Amplitude response means the output amplitude divided by the input amplitude at a given frequency. Phase response is the difference between output phase and input phase, or the phase shift. A graphical expression of this splitting is known as the BODE-diagram. One part shows the logarithm of amplitude  $A$  versus the logarithm of frequency  $f$  (or angular frequency  $\omega$ , or period  $T$ , see Figure 1a). The other part depicts the (linear) phase  $\phi$  versus the logarithm of frequency  $f$  (or  $\omega = 2\pi f$ , or  $T = 1/f$ , see Figure 1b). For  $A$  the terms *amplification* or *magnification* are also used.

### 1.1 The signal chain

The signal passes a chain of devices. Any single element of this chain can be described by its response. It is useful to split any response into elements of first or second order. At the end the overall amplitude response of the complete chain can be constructed by multiplying all single amplitude responses, and the overall phase response by adding all single phase shifts.

### 1.2 First and second order elements

For the amplitude response the double logarithmic scale of the amplitude diagram facilitates an easy and fast construction. Any element can be approximated by two straight lines. One horizontal line leads to the element corner frequency, and one line drops from that point with a slope depending on the order of the element.

A first order element is completely described by its amplification  $A$  and corner frequency  $f_c$ . The slope beyond  $f_c$  is one decade in amplitude per decade in frequency. The real amplitude value at  $f_c$  is dropped to 0.707 of the maximum amplitude (see Figure 2, full line). However, for our fast construction, we consider only a linear approximation to it (dash-dot lines).

A second order element exhibits a slope of two decades in amplitude per decade in frequency. Additionally it needs another parameter called damping  $D$ , describing the amplitude behaviour at frequencies near  $f_c$  (compare Figure 3).



## 2 The seismological signal chain

The seismological signal passes the chain

- mechanical receiver
- transducer
- preamplifier
- filter
- recording unit (to be recognized separately)

**Note!** Below we discuss in sections 2.1 and 2.2 amplitude responses related to ground displacement. Therefore, the ordinate axis of the BODE-diagram (amplitude  $A$ ) for the mechanical receiver has no unit (or the unit  $[m/m]$ ), and for the transducer the unit is  $[V/m]$ . Changes to other types of movement, being proportional to ground velocity or to ground acceleration, will be described in section 3.

### 2.1 The mechanical receiver

The mechanical receiver is a second order system. It describes the relative movement of the pendulum (i.e., a seismic mass attached to a frame by a spring) with respect to the frame. Damping  $D$  is set mostly to 0.707. Only at this damping value is the amplitude value at  $f_c$  also 0.707 (Figure 3, full curve). For higher values of damping one obtains a more flat curve (dashed). For lower values of  $D$  the dash-dot curve strongly exceeds the amplification level at  $f_c$ , indicating low-damped resonance oscillations of the pendulum, which can be stimulated by any signal.

The amplification of the mechanical receiver is  $A = 1$ . This means that for frequencies  $f > f_c$  the amplitude of the pendulum movement with respect to the frame is similar to the ground amplitude. For the phase shift see Figure 7b (HIGH Pass 2).

### 2.2 The transducer

The transducer transforms the relative movement of the pendulum into an electrical signal, i.e., in a voltage. The transducer constant  $G$  gives the value of the output voltage  $U$  depending on the relative pendulum movement  $z$ . There are three main types of transducers, distinguished by their proportionality to ground motion and its derivatives:

- |                |                    |               |                                             |
|----------------|--------------------|---------------|---------------------------------------------|
| - Displacement | $U \sim z$         | $G_d[V/m]$    | (capacitance or inductance bridges)         |
| - Velocity     | $U \sim dz/dt$     | $G_v[Vs/m]$   | (magnet-coil systems)                       |
| - Acceleration | $U \sim d^2z/dt^2$ | $G_a[Vs^2/m]$ | (piezo-electric systems, $U \sim F = m a$ ) |

The above proportionality of the transducer voltage to ground motion (i.e., to displacement, velocity or acceleration, respectively) is, of course, only given for frequencies  $f > f_c$ , i.e., for the horizontal part of the mechanical receiver response (see Figure 3).

All transducer amplitude responses can be drawn as straight lines over the full considered frequency range (Figure 4). They differ only in their slope.

The phase responses have a constant phase shift over the whole frequency range with values of  $0^\circ$  (displacement),  $90^\circ$  (velocity), or  $180^\circ$  (acceleration).

### 2.3 The preamplifier

The preamplifier is a first order LOW Pass. Its corner frequency is beyond the signal range of seismology - up to several 10 kHz. Thus, only the amplification is of interest (Figure 5). The response is a horizontal line drawn at the amplification level  $A$ .

The phase shift is  $\phi = 0^\circ$ , but one should keep in mind that, if using the inverted input, the phase shift will be  $\phi = -180^\circ$  over the whole frequency range.

### 2.4 First and second order LOW Passes

LOW Passes have constant amplifications  $A$  for frequencies lower than their corner frequencies  $f_c$ . For frequencies higher than  $f_c$  the amplification drops with a slope depending on the order of the filter (Figure 6a). LOW Passes cut the high frequencies, therefore, also the term High Cut is used. The phase shift for  $f < f_c$  is about  $0^\circ$  and for  $f > f_c$  it turns to  $-90^\circ$  (first order, LP1) or  $-180^\circ$  (second order, LP2; see Figure 6b), passing half of the phase shift exactly at  $f_c$ .

Of course, the given amplitude and phase values are approximations. In reality we would obtain  $\phi = 0^\circ$  only if inserting a frequency of 0 Hz, and  $\phi = -90^\circ$  ( $-180^\circ$ ) for infinite frequency values. However, the accuracy is sufficient for our fast construction.

### 2.5 First and second order HIGH Passes

HIGH Passes have constant amplifications  $A$  for all frequencies higher than their corner frequency  $f_c$ . For frequencies lower than  $f_c$  the amplification drops with a slope depending on their order (Figure 7a). They cut the low frequencies, so one can also find the term Low Cut.

The phase shift for  $f > f_c$  is about  $0^\circ$ , and for  $f < f_c$  it turns to  $+90^\circ$  (first order, HP1) or  $+180^\circ$  (second order, HP2; see Figure 7b), passing half of the phase shift exactly at  $f_c$ .

Comparable to the description of LOW Passes the given amplitude and phase values are approximations.

### 2.6 Second order BAND Pass

The second order BAND Pass (BP2) can be explained as a combination of a first order LOW Pass and a first order HIGH Pass. It suppresses all frequencies, except  $f_c$ , with a slope of one decade in amplitude per decade in frequency (Figure 8a). The peak at  $f_c$  can be turned into a horizontal line (symmetrical to  $f_c$ ) by increasing the damping to values  $D > 1$ . Thus it is possible to construct a BAND Pass by combining a HIGH Pass with a LOW Pass.

The phase shift for  $f < f_c$  is about  $+90^\circ$ , and for  $f > f_c$  it turns to  $-90^\circ$  (see Figure 8b), passing half of the phase shift at  $f_c$ .

### 3 The overall response

The construction of the overall response should be divided into two steps:

- from mechanical receiver to the final filter stage; and
- adding the recorder response.

The first result, the electrical output, is useful for fitting the signal to the recorder input. It has to be fixed, meaning that changes in magnification (or signal resolution) should be done by setting up the recorder only.

#### 3.1 From the mechanical receiver to the final filter

As defined in section 2, the amplitude response is constructed related to ground displacement. Multiplying all the units of our signal chain, we get the unit  $[V/m]$  for the ordinate axis. All elements, including mechanical receiver, transducer, and filter stages can be implemented in the same sheet with a double logarithmic grid, each element with its magnification and its corner frequency. Then the resulting amplitude response has to be constructed point by point at certain frequencies. This can be done either by multiplying the amplitudes of all elements at these frequencies, which is the more secure method, or, alternatively, by adding the distances (e.g., in millimetres) of all element amplitudes to the amplitude level line  $A = 1$ , with positive distances if above this line and negative ones if below. This method is faster. A linear addition is, in this logarithmic scale, a multiplication of the amplitude values. The final amplitude response curve can be drawn on the same sheet, together with the single elements.

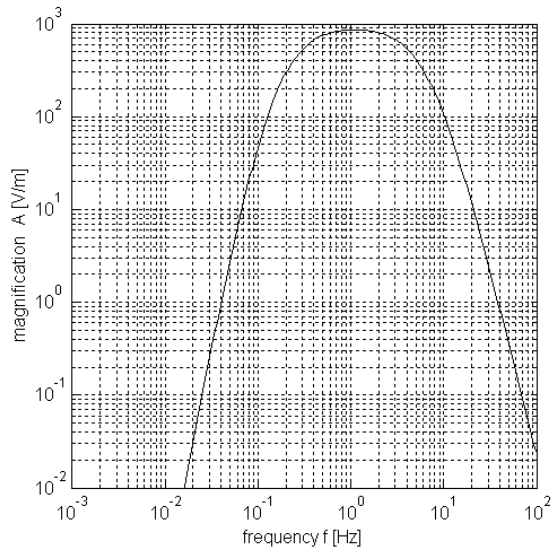
#### 3.2 Adding the recorder

In reality, at the end of our signal chain we will find a commercially available recorder, transforming the obtained voltage back into movement (drum recorder) or into computable digital values (Analogue-to-Digital Converter = ADC). Its main parameter is the input sensitivity  $H$ . In the case of a drum recorder,  $H$  is the pen deflection per Volt (in units  $[m/V]$ ). For an ADC,  $H$  is the digital count per Volt (in units  $[digit/V]$ ).

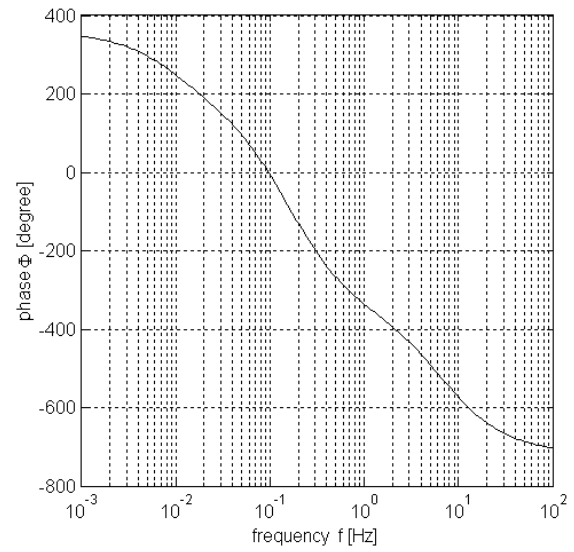
Thus the overall amplitude response needs a separate BODE-diagram for each recorder type. Multiplying the units we obtain the units  $[m/m]$  for the drum recorder, and  $[digit/m]$  for the ADC. You will also find derivatives of this unit, like  $[digit/nm]$  or  $[counts/nm]$ .

#### 3.3 Introducing ground velocity and ground acceleration

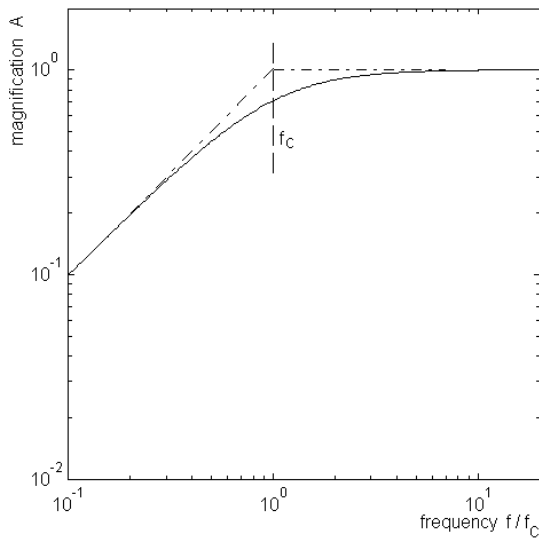
If the amplitude response curve has to be constructed related to ground velocity (or ground acceleration), it is sufficient to redraw either the response of the mechanical receiver or the transducer. The simpler method is to change the transducer response. Each slope will change by one order if going from displacement to velocity, or from velocity to acceleration. The unit of the ordinate changes from  $[V/m]$  (displacement) via  $[Vs/m]$  (velocity) to  $[Vs^2/m]$  (acceleration). These units will also be the units of the amplitude response from the mechanical receiver to the final filter stage. Beyond this the construction of the overall amplitude response is similar to section 3.1. The units of the recorder amplitude response will alter to  $[m\cdot s/m]$  (velocity) or  $[m\cdot s^2/m]$  (acceleration) for the drum recorder. For the ADC we obtain  $[digit\cdot s/m]$  (velocity) or  $[digit\cdot s^2/m]$  (acceleration).



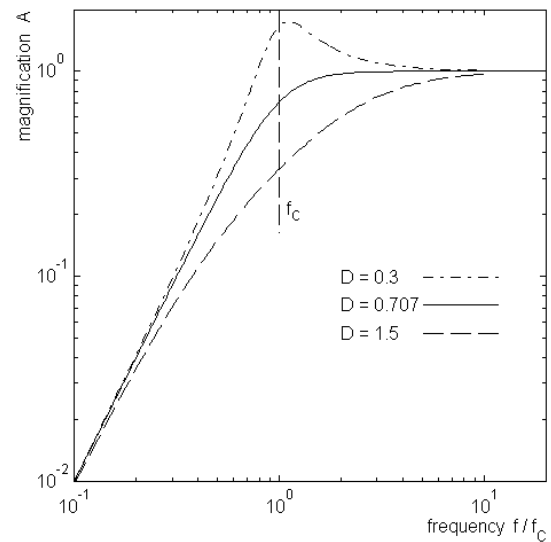
**Figure 1a** Amplitude Response.



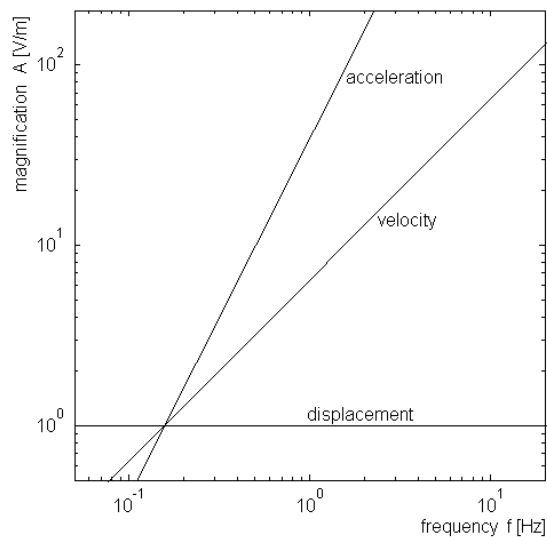
**Figure 1b** Phase Response.



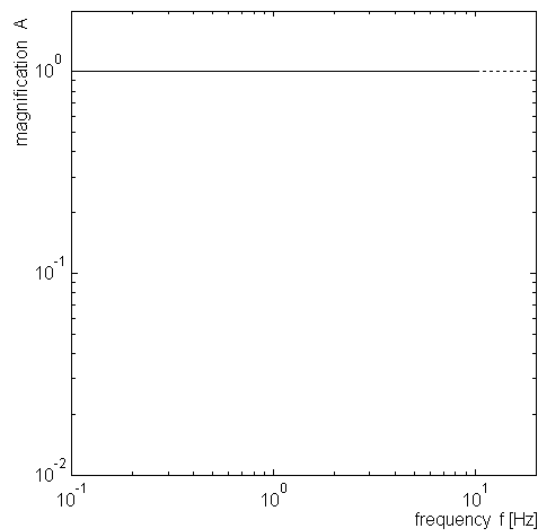
**Figure 2** First Order HIGH Pass (HP1).



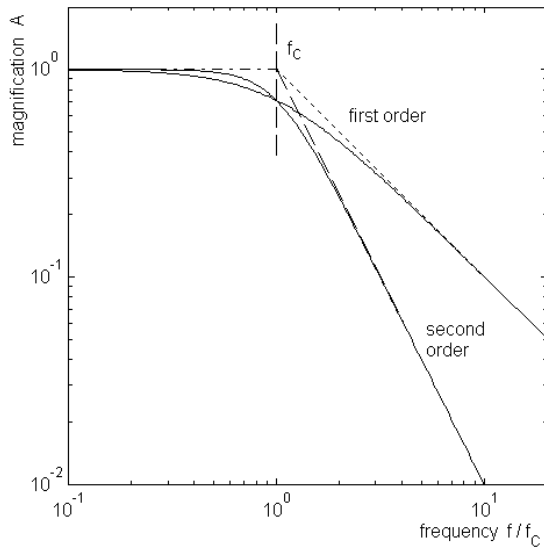
**Figure 3** Second Order HIGH Pass (HP2) or Mechanical Receiver.



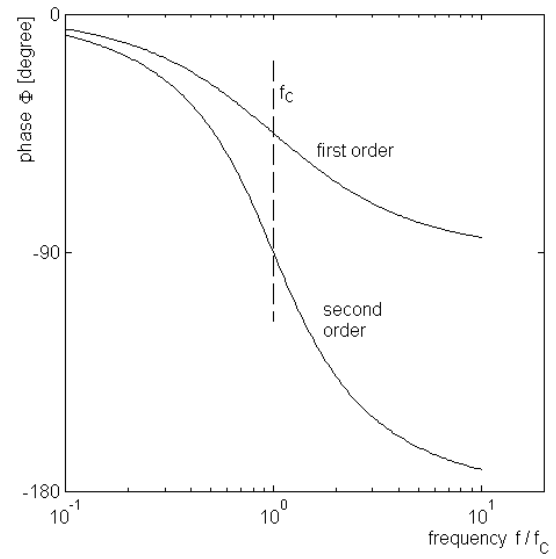
**Figure 4** Transducer Amplitude Response.



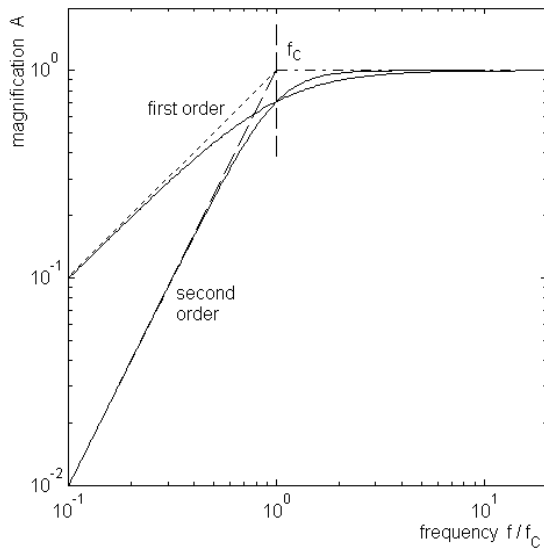
**Figure 5** Preamplifier Amplitude Response.



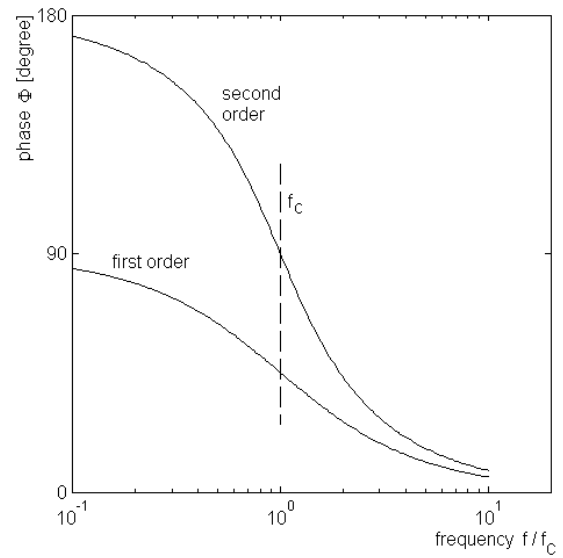
**Figure 6a** LOW Pass Amplitude Response.



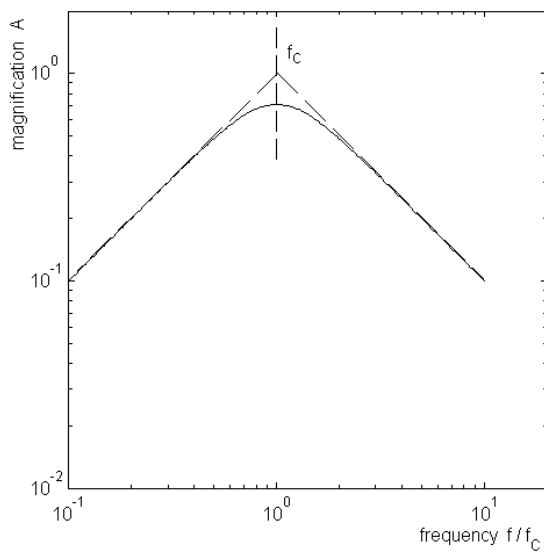
**Figure 6b** LOW Pass Phase Response.



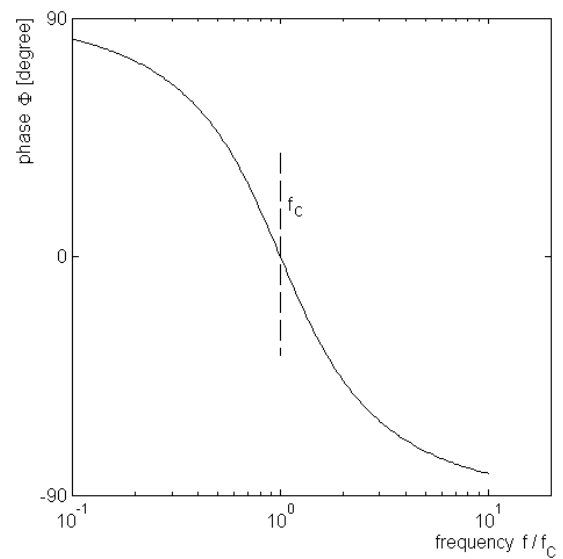
**Figure 7a** HIGH Pass Amplitude Response.



**Figure 7b** HIGH Pass Phase Response.



**Figure 8a** BAND Pass Amplitude Response.



**Figure 8b** BAND Pass Phase Response.

| Topic   | Measuring rotational ground motions in seismological practice                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|---------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p><b>William. H. K. Lee<sup>1)</sup>, John R. Evans<sup>1)</sup>, Bor-Shouh Huang<sup>2)</sup>, Charles R. Hutt<sup>3)</sup>, Chin-Jen Lin<sup>2)</sup>, Chun-Chi Liu<sup>2)</sup>, and Robert L. Nigbor<sup>4)</sup></b></p> <p><sup>1)</sup> U. S. Geological Survey, MS 977, 345 Middlefield Road, Menlo Park, CA 94025, USA; contact via E-mail: <a href="mailto:lee@usgs.gov">lee@usgs.gov</a>;</p> <p><sup>2)</sup> Institute of Earth Sciences, Academia Sinica, P.O. Box 1-55, Nankang, 11529 Taipei, Taiwan;</p> <p><sup>3)</sup> U. S. Geological Survey, P.O. Box 82010, Albuquerque, NM 87198, USA;</p> <p><sup>4)</sup> Department of Civil Engineering, University of California at Los Angeles, Los Angeles, CA 90095, USA.</p> |
| Version | February 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_5.3                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |

|                                                                               | page |
|-------------------------------------------------------------------------------|------|
| <b>1 Introduction</b>                                                         | 1    |
| <b>2 Early attempts to study rotational motions</b>                           | 3    |
| <b>3 Techniques for measuring rotational motions</b>                          | 4    |
| 3.1. Theoretical basis                                                        | 4    |
| 3.2 Measuring considerations                                                  | 5    |
| <b>4. Measuring rotational motions in practice</b>                            | 6    |
| 4.1 Ring laser gyros                                                          | 6    |
| 4.1.1 Ring laser gyro technology                                              | 6    |
| 4.1.2 G Ring laser and recording teleseisms                                   | 7    |
| 4.2 Strong-motion inertial and angular sensors                                | 9    |
| 4.2.1 The eentec <sup>TM</sup> Model R-1 <sup>TM</sup> rotational seismometer | 10   |
| 4.2.2 Laboratory calibration of rotational sensors                            | 12   |
| 4.2.3 Field deployment of rotational sensors                                  | 14   |
| <b>5 Discussion of results and benefits of studying rotational motions</b>    | 14   |
| 5.1 Linear and non-linear elasticity                                          | 16   |
| 5.2 Near-field seismology                                                     | 16   |
| 5.3 Using explosions to study rotational motions                              | 18   |
| 5.4 Processing collocated measurements of translations and rotations          | 19   |
| <b>6 Conclusions</b>                                                          | 20   |
| <b>Acknowledgments</b>                                                        | 21   |
| <b>References</b>                                                             | 21   |

---

*Hypertext links and other references to products are provided for information only and do not constitute endorsement or warranty, expressed or implied, as to their suitability, content , usefulness, functioning, completeness, or accuracy by the United States Geological Survey.*

# 1 Introduction

Rotational seismology is an emerging study of all aspects of rotational motions induced by earthquakes, explosions, and ambient vibrations. The subject is of interest to several disciplines, including seismology, earthquake engineering, geodesy, and earth-based detection of Einstein's gravitation waves. In this chapter, we will present basic elements of measuring rotational ground motions in seismological practice and an overview of recent observations. Because there are terms in this Chapter that may not be familiar to the readers, readers may consult either the amended **Glossary** of this manual, or for an even more detailed glossary on rotational seismology Lee (2009).

Likely rotational effects of earthquake waves together with rotations caused by soil-structure interaction have been observed for centuries (e.g., rotated chimneys, monuments, and tombstones relative to their supports). A summary of historical examples of observations on earthquake rotational effects is provided by Kozák (2009), including reproduction of the relevant sections from Mallet (1862) and Reid (1910). Figure 1 shows the rotation of the monument to George Inglis (erected in 1850 at Chatak, India) as observed by Oldham (1899) after the 1897 Great Shillong earthquake.



**Figure 1** Rotation of the monument to George Inglis (erected in 1850 at Chatak, India) as observed by Oldham (1899) after the 1897 Great Shillong earthquake. This monument had the form of an obelisk rising about 20 m high from a base 4 m on each side. During the earthquake, the topmost 2 m section was broken off and fell to the south and the next 3 m section was thrown to the east. The remnant is about 6.5 m in height and is rotated about 15° relative to the base. Such rotations can be due to strictly translational accelerations and don't necessarily require rotational ground motion (see text).

A few early authors proposed rotational waves or at least some “vortical” motions. Many different terms were used for the rotational motion components at this early development stage. For example, “rocking” is rotation around a horizontal axis, sometimes also referred to as “tilt”. Mallet (1862) proposed that rotations of a body on the Earth's surface are due to a sequence of different seismic phases emerging at different angles. Computational modelling by Hinzen (2010) also demonstrates such effects.

Reid (1910) studied this phenomenon, which was observed in the 1906 San Francisco earthquake, and pointed out that the observed rotations are too large to be produced by waves of linear elastic distortion. Such waves “*produce very small rotations, whose maximum amount, ... is given by the expression  $2\pi A = \lambda$ , where  $A$  is the amplitude and  $\lambda$  the wavelength;*

with a wave as short as 10,000 feet (3 km) and an amplitude as large as 0.2 of a foot (6 cm), the maximum rotation would only be about 0.25 of a minute of arc [ $7.3 \mu\text{rad}$ ], a quantity far too small to be noticeable.” (Reid, 1910, p. 44). A modern analysis of such rotational effects is presented in Todorovska and Trifunac (1990).

Observational seismology is based mainly on measuring *translational* motions because of a widespread belief that *rotational* motions are insignificant. For example, Richter (1958, footnote on p. 213) states that “*Theory indicates, and observation confirms, that such rotations are negligible.*” Richter provided no references (but perhaps after Reid, 1910, p. 44), and there were no instruments at that time sensitive enough to either confirm or refute this claim. Recent advances in rotational seismology became possible because sensitive rotational sensors have been developed in aeronautical and astronomical instrumentation.

## 2 Early attempts to study rotational motions

Ferrari (2006) summarized two models of an electrical seismograph with sliding smoked paper, developed by P. Filippo Cecchi in 1876 to record three-component *translational* motions and also the *torsional* movements from earthquakes. Although these instruments operated for several years, no rotational motion could be recorded because of low transducer sensitivity. Pioneers in several countries attempted to measure rotational motions induced by earthquakes. Nearly a century ago, Galitzin (1912) suggested using two identical pendulums installed on different sides of the same axis of rotation for separate measurement of rotational and translational motion. This was later implemented, for example, by Kharin and Simonov (1969) in an instrument designed to record strong ground motion. Using an azimuthal array of seismographs, Droste and Teisseyre (1976) derived rotational seismograms for rock bursts in a nearby mine. Inspired by Walter Munk, Farrell (1969) constructed a gyroscopic seismometer, and obtained a static displacement of  $<1$  cm and a tilt of  $<0.5 \mu\text{rad}$  at La Jolla, California, during the Borrego Mountain earthquake of 09 April 1968 (magnitude 6.5) at an epicentral distance of 115 km.

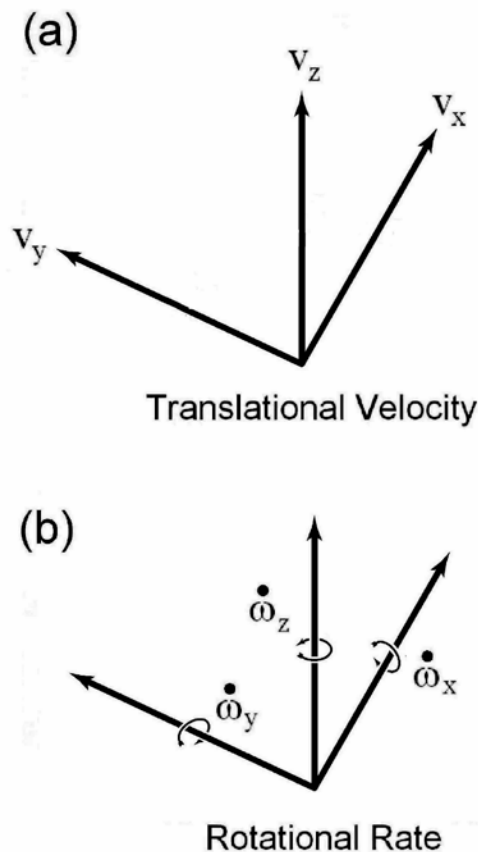
Early efforts also included studies of explosions. For example, Graizer (1991) recorded tilts and translational motions in the near-field of two nuclear explosions using seismological observatory sensors to measure point rotations directly. Nigbor (1994) measured rotational and translational point ground motions directly with a commercial rotational micro-electro-mechanical system (MEMS) sensor and found significant near-field rotational motions ( $660 \mu\text{rad}$  at 1 km distance) from a one-kiloton explosion.

Rotations and strains of the ground and in the response of structures have been deduced indirectly from accelerometer arrays using methods valid for seismic waves having wavelengths that are long compared to the distances between sensors (e.g., Trifunac 1979; 1982; Oliveira and Bolt, 1989; Spudich et al., 1995; Bodin et al., 1997; Huang, 2003; Suryanto et al., 2006; Wassermann et al., 2009). The rotational components of ground motion have also been estimated theoretically, using kinematic source models (Bouchon and Aki, 1982; Wang et al., 2009) and linear elastodynamic theory of wave propagation in elastic solids (Lee and Trifunac, 1985, 1987). Recently, Graizer and Kalkan (2010) presented a history of strong motion seismology and rotations, outlining also future directions.



### 3 Techniques for measuring rotational motions

The general motion of the particles or a small volume in a solid body can be divided into three parts: translation (along the X-, Y-, and Z-axes), rotation (about the X-, Y-, and Z-axes), and strain (six components). Figure 2(a) shows the axes in a Cartesian coordinate system for *translational velocity* measured by the usual sort of seismometers used in seismology, and Figure 2(b) shows the corresponding axes of *rotation rate* measured by rotational sensors (Evans and the International Working Group on Rotational Seismology, 2009). These are “body attached” coordinates, those that a seismic instrument would measure at a given instant as the sensors move and rotate through space. Converting an extended record of these body-fixed motions to recover motions in an Earth-fixed, quasi-inertial coordinate system has been performed for decades in “strapped down” inertial navigation systems such as those attached to a moving airplane. Lin et al. (2010) introduce these equations into seismology and earthquake engineering for recovering inertial-frame ground and structure motions.



**Figure 2** Coordinate system for translational velocity (adopted from Figure 1 of Evans and the International Working Group on Rotational Seismology, 2009). (a) Coordinate system for translational velocity. (b) Coordinate system for body-fixed or instantaneous rotational rate. This is the preferred nomenclature and sign conventions for observed translational and rotational motions in seismology and earthquake engineering. X and Y axes point into the page and are normally in the horizontal plane. Note the positions of rotation gyres with respect to the axes (the line in front breaks the line behind). Axes point in the positive directions of both the translational vectors and rotation vectors.

Annotations  $v$  and  $\dot{\omega}$  are translational velocity (m/s) and rotation rate (rad/s); the subscripts x, y, and z refer to the X, Y, and Z components.

#### 3.1 Theoretical basis

Rotational motion in seismology is usually formulated in the framework of linear elasticity theory with the assumption of *infinitesimal deformation* (Cochard et al., 2006). In a linear elastic medium the displacement  $\mathbf{u}$  of a point  $\mathbf{x}$  is related to a neighboring point  $\mathbf{x} + \delta\mathbf{x}$  by

$$\mathbf{u}(\mathbf{x} + \delta\mathbf{x}) = \mathbf{u}(\mathbf{x}) + \boldsymbol{\varepsilon} \delta\mathbf{x} + \boldsymbol{\omega} \times \delta\mathbf{x} \quad (1)$$

where  $\boldsymbol{\varepsilon}$  is the strain tensor and

$$\boldsymbol{\omega} = \frac{1}{2} \nabla \times \mathbf{u}(\mathbf{x}) \quad (2)$$

is a pseudo-vector representing the infinitesimal angle of rigid rotation generated by the disturbance. The three components of rotation about the X-axis, Y-axis and Z-axis are given by the following equations for such infinitesimal motions:

$$\begin{aligned} \omega_x &= \frac{1}{2} (\partial u_z / \partial y - \partial u_y / \partial z), \\ \omega_y &= \frac{1}{2} (\partial u_x / \partial z - \partial u_z / \partial x), \\ \omega_z &= \frac{1}{2} (\partial u_y / \partial x - \partial u_x / \partial y) \end{aligned} \quad (3)$$

Therefore, rigid rotations can be observed: (i) indirectly, by an array of translational seismometers for “cord” rotations associated with long wave lengths by assuming that contamination of translational signals by rotational motions is small, and that the linear elasticity theory is valid (e.g., Spudich and Fletcher, 2008), or (ii) directly, by rotational sensors for “point” body-fixed rotations (e.g., Lee et al., 2009b, and Lin et al., 2010).

Although classical elasticity theory works well for earthquakes in the *far* field recorded by vaulted seismographs, this linear theory may not be applicable in the *near* field. Theoretical work suggests that in shallow granular, cracked or biologically-disturbed, often dry continua (e.g., soil or weathered rock at or near Earth’s surface), asymmetries of the stress and strain fields can create rotations separate from and larger than those predicted by classical elastodynamic theory (e.g., Teisseyre et al., 2006; 2008; Teisseyre, 2010). If so, these exaggerated rotations join previously known translational “site effects” for strong-motion stations. Further, unlike the traditional fault-slip model (Aki and Richards, 2002, p. 38-59) assuming a fault zone of *zero thickness*, Knopoff and Chen (2009) consider the case of faulting that takes place on a fault zone of *finite thickness*. They show that there is an additional single-couple term in the body-force equivalence and additional terms in the far-field displacement. They also show that the single-couple equivalent does not violate the principles of Newtonian mechanics because the torque imbalance in the single-couple is counterbalanced by rotations within the fault zone, with torque waves being radiated.

### 3.2 Measuring considerations

Most seismological instruments used in measuring translational ground motion are pendulum seismometers and accelerometers, and a tutorial on measuring rotations using multi-pendulum systems is given in Graizer (2009). It is important to note that pendulums are sensitive to translational motion and rotations. Horizontal pendulums are sensitive to the acceleration of linear motion, tilt, angular acceleration, and cross-axis excitations; and vertical pendulums are sensitive to the acceleration of linear motion, angular acceleration, and cross-axis excitations but not to tilt. Graizer (2005, 2010) provided complete equations of pendulum sensors used in seismology and their approximation for strong-motion seismology and also presented a method (Graizer, 2006) to extract residual tilt from uncorrected strong-motion accelerograms.

Measuring rotational ground motions is closely related to the traditional seismological observations of translational ground motions, and measurements of tilts and strains. However, it is beyond the scope of this Chapter to discuss tiltmeters and strainmeters. For these instruments, interested readers may consult, e.g., Agnew (1986), Segall (2010) and IS 5.1 by Zürn (2002) and in this NMSOP2.

## 4 Measuring rotational motions in practice

In the past decade, rotational motions from small local earthquakes to large teleseisms have been recorded successfully by sensitive rotational sensors in several countries (e.g., Takeo, 1998; McLeod et al., 1998; Igel et al., 2005, 2007; Suryanto et al., 2006; Cochard et al., 2006). In particular, the application of Sagnac interferometry in large ring-laser gyros provided greatly improved sensitivity to rotations at teleseismic distances and showed that the measured rotations are a good match to those estimated from linear elastic wave theory. Combined with translations, such rotational motions provide additional observations that lead to new approaches to the seismic inverse problem (Bernauer et al., 2009; Fichtner and Igel, 2009). Recently Igel et al. (2010) reported the first observations of Earth's free oscillations using ring laser recordings, and opened up potential applications of rotational seismology at long periods.

In contrast, strong-motion observations near the source in both Japan and Taiwan show that the amplitudes of these rotations can be one to two orders of magnitude greater than that expected from linear elasticity theory (e.g., Takeo, 1998; Lee et al., 2009b).

### 4.1 Ring laser gyros

An unexpected advance in studying rotational ground motions came from a different field of geophysics. Recent developments of highly sensitive ring laser gyroscopes to monitor the Earth's rotation also yield valuable data on rotational motions from large teleseismic events (Schreiber, 2009). The most important property that makes such rotation sensors useful for seismology is its very low noise floor and high sensitivity to rotational motions and its insensitivity to translational and cross-rotational motions. Very recently, the 3 September 2010 Canterbury, New Zealand earthquake ( $M_w$  7.0) occurred near the Canterbury ring-laser station, which recorded many of the aftershocks (Schreiber, personal communication, 2010).

The rotation rates expected and observed in seismology range from the order of  $10^{-1}$  rad/s (e.g., Nigbor, 1994; Trifunac 2009) for near seismic sources down to order  $10^{-11}$  rad/s for large earthquakes at teleseismic distances (e.g., Igel et al., 2005, 2007). This range spans at least 10 orders of magnitude (200 dB), much as do translational motions, and it is unlikely that one instrument or one instrumental technology will be capable of providing accurate measurements over such a large range of amplitudes. Ring laser technology is currently the most promising approach to recording the small rotational motions induced by teleseisms, but to reach the sensitivities needed for seismic measurements large sizes (several square meters) are necessary and therefore the primary drawback of this technology is its very high cost.

#### 4.1.1 Ring laser technology

Ring lasers detect the Sagnac beat frequency of two counter-propagating laser beams (Stedman, 1997; and Figure 3(c)), and Schreiber (2010) reviews the recent progress in ring laser technology. These active interferometers generally form triangular or square closed loops several meters across. The cavity is accurately evacuated and filled with a mixture of helium and neon for the laser, and the instrument is vacuum tight. If this instrument is rotating on a platform with respect to inertial space, the effective cavity length between the co-rotating and counter-rotating laser cavities differ and one observes frequency splitting

resulting in a beat frequency. This beat frequency  $\delta f$  is directly proportional to the rotation rate  $\Omega$  around the surface normal  $\mathbf{n}$  of the ring laser system, as given by the Sagnac equation:

$$\delta f = \frac{4A}{\lambda P} \mathbf{n} \cdot \Omega, \quad (4)$$

where  $P$  is the perimeter of the instrument,  $A$  its area, and  $\lambda$  the laser wavelength. This equation has three contributions that influence the beat frequency  $\delta f$ : (a) variations in the scale factor ( $4A/\lambda P$ ) have to be avoided by means of a servo system which keep the perimeter of the ring constant; (b) changes in orientation  $\mathbf{n}$  (tilting relative to Earth's rotation axis) enter the beat frequency *via* the inner product; and (c) variations in  $\Omega$  (e.g., changes in Earth's rotation rate and seismically induced rotations). The dominant contribution to  $\delta f$  is  $\Omega$ . Note that translations do not contribute to the Sagnac frequency unless they affect  $P$  or  $A$  in some indirect manner. Next generation ring-laser gyro should have 3 axes attached to the same monument.

Ring lasers are sensitive to rotations only, assuming stable ring geometry and lasing. However, for co-seismic observations at the Earth's surface the horizontal components of rotation (i.e., tilts) will contribute to the vertical component of rotation rate (via  $\mathbf{n}$  in Equation (4)). As shown by Pham et al. (2009), the tilt-coupling effect is several orders of magnitude below the level of the earthquake-induced rotational signal unless one is very close to the source (where sensitive ring lasers would not be the appropriate technology; see next Section).

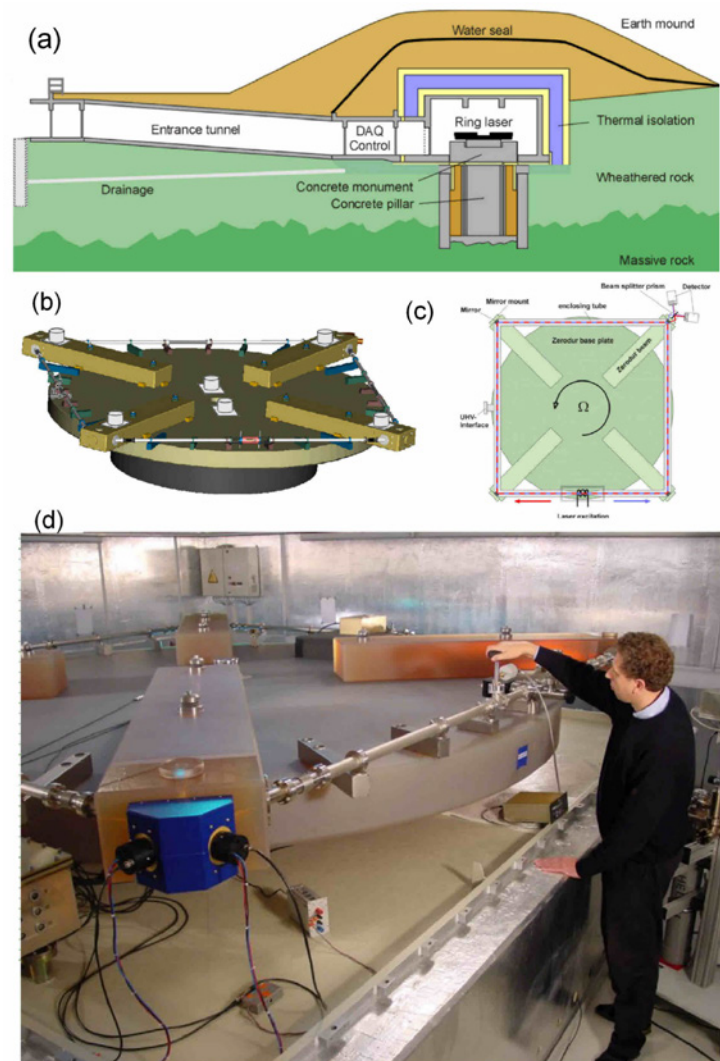
At present, there are ring laser gyros capable of measuring rotations induced by small local earthquakes or distant large teleseisms at four sites: 1.) Cashmere cavern, Christchurch, New Zealand (McLeod et al., 1998), 2.) Wettzell, Germany (Schreiber et al., 2005), 3.) Conway, Arkansas (Dunn et al, 2009), and 4.) Piñon Flat, California (Schreiber et al., 2009a). The ring laser gyros in these 4 sites all measure the vertical component of rotation rate (rotation about the vertical axis). However, measuring the horizontal components (rotations about horizontal axes) using ring laser technology is under development, e.g., VIRGO, Pisa, Italy (Di Virgilio, 2010).

#### 4.1.2 G Ring laser and recording teleseisms

Since 2001, the “G Ring” laser (capable of measuring rotation rate of about  $10^{-12}$  rad/s) has been operating at the primary geodetic station (Fundamentalstation) at Wettzell, in Bavaria (<http://www.fs.wettzell.de/>, last accessed 21 January 2011), along with many other geodetic instruments. A cross-sectional view of the site of the G Ring Laser is shown in Figure 3(a). The instrument is resting on a polished table consisting of a Zerodur™ disk (originally planned for a telescope) (Figure 3(b)) embedded in a 90-ton concrete monument. As shown in Figure 3(a), the monument is attached to a massive 2.7-m diameter concrete pillar and this is founded on crystalline bedrock 10 m below. A system of concrete rings and isolation material shields the monument and pillar from adjacent weathered-rock to eliminate its deformation and heat-flow contributions. The G Ring Laser is protected against external influences by a subsurface installation with passive thermal stability provided by a 2-m layer alternating between Styrofoam and wet clay, this beneath a 4-m soil mound. A lateral entrance tunnel with five isolating doors and a separate control room minimize thermal perturbations during maintenance. In addition the complete Zerodur™ table is covered by a

pressure tight casing which is lowered after maintenance - this results in a huge improvement of the signal/noise ratio.

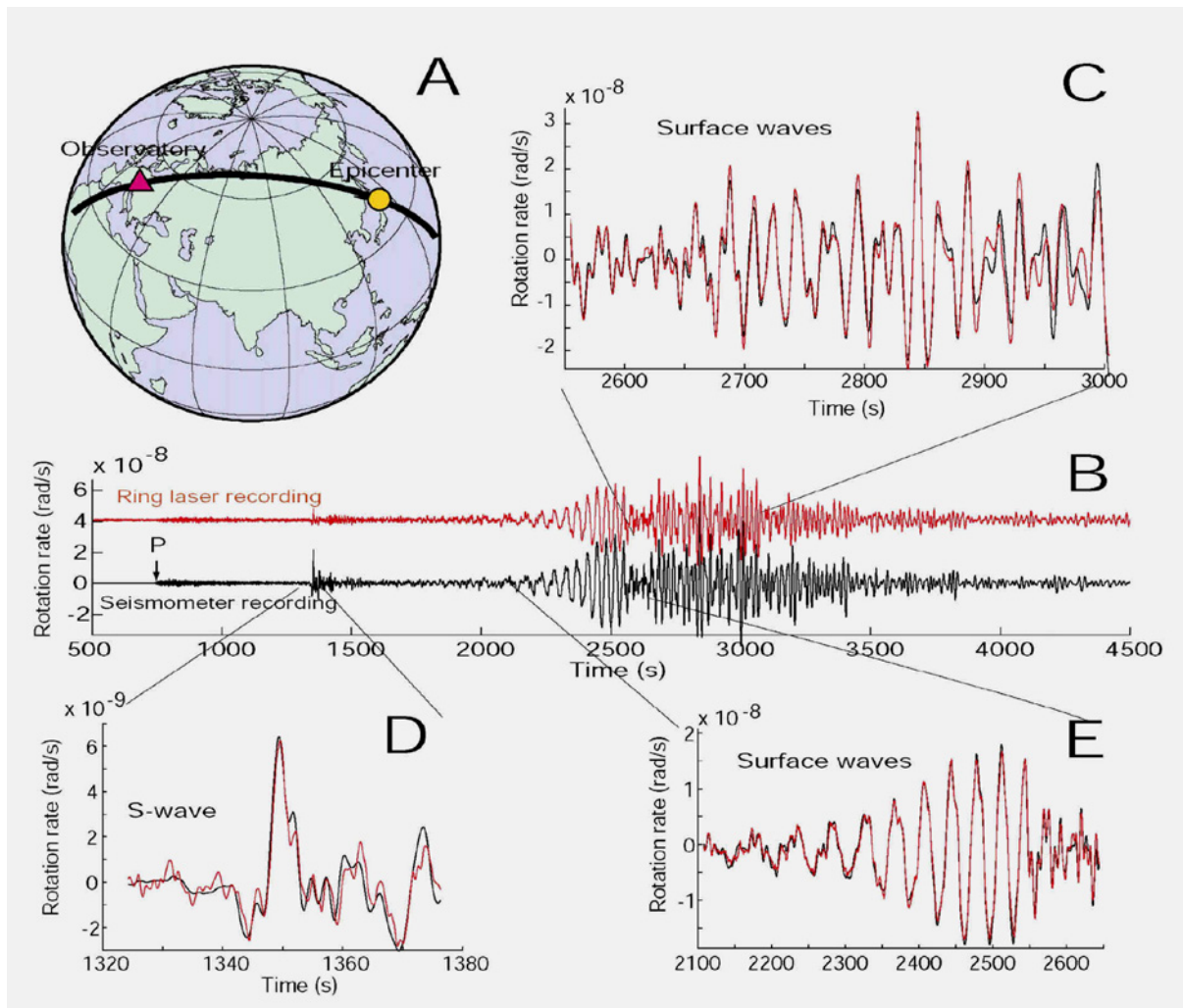
After two years of thermal adaptation, the average temperature reached 12.2 °C with seasonal variations of less than 0.6 °C. Figure 3(c) shows the schematic drawing of instrument, and Figure 3(d) is a photo of the G Ring laser with its designer, Ulli Schreiber. The G Ring laser is large and sit on top of an impressive monument to reach extremely high sensitivity to study variations of the Earth's rotation. For seismological practice, a ring laser on top of a lighter monument will be adequate, as is the ring laser at Piñon Flat, California.



**Figure 3** G Ring laser gyro at the Wettzell Superstation, Germany. (a) Cross-sectional view of the instrument site. (b) Instrument resting on a granite table. (c) Schematic drawing. (d) Photo of G Ring laser gyro with its designer, Ulli Schreiber.

Figure 4 is a comparison of direct point measurements of ground rotations around a vertical axis (red lines) to transverse accelerations (black lines, converted to rotation rate for each time window) for the M8.1 Tokachi-oki earthquake, 25 September 2003 (Igel et al., 2005). Figure 4(a) is a schematic view of the great-circle-path through the epicenter in Hokkaido, Japan, and the observatory in Wettzell, Germany. Figure 4(b–e) show the superposition of the rotation

rate derived from transverse translations (black) and measured directly (red) for various time windows: (b) the complete signal, (c) the latter part of the surface wave train, (d) the direct S-wave arrival, and (e) for the initial part of the surface wave train. These results confirm the expectation from linear elasticity that the waveforms of transverse acceleration and rotation rate (around the vertical axis) should be identical assuming plane harmonic waves. Information on subsurface structure is contained in the ratio between the corresponding motion amplitudes.



**Figure 4** Comparison of direct measurements of ground rotational motions around a vertical axis (red lines) with transverse accelerations (black lines, converted to rotation rate for each time window) for the M8.1 Tokachi-oki earthquake, 25 September 2003 (after Igel et al., 2005).

## 4.2 Strong-motion inertial angular sensors

In aerospace, automotive, and mechanical engineering, smaller rotational-motion sensors are common and generically known as gyroscopic or inertial angular sensors. Nigbor (1994) used a MEMS-based Coriolis rotation-rate sensor to measure the rotational components of strong ground motions close to a large chemical explosion, and unsuccessfully attempted to resolve rotational earthquake motions in southern California for a decade. Similar sensors were used



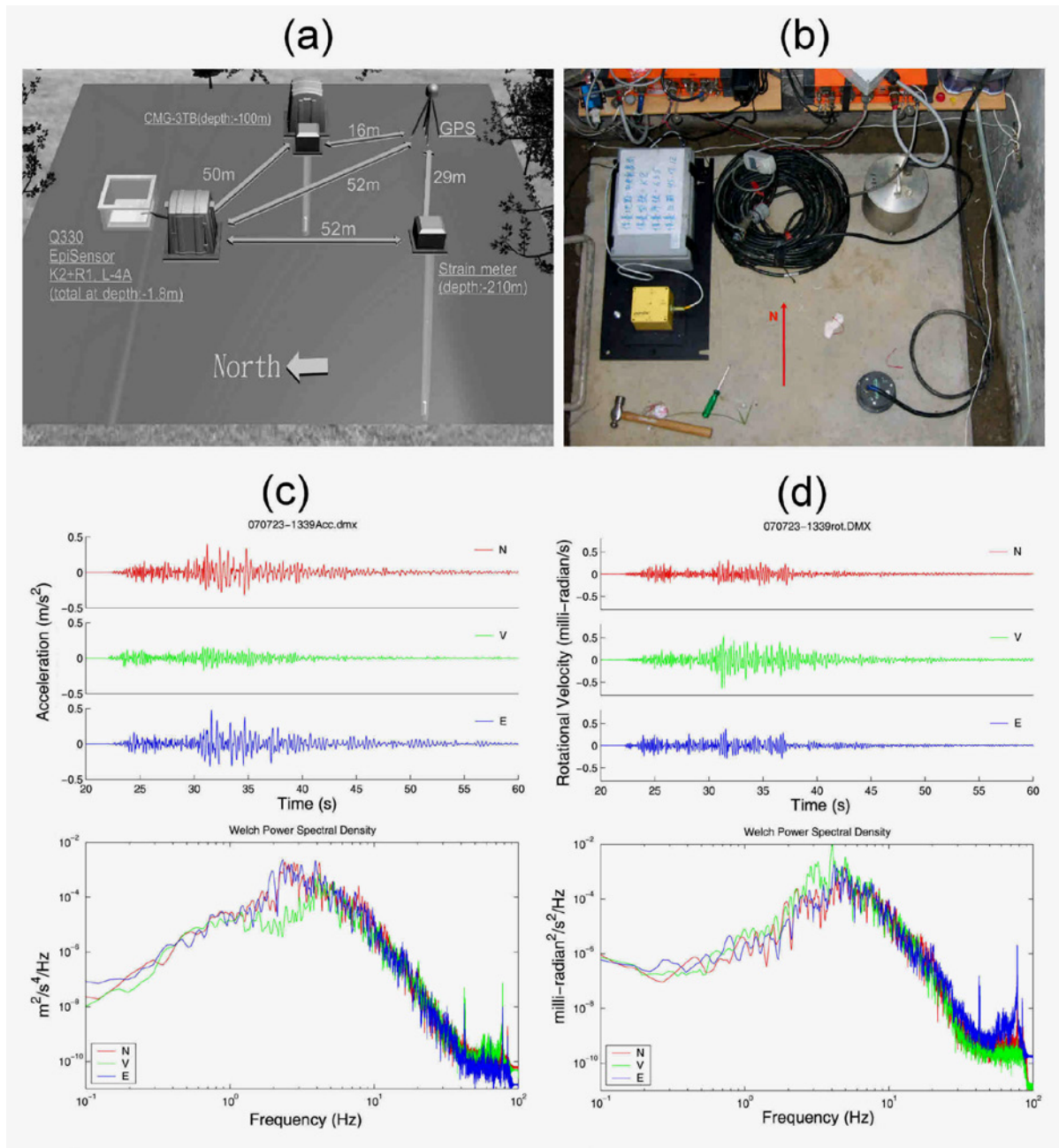
by Takeo (1998) to measure rotational motions from an earthquake swarm ~3 km away. However, such sensors do not have the sensitivity to record rotations from small local earthquakes (magnitude  $\approx 4$ ) at distances of tens of kilometers.

#### 4.2.1 The eentec™ Model R-1™ rotational seismometer

The eentec™ model R-1™ rotational seismometer is the first modestly-priced (about US\$5,000 each) three-component sensor capable of recording small earthquakes at distances up to several tens of kilometers. It uses electrochemical technology in which the motion of an electrolytic fluid inside a torus is sensed electronically, yielding a voltage signal proportional to rotational velocity. (For more information on this pioneering transducer principle see Vol. 1, Chapter 5, subsection 5.3.10). Nigbor et al. (2009) carried out extensive tests of commercial rotational sensors and concluded that the R-1 sensor generally meets the specifications given by the manufacturer but that clip level and frequency response vary from those specifications and between individual channels enough that more detailed calibrations are warranted for each unit. A typical transfer function for the R-1 can be found at the manufacturer's website (<http://www.eentec.com/>, last accessed 21 January 2011). The instrument response is roughly flat from 0.05 to 20 Hz and its self noise (rms) is  $<10 \mu\text{rad/s}$  over the same frequency band.

Schreiber et al. (2011) conducted a comparison test of an R-1 rotational seismometer and a fiber optic gyro  $\mu\text{FORS}$  (Litef GmbH) at the facility of Physikalisch-Technische Bundesanstalt (PTB) in Germany (Department "Längen- und Winkelteilungen"). The two sensors were co-located at the rotation turn table that allows the precise measurements of the revolution angles and provides the time marks for calculation of the rotation rate. The authors found that the R-1 could not adequately detect low-frequency variations of rotation rate, but at frequencies above 0.1-1 Hz, its signal was quite similar to that of the  $\mu\text{FORS}$ . They concluded that "*The declared accuracy characteristics of R-1 are [according to the manufacturer, the] following: frequency bandwidth 0.033-50 Hz, resolution  $1.2 \times 10^{-7} \text{ rad/sec}$ , clip level 1 rad/sec. According to our tests, the resolution, or minimal detectable rotation rate change is about  $2 \times 10^{-5} \text{ rad/sec}$ , and the clip level is  $2.5 \times 10^{-2} \text{ rad/s}$ .*" Moreover, as found by Joachim Wassermann (personal communication, 2011), the R-1 rotational seismometer is very sensitive to temperature. The overall gain changes with increasing temperature, measured from 15 to 50 °C, by a factor of at least 30%.

The R-1 rotational seismometers recorded several hundred local earthquakes and two explosions in Taiwan (Lee et al., 2009b). The top panels in Figure 5 show the instruments deployed at station HGSD in eastern Taiwan (Liu et al., 2009). Figure 5(a) is a schematic drawing of the various seismic, geodetic, and strain instruments there, and Figure 5(b) shows the subset of the instruments deployed in the shallow vault at the left hand side of the upper drawing. These instruments include a data logger (Quanterra Q330), an accelerometer (Kinematics Episensor), a six-channel digital accelerograph (Kinematics K2 with an external rotational seismometer, R-1 by eentec), and a short-period seismometer (Mark Products L-4A). The K2+R-1 instrument is at the left hand side, and the yellow-color box is the R-1 rotational seismometer.



**Figure 5** (a) Schematic of instrument layout at station HGSD in Taiwan. (b) Photo showing the instrument vault containing a K-2 accelerograph (upper left), an R-1 rotational seismometer (yellow box), an EpiSensor accelerometer (lower right-hand side), and an L-4A short-period velocity seismometer (upper right-hand side). (c) Recorded translational accelerations (upper panel) and their spectra (lower panel) from an  $M_w$  5.1 earthquake 51 km from this site. (d) Recorded rotational rates (upper panel) and their spectra (lower panel) from the same earthquake.

The largest peak rotational rate recorded at the HGSD station (to early 2008) is from an  $M_w$  5.1 earthquake at a hypocentral distance of 51 km at 13:40 UTC, 23 July 2007. Figure 5(c) shows the amplitudes and spectra of translational accelerations recorded by the K2's accelerometer. The peak ground acceleration was  $0.47 m/s^2$  (4.8% g) and the two horizontal components have much higher peak amplitude than the vertical. Figure 5(d) shows the amplitudes and spectra for rotational rates recorded from the external R-1 seismometer. The



peak rotational rate was 0.63 mrad/s for the vertical component, which is more than twice the rotational rate recorded on either of the horizontal components. The dominant frequency band in ground acceleration is about 2–5 Hz (horizontal components), whereas for the ground rotation rate it is about 2.5–5.5 Hz for the vertical component. Other studies report observations made with the R-1 sensor and compare these point measurements of rotation to array-derived area rotations (e.g., Wassermann et al., 2009).

A new model, the R-2™, became available from eentec in 2010. As of this writing, laboratory tests on twelve R-2 units indicated that their performance did not meet the technical specifications given by the manufacturer, though the design itself appears sound (Evans et al., 2010). Other promising efforts to develop new rotational sensors include that of Northrop Grumman Lites GmbH, Freiburg, Germany, which is developing a rotational sensor specifically for seismological applications and based on fiber optic gyroscope principle, and that of ATA Sensors (Smith, 2010), which is applying an entirely different technology to develop sensors for weak motion (using a solid proof mass) and for strong motion (using their existing patented magneto-hydrodynamic, or MHD, technology).

#### 4.2.2 Laboratory calibration of rotational sensors

Most rotational sensors output a signal proportional to rotational velocity but a few produce rotational acceleration. In either case the only natural reference signal commonly available is the Earth's rotation about its pole, which is about 73  $\mu$ rad/s about the North Pole (right-handed rotation vector pointing north). While the Earth's rotation is a very useful calibration signal for weak-motion rotational sensors such as large-area ring-laser gyroscopes, it is near or below the noise level of many existing strong-motion sensors, and so is effectively unavailable for their calibration. Further, Earth rotation is a roughly 0-Hz signal ("DC"), and thus it is not useful for establishing any rotational sensor's linearity or transfer functions. Thus, strong-motion rotational sensors require the use of a precision rotational shake table, and these are costly. In what follows, we focus on calibration of strong-motion rotational sensors for which the Earth's rotation is not a viable proof signal.

In all cases, operators are advised to use a right-handed set of axes including right-handed definitions of rotational pseudovectors (e.g., Evans et al., 2009, their Figure 1). Note, of course, that many seismic instruments record in the channel order Up, North, East; so  $x$  should be thought of as East,  $y$  as North, and  $z$  as Up for the sensor to be right handed.

IEEE Standard 671-1985 (1985) provides useful guidance, while procedures specific to seismic rotational instruments are in development and will draw on both this standard and experiences at several laboratories in the early years of testing such instruments. At present, the primary concerns of those laboratories are the transfer function (response function) of rotational sensors, the self-noise levels of these sensors, and the cross-axis terms, including both rotation-to-rotation and translation-to-rotation cross-axis sensitivities. Existing works on testing rotational seismic sensors include Nigbor et al. (2009), Evans et al. (2010), Schreiber et al. (2011), and various private reports of sensor testing by the laboratories active in this field.

Transfer function. Early rotational-velocity sensors for strong-motion work have proved to be less than flat in response and to vary even in mid-band sensitivity by far more than the 1% industry norm for sensitivity accuracy. Thus, it is essential that all such sensors be calibrated individually on a shake table to determine amplitude and phase responses, from which

Laplace poles and zeros may be estimated by a parametric fit. In practice, this calibration proceeds by inputting known sine wave motions on axis with the sensor under test, and that intervals of at least three cycles of each sine frequency (preferably many more) be recorded and compared to these known inputs. Alternatively, a slowly swept sine wave spanning the same frequency band may be used. This test generally should proceed at half-octave steps in frequency from one octave below the low-frequency corner of the sensor to one octave above its high-frequency corner (some shake table systems cannot drive the sensors to this peak frequency, but one should get as high as possible without damaging the table). The sensor output and the input motions must be recorded on a common time base (ideally with simultaneous sampling to within about 10  $\mu$ s) in order to recover phase accurately. In practice, it is common to use either the positional feedback of the shake table or a reliable reference sensor to measure this input motion, so the resulting transfer function is relative to this reference signal. Empirical evidence suggests that amplitude response is best measured by comparing peak power between PSDs of the test and reference traces while phase appears to be most stably measured by cross correlation of the traces. In co-testing different instruments one should keep them near and at the same distance from the rotation point of the table to limit centrifugal inputs.

Self noise. At present, strong-motion rotational sensors appear to have self noise levels well above Earth's ambient rotational seismic background motions. The level of rotational Earth noise is not very well known at present. However, the absence of microseism-band increases in typical instrument-noise figures argues against a significant Earth-noise input. Thus, the single-sensor method of *Evans et al. (2010)* may be used until sensor noise levels drop and sensitivities rises (The same will not be true for weak motion sensors.) When that stage of sensor development is reached, there is no reason in principal not to use the two- or three-sensor methods described in that paper and works cited therein. The single-sensor method is to record sensor output with a sensitive recorder at some quiet location during a quiet interval, generally overnight. One computes the PSD and simply attributes all of it to sensor self noise. Details of a "standard" method for this computation and subsequent "operating range diagram" representation of sensor noise are given by *Evans et al. (2010)*.

Rotation-to-rotation cross-axis sensitivity. Industry norms for translation-to-translation and rotation-to-rotation cross-axis sensitivities are typically about 1%. For example, a 1-g input to a translational accelerometer oriented at right angles to that input acceleration should register no more than 0.01 g in response. The unit "g" means Earth surface gravitational acceleration, typically about 9.8 m/s<sup>2</sup>. It is a unit preferred by the earthquake engineers. This 1% norm is reasonable for seismology and earthquake engineering applications as well. That is, a rotation rate of 1 mrad/s should register on the orthogonal components as no more than a 10  $\mu$ rad/s output signal. In most sensors, imperfect alignment between the sensor case and the true active axes of the sensor dominate this cross-axis term (1% corresponds to an alignment error of about 0.6°).

Translation-to-rotation cross-axis sensitivity. It can be difficult to compare motions with differing units and fundamental physics, in this case translation and rotation, but there is a simple method that accounts for the wide anticipated use of rotational sensors for correcting tilt-induced errors in horizontal translational sensors. Tilt within Earth's gravitational field applies  $\sim 1 \mu$ g of acceleration to horizontal translation sensors for every 1  $\mu$ rad of tilt. This is a first-order sensitivity in the presence of any significant rotational motions. Thus, one should compare the output of a rotational sensor excited by translational input motions to the true translational inputs *via*  $g \sin \theta$ , where  $g$  is local gravitational acceleration and the translational sensor is rotated by an angle of  $\theta$  about a horizontal axis (i.e., tilted). Angle  $\theta$  is generally

obtained by integrating rotational velocity to rotation angle in a collocated rotational sensor. Because rotational motions in earthquakes are relatively small, perhaps 1 to 10% of translational motions when so compared, the translation-to-rotation cross-axis sensitivity of rotational sensors must be well under one part per thousand (1 PPT) in order to correct the translational sensors accurately. A similar level of immunity to translational motions is needed to make reliable use of rotational data for other purposes, such as single-site computation of surface-wave phase velocity or reducing the size of response kernels in seismic tomography. This restriction is quite stringent but has been met by well designed sensors.

#### 4.2.3 Field deployment of rotational sensors

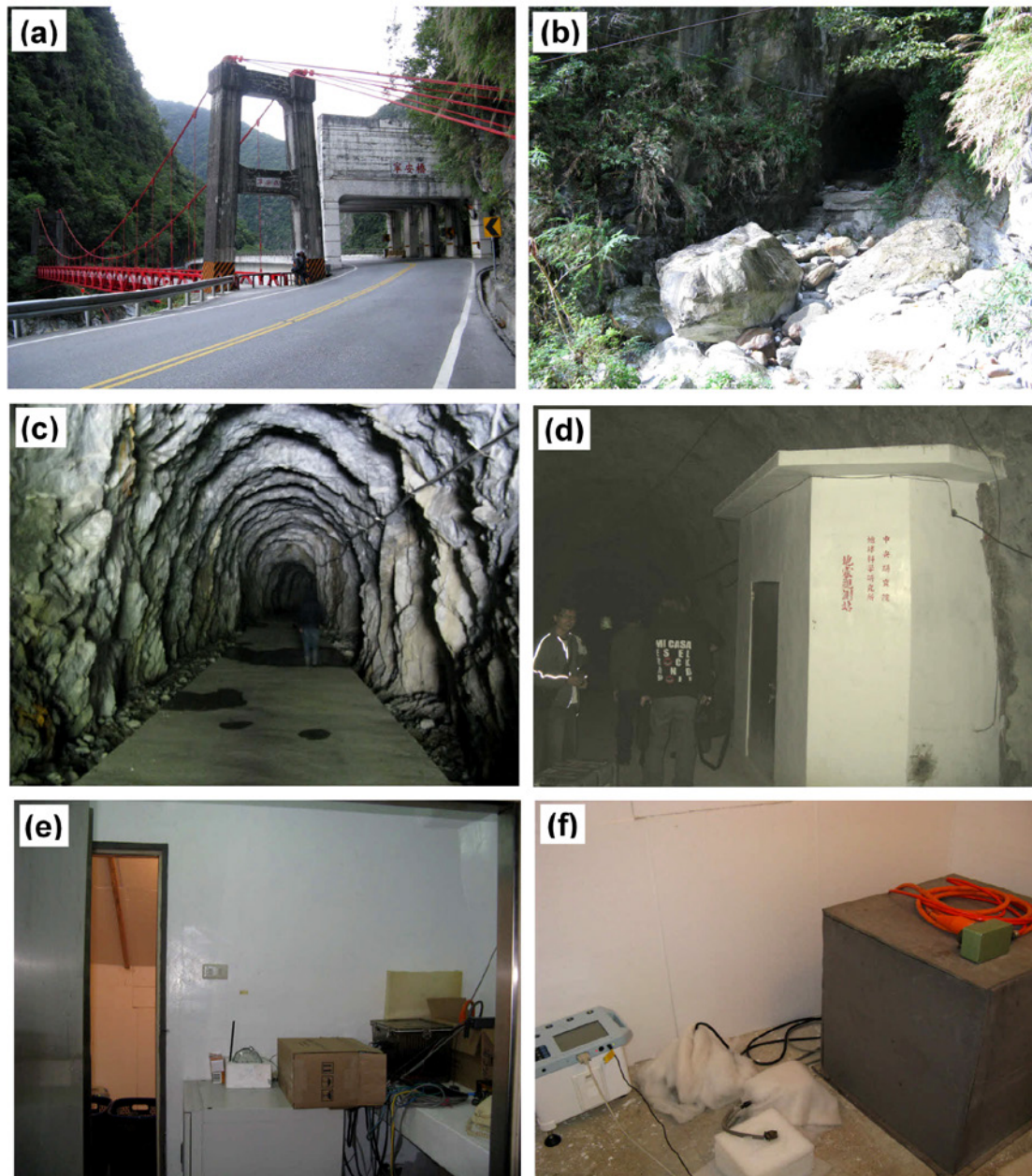
Field deployment of rotational sensors requires site selection and preparation similar to that for any seismological instruments (Trnkoczy, et.al., 2002 and also Chapter 7 in NMSOP2). The R-1 rotational seismometer can be connected to a data logger, as with any other analog-output seismic sensor. An earlier approach in Taiwan was to connect the R-1 to a six-channel accelerograph (Kinometrics model K2) forming a six-degree-of-freedom (6-DOF) self-contained instrument. In this case, the K2 and R-1 are mounted together on a thick metal plate. A vault was constructed at the HGSD station to house this K2+R1 instrument. In addition, a six-channel, 24-bit Quanterra Q330 data logger was deployed to record ground acceleration using a Kinometrics EpiSensor and to record ground velocity using a Mark Product short-period seismometer (Model L-4A; 2 Hz natural frequency). Instruments in operation during the Phase 2 operation of Liu et al. (2009) have been shown in Figure 5 above.

Another configuration used in Taiwan is to connect an R-1 rotational seismometer directly to a 24-bit Quanterra Q330 datalogger. Data are telemetered from the field to the Institute of Earth Sciences in Taipei in real time at 20 sps (samples per second), and continuous data are also recorded on an external hard disk in the field at 100 sps. The disk is collected every three to four months when the station is being serviced. Figure 6 (on next page) shows an installation of broadband seismometer, accelerometers, and rotational sensor at the Ninganchiao station (NACB) operated by the Institute of Earth Sciences, Academia Sinica, Taiwan. Station NACB is near the back of a 60-m tunnel into a hill composed of marble, and is located at 24.1738° N, 121.5947° E, and 130 m above the sea level. This station was established in 1995 as one of the Broadband Array in Taiwan for Seismology (BAT) stations (<http://bats.earth.sinica.edu.tw/>, last accessed 21 January 2011). In April, 2008, a Model R-1 rotational seismometer of eentec and an accelerometer (Model TSA-100S of Metrozet) were installed at this site. This set of instruments forms the NAC1 station and is recorded by a 6-channel Q330 datalogger of Quanterra. In 2010, the Metrozet accelerometer was replaced by a Model AGI 520 tiltmeter.

## 5 Discussion of results and benefits of studying rotational motions

Many authors have emphasized the benefits of studying rotational motions (e.g., Twiss et al., 1993; Spudich et al., 1995; Takeo and Ito, 1997; Teisseyre et al., 2006; Trifunac, 2006; Igel et al., 2007; Fichtner et al., 2009; and Teisseyre, 2010), and its relevance to engineering applications (e.g., Trifunac, 2009; Castellani and Guidotti, 2010). Rotational seismology is also of interest to physicists using Earth-based observatories for detecting Einstein's gravitational waves (e.g., Lantz et al., 2009) because they must correct for the underlying

Earth motion (e.g., DeSalvo, 2009; Di Virgilio, 2010). Here we briefly discuss some basic issues.



**Figure 6** Photos of an installation of broadband seismometer, accelerometers, and rotational sensor at the Ninganchiao station (NACB) operated by the Institute of Earth Sciences, Academia Sinica, Taipei, Taiwan: (a) The Ninganchiao Bridge in the National Park northwest of Hualien, Taiwan, (b) the entrance to a tunnel near the Ninganchiao Bridge, (c) the tunnel of about 60 m, leading to the Ninganchiao station, (d) the front entrance of the Ninganchiao station, (e) the recording room, showing a 6-channel Q330HR recorder to the right and the entrance to the instrument room, (f) the STS-2 broadband velocity seismometer, which is enclosed in a metal box to the right, together with two accelerometers (Model Episensor ES-T of Kinemetrics, and Model TSA-100S of Metrozet) in the middle (covered by insulation material), and a Model R-1 rotational seismometer of eentec near the bottom to the left (also covered by insulation material).

## 5.1 Linear and nonlinear elasticity

Real materials of the Earth are heterogeneous, anisotropic, and nonlinear, especially in the damaged zones surrounding faults and in poorly consolidated sediments, soil, and weathered, fractured rock near the surface. These characteristics are typical of site conditions just beneath seismic instruments, particularly for strong-motion instruments that are intended to record the sort of motions experienced by shallow-foundation structures, and in consequence, commonly are placed on the surface. In the presence of significant nonlinearity we are forced to consider the mechanics of chaos (Trifunac, 2009), and we must record both the rotational and translational components of strong motion in order to interpret such complexities.

Seismology is primarily based on the classical, linear elasticity theory, which is applicable to simple homogeneous materials under infinitesimal strain. “Curl” rotation is defined for such materials as the curl of the displacement field in Equation (2), and in this classical elasticity theory, the rotational components of motion are contained in the S waves. Meanwhile, continuum mechanics has advanced far beyond such classical theory. In particular, the elasticity theory of the Cosserat brothers (Cosserat and Cosserat, 1909) incorporates both a local rotation of continuum particles and the translational motion assumed in classical theory, as well as a couple stress (a torque per unit area) in addition to the usual the force stress (force per unit area). In the constitutive equation of classical elasticity theory for an isotropic material there are two independent elastic constants, while in Cosserat elastic theory there are at least six elastic constants. Pujol (2009) provides a tutorial on rotations in the theories of finite deformation and micropolar (Cosserat) elasticity. Twiss (2009) derives an objective asymmetric micropolar moment tensor from a discrete-block model for a deforming granular material. He also investigates seismogenic deformation associated with volumes of distributed seismicity in three different geographic areas, and finds support in the micropolar model for the effects of a granular substructure on the characteristics of seismic focal mechanisms.

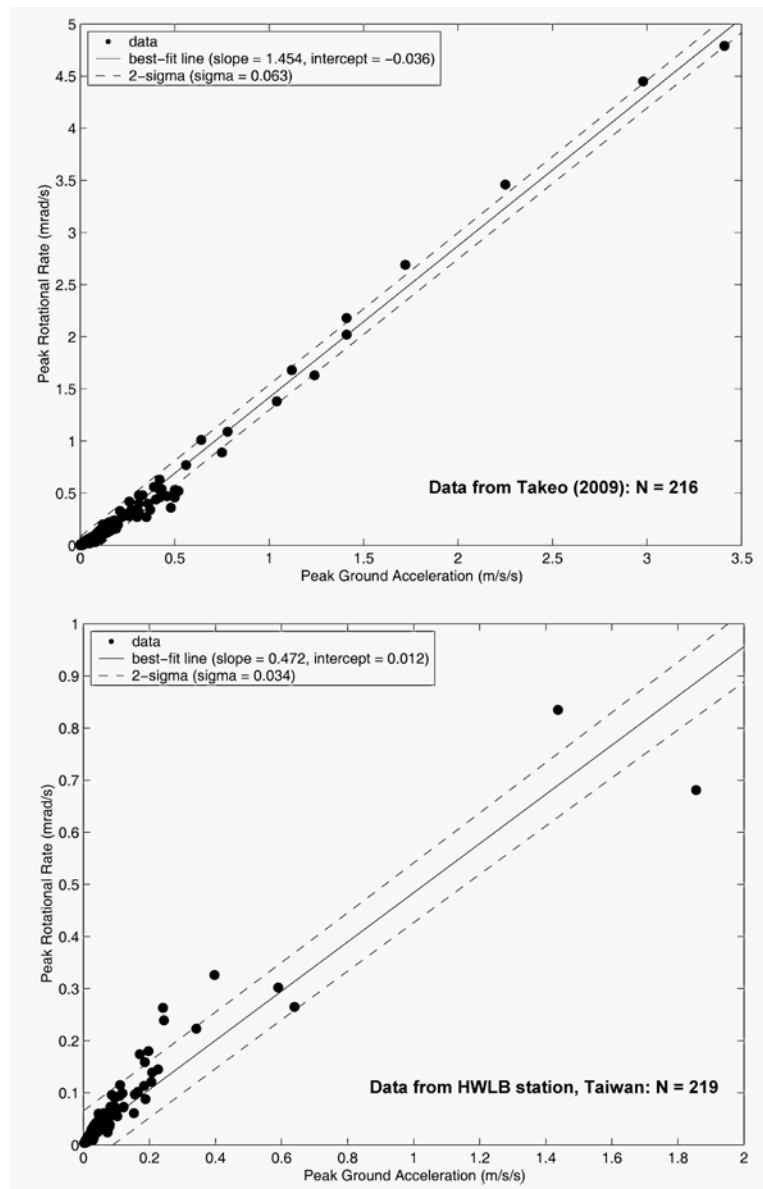
Angular squeeze strain deformations were found by Gombert and Agnew (1996) for all earthquakes they studied on the basis of their experimental strainmeter data. Such squeeze circular deformations cannot be explained in a frame work of the classical continuum theory, and therefore, the authors attributed such founding to some systematic errors. However, as pointed out by Roman Teisseyre (personal communication, 2011), these deformation may exist without such errors according to the asymmetric continuum theory (Ayden et al., 1993; Teisseyre, 2010).

## 5.2 Near-field seismology

Although the observed rotational motions for teleseisms recorded in a rock vault agree well with estimates based on the classical elasticity theory (e.g., Igel et al., 2005; 2007), this is not the case for local earthquakes recorded by strong-motion installations. As first noted by Takeo (1998) and later confirmed by Lee et al. (2009b), observed rotational rates from local earthquakes are much larger than those predicted by the classical elasticity theory. For example, Bouchon and Aki (1982) determined theoretically that a maximum rotational rate of 1.5 mrad/s is expected for an M6.5 earthquake, whereas maximum rotational rates exceeding 1 mrad/s have been observed for much smaller (M4.5–5.5) earthquakes in Japan and Taiwan, for which the classical elasticity theory predicts far smaller rotations. Takeo (1998) reported the largest rotational rate to be 26 mrad/s around the north-south horizontal axis from the second largest earthquake (M5.2 at 14:09 UTC, 03 March) during the 1997 swarm, east of Cape Kawana, offshore Ito, Japan (epicentral distance ~3 km). As of the end of 2009, the

largest rotational rate recorded at the HWLB station (Hualien, Taiwan) was 2.58 mrad/s around the east-west horizontal axis for an  $M_w$  6.4 earthquake offshore that occurred at 13:02 UTC on 19 December 2009, and was located offshore at a hypocentral distance of about 49 km. The peak rotational rate was 1.57 mrad/s around the north-south axis, and 0.68 mrad/s around the vertical axis. The corresponding peak ground accelerations were 1.16, 1.85, and 0.50  $\text{m/s}^2$  for the east-west, north-south, and vertical component, respectively (5.1–18.9% g).

Figure 7 shows the vertical-axis peak rotational rate versus horizontal-axis peak ground acceleration for the earthquake data set from Takeo (2009) (top frame), and for the earthquake data set (22 August 2008 to 25 December 2009) recorded (at different distance ranges) at station HWLB, Taiwan (bottom frame; N is the number of data points).



**Figure 7** Vertical peak rotational rate (PRR) versus horizontal peak ground acceleration (PGA) for the earthquake data set from Takeo (2009) (**top frame**), and for the earthquake data set (22 August 2008 to 25 December 2009) recorded at the HWLB station, Taiwan (**bottom frame**). N is the number of data points. Best-fit lines are determined using linear least-squares. Note the difference in the scaling of the axes for the two plots.



The slopes for the two data sets (1.454 versus 0.472) are significantly different. But, as noted below, this may reflect a difference in site conditions. While the Taiwan data appears to exhibit more scatter than in the Takeo (2009) data, its variance actually is only about half. One should also note that the range of the observations in the two plots is significantly different, which may account, at least in part, for the difference in scatter. Whereas the earthquake sources for the Takeo (2009) data belong all to a compact, nearby swarm, the Taiwan data relate to widely separated earthquake sources. Several authors have noted similar linear relationships (e.g., Spudich and Fletcher, 2008; Stupazzini et al., 2009; Takeo, 2009; and Wang et al., 2009). In particular, Takeo (2009) showed “*a linear correlation between the maximum rotational displacements around vertical axis and the maximum [ground] velocities*”. The two plots in Figure 7 are equivalent to Takeo’s linear relationship, without the need to perform the integration for the measured rotational rate and ground acceleration to obtain rotational displacements and ground velocities. The linear slope in Figure 7 has units of s/km, i.e., of slowness. Spudich and Fletcher (2008) interpreted this “slowness” as the inverse of an “apparent velocity”, characterizing the seismic wavefield beneath the recording station.

Spudich and Fletcher (2008) computed peak values of ground strain, torsions, and tilts for the 2004 Parkfield earthquake ( $M_w$  6.0) and four of its aftershocks ( $M_w$  4.7–5.1) using the data recorded by the UPSAR array of accelerators. The observed peak horizontal acceleration and velocity were 0.45 g and 27 cm/s during the mainshock, and derived a maximum rotation rate of 1.09 mrad/s for the vertical component, which agrees well with the theoretical result of Bouchon and Aki (1982) mentioned above. Takeo (2009) noted that his observations of peak rotation values in the near-field region during an 1998 earthquake swarm offshore Ito, Izu Peninsular, Japan, are much larger than those calculated by Spudich and Fletcher (2008), and he proposed the following explanations: the different spatial scale of rotational motion measured by a single point gyro measurement and by an array observation (that is, the point observation not benefiting from averaging out local site variations), the effects of topography, and the difference of the degree of geologic maturation between the San Andreas fault and the swarm volume offshore Ito. Two arrays of both rotational and translational sensors have been deployed by Wu et al. (2009) in Taiwan to study this discrepancy.

### 5.3 Using explosions to study rotational motions

Because large earthquakes rarely occur near existing seismic stations, explosions have been used by several pioneers to study rotational motions (e.g., Graizer, 1991; Nigbor, 1994). Lin et al. (2009) deployed an array of eight triaxial rotational sensors, 13 triaxial accelerometers, and 12 six-channel, 24-bit data loggers with GPS time receivers to record two explosions in northeastern Taiwan. These instruments were installed at about 250 m (1 station), 500 m (11 stations), and 600 m (1 station) from the explosions. The group of 11 stations forms a “Center Array” with station spacing of about 5 m. The code name for the first shot with 3,000 kg explosives is “N3P”, and that for the smaller second shot (750 kg) is “N3”. Although the N3P shot used four times as much explosive as shot N3, the peak ground translational acceleration and rotational rates at all 13 station sites for N3P are only about 1.5 times larger than that for N3, suggesting some nonlinear process. Large variations (by tens of percent) in translational accelerations and rotational rates were observed within the small Center Array. The largest observed peak rotational rate was on the horizontal transverse component at 2.74 and 1.75 mrad/s at a distance of 254 m from the N3P and N3 shots. As pointed out by Joachim Wassermann (personal communication, 2011), this experiment may have problems with strain-tilt coupling that Lin et al. (2009) did not consider.

The acceleration data from these two explosions were used by Langston et al. (2009) to compute acceleration spatial gradients, horizontal strains and horizontal rotations, and to perform a radiometric analysis of the strong ground motion wave train. The analysis yields a complex, frequency-dependent view of the nature of seismic wave propagation over short propagation distances that imply significant lateral velocity changes in the near-surface crustal structure. Areal strain and rotation about the vertical have equal amplitudes and suggest significant wave scattering within the confines of the river valley where the experiment was performed and/or significant departure from an axisymmetric explosion source. Gradiometry shows that the P wave arrives at the array 35 degrees off-azimuth clockwise from the straight-line path and appears to have been refracted from the northern side of the valley (the shots are west of the Center Array). Chi et al. (2011) successfully recovered the first order features of vertical rotation-rate ground motions from the translational velocity waveforms in the frequency band 0.5–20 Hz by using the software of Spudich and Fletcher (2009); they deduced strain as large as  $10^{-4}$ . To fulfill the uniform-rotation assumption in the linear elasticity theory, one must use a small-aperture array. However, inverting data from an array of small spatial dimension requires accurate waveforms of high signal-to-noise ratio and high sampling rates. Since waveforms from adjacent stations are very similar, low levels of noise can have a strong influence on estimated displacement gradients. The recordings of Lin et al. (2009) were limited to 200 samples per second (sps). However, much higher sampling rates are required in the near field recording of explosions, which also radiate rather high-frequency (>100 Hz) waves.

## 5.4 Processing collocated measurements of translations and rotations

Processing collocated observations of rotation and translation is routinely performed in the inertial navigation units of aircraft and other vehicles. A similar analysis is possible for various combinations of strain components, rotations, and translations in seismology and earthquake engineering. With the exception of velocity-strain combinations (e.g., Gomberg and Agnew, 1996) these methods were largely unexplored until the recent work of Lin et al. (2010). Lin et al. demonstrated an appropriate set of these equations for earthquake engineering and seismology, and used them to recover inertial-frame displacements and rotations. Further, it is already apparent that rotational motions provide useful additional analysis opportunities — simply put, more data at a site yield more results:

Phase velocities and propagation directions. A simple calculation for non-dispersive linear-elastic plane waves with transverse polarization shows that the ratio of transverse acceleration to vertical-axis rotational velocity is proportional to local phase velocity. This result implies that information on subsurface velocity structure (otherwise only accessible through seismic array measurements and combined analyses) is contained in any single-point measurement that includes rotational sensors. It has been shown that such ratio-derived phase velocities agree with velocities predicted by theory (Igel et al., 2005; Kurrle et al., 2010). In a recent theoretical study based on full ray theory for Love waves (normal mode summation) Ferreira and Igel (2009) demonstrated that the Love wave dispersion relation also can be obtained by taking the spectral ratio of transverse acceleration to vertical-axis rotation rate. This result implies that seismic surface wave tomography is possible without requiring sub-arrays to determine local mean phase velocities. Information on the direction of propagation also is contained in the azimuth-dependent phase fit between rotations and translations. Since this fit is optimal in the direction of propagation the back azimuths can be estimated to within a few degrees (Igel et al., 2007). Linking observational translations, strains, and rotations together



also is advocated by Langston (2007) to yield a snapshot of the wavefield including direction, slownesses, and radial/azimuthal amplitude gradients independently at each such station.

Towards a new kind of tomography. The possibility of deriving local dispersion relations from single-station records leads to the question of what subsurface volume one resolves and to what degree depth-resolution velocity perturbations can be recovered. The method of choice to answer this type of question is the adjoint method (Fichtner and Igel, 2009), with which sensitivity kernels (first Fresnel zones) can be calculated to indicate the volume in which the observable parameters (typically travel times) are sensitive to structural perturbations. Fichtner and Igel (2009) introduced a new observable quantity — apparent shear wave velocity — which is a time-windowed ratio of the moduli of translational velocity and rotation angle. It turns out that sensitivity near the source vanishes, leading to a new type of resolution kernel with high sensitivity only in the vicinity of the receiver and in a somewhat smaller portion of that volume than the kernels of translational motions alone. This result implies that a superior tomographic inversion for near-receiver structures based on rotations and translations is possible. Synthetic tomographic inversions of this kind are given in Bernauer et al. (2009).

Scattering properties of the crust: Partitioning of P and S waves. The partitioning of P and S energy and stabilizing the ratio between the two is an important constraint on the scattering properties of a medium. Igel et al. (2007) discovered surprisingly great rotational energy in a time window prior to teleseismic S, belonging to the P-wave coda. Detailed analysis of the signals and modelling of wave propagation through three-dimensional random media demonstrate that these signals can be explained with P-SH scattering in the crust with scatterers of roughly 5-km correlation length and rms perturbation amplitude of 5% (the latter is better constrained). Additionally, rotation measurements can be used as a filter for SH type motion (Takeo and Ito, 1997). These results further illustrate the efficacy of rotation measurements in their own right.

## 6 Conclusions

Seismology has been very successful in vaults in the *far field* because classical elasticity theory works very well for interpreting recorded translational motions of bedrock at large distances. Because of this success and limited instrumentation options, most funding for earthquake monitoring historically has gone into global and regional seismic networks using only translational seismometers. However, to improve our understanding of damaging earthquakes we must also deploy rotational and translational instruments in the *near field* of active faults where potentially damaging earthquakes (magnitude > 6.5) occur *infrequently*. For strong-motion seismology and engineering, this requires long-term commitment to monitoring because a damaging earthquake on any given fault may not take place for hundreds of years. Given that financial resources are often limited, recording ground motions in the near field would require extensive seismic instrumentation along some well-chosen active faults and a bit of luck.

Ring laser observations at Wettzell, Germany, and at Piñon Flat, California, demonstrated consistent measurements of rotational ground motions in the *far field*. So far this success can only be demonstrated with one component of rotation. The high cost of present high-precision ring laser gyros makes widespread deployment of these instruments unlikely. Less expensive (but less sensitive) alternatives are being pursued by several academic groups (Cowsik et al., 2009; Dunn et al., 2009; Jedlička et al., 2009; Schreiber et al., 2009b;

Takamori et al., 2009; Brokesova, 2010; Kozák, 2010; and Velikoseltsev, 2010), and by commercial companies (e.g., Northrop Grumman Litef GmbH, and ATA Sensors).

Schreiber (2011, personal communication) wished to emphasize that their development of a fiber optic gyro (FOG), as reported in Schreiber et al. (2009b), has shown that FOGs are very suitable for measuring ground rotation in the near field of earthquakes, and their recent deployment of a FOG in a building on the campus of the University of Canterbury in Christchurch, New Zealand recorded clear rotational motions in November 14, 2010, from aftershocks of the 3 September 2010  $M_w$  7 earthquake nearby.

As of late 2010, only Taiwan has a modest program (Lee et al., 2009b) to monitor both translational and rotational ground motions from local and regional earthquakes at several permanent seismic stations, as well as at two arrays in a building and a nearby free-field site. These two arrays are designed to capture a repeat of the 1906 Meishan earthquake (magnitude 7.1) in the *near field* with both translational and rotational instruments in 6-DOF point and array observations (Wu et al., 2009).

Based on the developments described in the BSSA Special Issue on rotational seismology and engineering (Lee et al., 2009a), observation, analysis, and interpretation of both rotational and translational ground motions is beginning to play a significant role in seismology and earthquake engineering. An International Working Group on Rotational Seismology (IWGoRS) was organized in 2006 to promote such investigations, evaluate their implications, and share experience, data, software, and results in an open Web-based environment (Todorovska et al., 2008). Anyone can join IWGoRS at <http://www.rotational-seismology.org> (last accessed 21 January 2011), subscribe to the mailing list, and contribute to the content (publications, data, links, etc.). Finally, the Second IWGoRS Workshop (convened by Johana Brokesova and Heiner Igel) took place 10–13 October 2010 in Prague, Czech Republic (<http://www.rotational-seismology.org/events/workshops/2010/2nd-iwgors-workshop>, last accessed 21 January 2011) and included contributions from many institutions in several countries.

## Acknowledgements

We wish to thank Riccardo DeSalvo, Angela Di Virgilio, Vladimir Graizer, Jan Harms, Brian Kilgore, Jan Kozák, Ulli Schreiber, Chris Stephens, Roman Teisseyre, and Joachim Wassermann for their comments and suggestions, which greatly improve our manuscript.

## References

- Agnew, D. C. (1986). Strainmeters and tiltmeters. *Rev. Geophys.*, **24**, 579-624.
- Aki, K., and Richards, P. G. (2002). *Quantitative Seismology*. 2<sup>nd</sup> Edition, University Science Books, Sausalito, CA.
- Ayden Ö., Akagi, T., and Kawamoto, T. (1993). The squeezing potential of rocks around tunnels: Theory and prediction. *Rock Mech. Rock Eng.*, **2**, 137-163.
- Bernauer, M., Fichtner, A., and Igel, H. (2009). Inferring Earth structure from combined measurements of rotational and translational ground motions. *Geophysics*, **74**(6), WCD41-WCD47, doi:10.1190/1.3211110.

- Bodin, P., Gomberg, J., Singh, S. K., and Santoyo, M. (1997). Dynamic deformations of shallow sediments in the valley of Mexico, part I: Three dimensional strains and rotations recorded on a seismic array. *Bull. Seism. Soc. Am.*, **87**, 528–539.
- Bouchon, M., and Aki, K. (1982). Strain, tilt, and rotation associated with strong ground motion in the vicinity of earthquake faults. *Bull. Seism. Soc. Am.*, **72**, 1717–1738.
- Brokesova, J. (2010). Rotaphone - A self-calibrated mechanical seismic sensor for field rotation rate measurements. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Abstract and presentation available at: <http://iwgors.pavucina.org/program.php>, last accessed 21 January 2011].
- Castellani, A., and Guidotti, R. (2010). Free field rotations: Relevance on buildings in near field. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Castellani.pdf>, last accessed 21 January 2011].
- Chi, W. C., Lee, W. H. K., Lin, C. J., and Liu, C. C. (2011). Inversion of ground motion data from a seismometer array using a modification of Jeager's method. [Submitted to the *Bull. Seism. Soc. Am.*, November, 2010.]
- Cochard, A., Igel, H., Schuberth, B., Suryanto, W., Velikoseltsev, A., Schreiber, U., Wassermann, J., Scherbaum, F., and Vollmer, D. (2006). Rotational motions in seismology: theory, observation, simulation. In: Teisseyre, R., Takeo, M. and Majewski, E. (ed.), *Earthquake Source Asymmetry, Structural Media and Rotation Effects*, Springer Verlag, Heidelberg, 391–411.
- Cosserat, E., and Cosserat, F. (1909). *Théorie des Corps Déformables*, Paris: Hermann (available from the Cornell University Library Digital Collections) (in French).
- Cowsik, R., Madziwa-Nussinov, T., Wagoner, K., Wiens, D., and Wyssession, M. (2009). Performance characteristics of a rotational seismometer for near-field and engineering applications. *Bull. Seism. Soc. Am.*, **99**(2B), 1181–1189.
- DeSalvo, R. (2009). Review: accelerometer development for use in gravitational wave-detection interferometers. *Bull. Seism. Soc. Am.*, **99**(2B), 990–997.
- Di Virgilio, A. (2010). Installation of the ringlaser G-Pisa inside the Virgo area. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: [http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Di\\_Virgilio.pdf](http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Di_Virgilio.pdf), last accessed 21 January 2011 ]
- Droste, Z., and Teisseyre, R. (1976). Rotational and displacemental components of ground motion as deduced from data of the azimuth system of seismograph. *Pub. Inst. Geophys. Polish Acad. Sci.*, **97**, 157–167.
- Dunn, R. W., Mahdi, H. H., and Al-Shukri, H. J. (2009). Design of a relatively inexpensive ring laser seismic detector. *Bull. Seism. Soc. Am.*, **99**(2B), 1437–1442.
- Evans, J. R., and the International Working Group on Rotational Seismology (2009). Suggested notation conventions for rotational seismology. *Bull. Seism. Soc. Am.*, **99**(2B), 1429–1436.
- Evans, J. R., Hutt, C. R., Nigbor, R. L., and de la Torre, T. (2010). Performance of the new R2 Sensor. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Evans.pdf>, last accessed 21 January 2011].
- Farrell, W. E. (1969). A gyroscope seismometer: measurements during the Borrego earthquake. *Bull. Seism. Soc. Am.*, **59**, 1239–1245.
- Ferrari, G. (2006). Note on the historical rotation seismographs. In: Teisseyre, R., Takeo, M. and Majewski, E. (ed.), *Earthquake Source Asymmetry, Structural Media and Rotation Effects*, Springer Verlag, Heidelberg, 367–376.

- Ferreira, A., and Igel, H. (2009). Rotational motions of seismic surface waves in a laterally heterogeneous Earth. *Bull. Seism. Soc. Am.*, **99**(2B), 1073–1075.
- Fichtner, A., and Igel, H. (2009). Sensitivity densities for rotational ground-motion measurements. *Bull. Seism. Soc. Am.*, **99**(2B), 1302–1314.
- Galitzin, B. B. (1912). *Lectures on Seismometry*. St. Petersburg: Russian Academy of Sciences, (in Russian).
- Gomberg, J., and Agnew, D. (1996). The accuracy of seismic estimates of dynamic strains: an evaluation using strainmeter and seismometer data from Piñon Flat Observatory, California. *Bull. Seism. Soc. Am.*, **86**, 212–220.
- Graizer, V. M. (1991). Inertial seismometry methods. *Izv. USSR Acad. Sci., Phys. Solid Earth*, **27**(1), 51–61.
- Graizer, V. M. (2005). Effect of tilt on strong motion data processing. *Soil Dyn. Earthq. Eng.*, **25**, 197–204.
- Graizer V. (2006). Tilts in strong ground motion. *Bull. Seism. Soc. Am.*, **96**, 2090–2106.
- Graizer, V. (2009). Tutorial on measuring rotations using multipendulum systems. *Bull. Seism. Soc. Am.*, **99**(2B), 1064–1072.
- Graizer V. (2010). Strong Motion Recordings and Residual Displacements: What Are We Actually Recording in Strong Motion Seismology? *Seism. Res. Lett.*, **81**, 635–639.
- Graizer, V., and Kalkan, E. (2010). Strong motion seismology and rotations: History and future directions. Second IWGoRS Workshop, October 11–13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Graizer.pdf>, last accessed 21 January 2011].
- Hinzen, K. G. (2010). Back calculation of earthquake-rotated objects (EROs). Second IWGoRS Workshop, October 11–13, Prague, Czech Republic. [Abstract and presentation available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Hinzen.pdf>, last accessed 21 January 2011].
- Huang, B. S. (2003). Ground rotational motions of the 1991 Chi-Chi, Taiwan, earthquake as inferred from dense array observations. *Geophys. Res. Lett.*, **30**(6), 1307–1310.
- IEEE-SA Standards Board (1985). IEEE standard specification format guide and test procedure for nongyroscopic inertial angular sensors: jerk, acceleration, velocity, and displacement. *IEEE Std.*, **671-1985**, 69 pp.
- Igel, H., Schreiber, U., Flaws, A., Schubert, B., Velikoseltsev, A., and Cochard, A. (2005). Rotational motions induced by the M8.1 Tokachi-oki earthquake, September 25, 2003. *Geophys. Res. Lett.*, **32**: L08309, doi:10.1029/2004GL022336.
- Igel, H., Cochard, A., Wassermann, J., Schreiber, U., Velikoseltsev, A., and Pham, N. D. (2007). Broadband observations of rotational ground motions. *Geophys. J. Int.*, **168**, 182–197.
- Igel, H. (2010). Observations of long-period rotational ground motions: From ambient noise to Earth's free oscillations. Second IWGoRS Workshop, October 11–13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Igel.pdf>, last accessed 21 January 2011].
- Jedlička, P., Buben, J., and Kozák, J. (2009). Strong-motion fluid rotation seismograph. *Bull. Seism. Soc. Am.*, **99**(2B), 1443–1448.
- Kharin, D. A., and Simonov, L. I. (1969). VBPP seismometer for separate registration of translational motion and rotations. *Seismic Instruments*, **5**, 51–66 (in Russian).
- Knopoff, L., and Chen, Y. T. (2009). Single-couple component of far-field radiation from dynamical fractures. *Bull. Seism. Soc. Am.*, **99**(2B), 1091–1102.
- Kozák, J. T. (2009). Tutorial on earthquake rotational effects: Historical examples. *Bull. Seism. Soc. Am.*, **99**(2B), 998–1010.

- Kozák, J. (2010). Pilot sensors for rotation strong motion recording. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Kozak.pdf>, last accessed 21 January 2011].
- Kurrlle, D., Igel, H., Ferreira, A. M. G., Wassermann, J., and Schreiber, U. (2010). [Can we estimate local Love wave dispersion properties from collocated amplitude measurements of translations and rotations ?](#). *Geophys. Res. Lett.*, 37, L04307, doi:10.1029/2009GL042215.
- Langston, C.A. (2007). Wave gradiometry in two dimensions. *Bull. Seism. Soc. Am.*, **97**, 401–416.
- Langston, C. A., Lee, W. H. K., Lin, C. J., and Liu, C. C. (2009). Seismic-wave strain, rotation, and gradiometry for the 4 March 2008 TAIGER explosions. *Bull. Seism. Soc. Am.*, **99**(2B), 1287–1301.
- Lantz, B., Schofield, R., O'Reilly, B., Clark, D. E., and DeBra, D. (2009). Review: Requirements for a ground rotation sensor to improve Advanced LIGO. *Bull. Seism. Soc. Am.*, **99**(2B), 980–989.
- Lee, V. W., and Trifunac, M. D. (1985). Torsional accelerograms. *Soil Dyn. Earthq. Eng.*, **4**(3), 132–139.
- Lee, V. W., and Trifunac, M. D. (1987). Rocking strong earthquake accelerations *Soil Dyn. Earthq. Eng.*, **6**(2), 75–89.
- Lee, W. H. K. (Compiler) (2009). A glossary for rotational seismology. *Bull. Seismol. Soc. Am.*, **99**(2B), 1,082–1,090.
- Lee, W. H. K., Celebi, M., Igel, H., and Todorovska, M. I. (2009a). Introduction to the Special Issue on Rotational Seismology and Engineering Applications. *Bull. Seism. Soc. Am.*, **99**(2B), 945–957.
- Lee, W. H. K., Huang, B. S., Langston, C. A., Lin, C. J., Liu, C. C., Shin, T. C., Teng, T. L., and Wu, C. F. (2009b). Review: Progress in rotational ground-motion observations from explosions and local earthquakes in Taiwan. *Bull. Seism. Soc. Am.*, **99**(2B), 958–967.
- Lin, C. J., Huang, H. P., Liu, C. C., and Chiu, H. C. (2010). Application of Rotational Sensors to Correcting Rotation-Induced Effects on Accelerometers. *Bull. Seism. Soc. Am.*, **100**, 585 - 597.
- Lin, C. J., Liu, C. C., and Lee, W. H. K. (2009). Recording rotational and translational ground motions of two TAIGER explosions in northeastern Taiwan on 4 March 2008, *Bull. Seism. Soc. Am.*, **99**(2B), 1237–1250.
- Liu, C. C., Huang, B. S., Lee, W. H. K., and Lin, C. J. (2009). Observing rotational and translational ground motions at the HGSD station in Taiwan from 2004 to 2008. *Bull. Seism. Soc. Am.*, **99**(2B), 1228–1236.
- Mallet, R. (1862). Great Neapolitan Earthquake of 1857, I and II. Chapman and Hall, London.
- McLeod, D. P., Stedman, G. E., Webb, T. H., and Schreiber, U. (1998). Comparison of standard and ring laser rotational seismograms. *Bull. Seism. Soc. Am.*, **88**, 1495–1503.
- Nigbor, R. L. (1994). Six-degree-of-freedom ground motion measurement. *Bull. Seism. Soc. Am.*, **84**, 1665–1669.
- Nigbor, R. L., Evans, J. R., and Hutt, C. R. (2009). Laboratory and field testing of commercial rotational seismometers. *Bull. Seism. Soc. Am.*, **99**(2B), 1215–1227.
- Oldham, R. D. (1899). Report on the Great Earthquake of 12<sup>th</sup> June 1897, *Memoir of the Geological Survey of India*, **29**.
- Oliveira, C. S., and Bolt, B. A. (1989). Rotational components of surface strong ground motion. *Earthq. Eng. Struct. Dyn.*, **18**, 517–526.
- Pham, D. N., Igel, H., Wassermann, J., Cochard, A., and Schreiber, U. (2009). The effects of tilt on interferometric rotation sensors. *Bull. Seism. Soc. Am.*, **99**(2B), 1352–1365.



- Pujol, J. (2009). Tutorial on rotations in the theories of finite deformation and micropolar (Cosserat) elasticity. *Bull. Seism. Soc. Am.*, **99**(2B), 1011–1027.
- Reid, H. F. (1910). The Mechanics of the Earthquake. *The California Earthquake of April 18, 1906, Report of the State Earthquake Investigation Commission, Volume 2*. Carnegie Institution of Washington, 43–47.
- Richter, C. F. (1958). Elementary Seismology. W. H. Freeman & Co, San Francisco, CA.
- Schreiber, K. U., Igel, H., Cochard, A., Velikoseltsev, A., Flaws, A., Schuberth, B., Drewitz, W., and Müller, F. (2005). The GEOSensor project: rotations — a new observable for seismology. In: Flury, Rummel, J., R., Reigber, C., and Rothacher, M. (ed.), *Observation of the Earth System from Space*. Heidelberg: Springer Verlag, 1–19.
- Schreiber, K. U., Hautmann, J. N., Velikoseltsev, A., Wassermann, J., Igel, H., Otero, J., Vernon, F., and Wells, J.-P. R. (2009a). Ring laser measurements of ground rotations for seismology. *Bull. Seism. Soc. Am.*, **99**(2B), 1190–1198.
- Schreiber, K. U., Velikoseltsev, A., Carr, A. J., and Franco-Anaya, R. (2009b). The application of fiber optic gyroscopes for the measurement of rotations in structural engineering. *Bull. Seism. Soc. Am.*, **99**(2B), 1207–1214.
- Schreiber, K. U. (2010). Progress in ring laser technology. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Schreiber.pdf>, last accessed 21 January 2011].
- Schreiber, K. U., Velikoseltsev, A., Bosse, H., and Krause, M. (2011). The comparison of  $\mu$ FORS and R-1 rotation sensors, unpublished report.
- Segall, P. (2010). *Earthquake and Volcano Deformation*. Princeton Univ. Press, Princeton and Oxford.
- Smith, D. (2010). ATA's nanoradian-class rotational sensors. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Poster file available at: [http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Smith\\_poster.pdf](http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Smith_poster.pdf), last accessed 21 January 2011].
- Spudich, P., Steck, L. K., Hellweg, M., Fletcher, J. B., and Baker, L. M. (1995). Transient stresses at Parkfield, California, produced by the M 7.4 Landers earthquake of June 28, 1992: Observations from the UPSAR dense seismograph array. *J. Geophys. Res.*, **100**(B1), 675–690.
- Spudich, P., and Fletcher, J. B. (2008). Observation and prediction of dynamic ground strains, tilts, and torsions caused by the Mw 6.0 2004 Parkfield, California, earthquake and aftershocks, derived from UPSAR Array observations. *Bull. Seism. Soc. Am.*, **98**, 1898–1914.
- Spudich, P., and Fletcher J. B. (2009). Software for inference of dynamic ground strains and rotations and their errors from short baseline array observations of ground motions. *Bull. Seism. Soc. Am.*, **99**(2B), 1480–1482.
- Stedman, G. E. (1997). Ring laser tests of fundamental physics and geophysics. *Rept. Progress Phys.*, **60**, 615–688.
- Stupazzini, M., De La Puente, J., Smerzini, C., Kaser, M., Igel, H., and Castellani, A. (2009). Study of rotational ground motion in the near field region. *Bull. Seism. Soc. Am.*, **99**(2B), 1271–1286.
- Suryanto, W., Igel, H., Wassermann, J., Cochard, A., Schuberth, B., Vollmer, D., Scherbaum, F., Schreiber, U., and Velikoseltsev, A. (2006). First comparison of array-derived rotational ground motions with direct ring laser measurements. *Bull. Seism. Soc. Am.*, **96**, 2059–2071.
- Takamori, A., Araya, A., Otake, Y., Ishidoshiro, K., and Ando, M. (2009). Research and development status of a new rotational seismometer based on the flux pinning effect of a superconductor. *Bull. Seism. Soc. Am.*, **99**(2B), 1174–1180.

- Takeo, M. (1998). Ground rotational motions recorded in near-source region. *Geophys. Res. Lett.*, **25**(6), 789–792.
- Takeo, M. (2009). Rotational motions observed during an earthquake swarm in April, 1998, at offshore Ito, Japan. *Bull. Seism. Soc. Am.*, **99**(2B), 1457–1467.
- Takeo, M., and Ito, H. M. (1997). What can be learned from rotational motions excited by earthquakes? *Geophys. J. Int.*, **129**, 319–329.
- Teisseyre, R., Takeo, M., and Majewski, E. (editors) (2006). *Earthquake Source Asymmetry, Structural Media and Rotation Effects*. Springer-Verlag, Berlin-Heidelberg.
- Teisseyre, R., Nagahama, H., and Majewski, E. (editors) (2008). *Physics of Asymmetric Continua: Extreme and Fracture Processes: Earthquake Rotation and Soliton Waves*. Springer-Verlag, Berlin-Heidelberg.
- Teisseyre, R. (2010). Why rotation seismology: Confrontation between classic and asymmetric theories. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: [http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/R\\_Teisseyre.pdf](http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/R_Teisseyre.pdf), last accessed 21 January 2011].
- Todorovska, M. I., and Trifunac, M. D. (1990). Note on excitation of long structures by ground waves. *Am. Soc. Civil Eng., Eng. Mech. Div.*, **116**(4): 952–964.
- Todorovska, M. I., Igel, H., Trifunac, M. D., and Lee, W. H. K. (2008). Rotational earthquake motions – International working group and its activities. Proceedings of the 14<sup>th</sup> World Conference on Earthquake Engineering, October 12-17, Beijing, China, Paper ID: S03-02-0031.
- Trifunac, M. D. (1979). A note on surface strains associated with incident body waves. *Bull. European Assoc. Earthq. Eng.*, **5**, 85–95.
- Trifunac, M. D. (1982). A note on rotational components of earthquake motions on ground surface for incident body waves. *Soil Dyn. Earthq. Eng.*, **1**, 11–19.
- Trifunac, M. D. (2006). Effects of torsional and rocking excitations on the response of structures. In: Teisseyre, R., Takeo, M. and Majewski, E. (ed.), *Earthquake Source Asymmetry, Structural Media and Rotation Effects*, p. 569–582, Springer Verlag, Heidelberg.
- Trifunac, M. D. (2009). Earthquake engineering, nonlinear problems in. In: Meyers, R.A. (ed.), *Encyclopedia of Complexity and Systems Science*, p. 2421–2437, Springer, New York.
- Trnkoczy, A., P. Bormann, W. Hanka, L. G. Holcomb, and R. L. Nigbor (2002). Site Selection, Preparation and Installation of Seismic Stations. In: *New Manual of Seismological Observatory Practice (NMSOP)*, edited by Bormann, P., Chapter 7, GeoForschungsZentrum, Potsdam, Germany.
- Twiss, R. J. (2009). An asymmetric micropolar moment tensor derived from a discrete-block model for a rotating granular substructure. *Bull. Seism. Soc. Am.*, **99**(2B), 1103–1131.
- Twiss, R., Souter, B., and Unruh, J. (1993). The effect of block rotations on the global seismic moment tensor and patterns of seismic P and T axes. *J. Geophys. Res.*, **98**, 645–674.
- Velikoseltsev, A. (2010). Fiber optic gyroscope as a tool for detection of seismic rotations. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Velikoseltsev.pdf>, last accessed 21 January 2011].
- Wang, H., Igel, H., Gallovic, F., and Cochard, A. (2009). Source and basin effects of rotations: Comparison with translations. *Bull. Seism. Soc. Am.*, **99**(2B), 1162–1173.
- Wassermann, J., Lehndorfer, S., Igel, H., and Schreiber, U. (2009). Performance test of a commercial rotational motions sensor. *Bull. Seism. Soc. Am.*, **99**(2B), 1449–1456.

Wu, C. F., Lee, W. H. K., and Huang, H. C. (2009). Array deployment to observe rotational and translational ground motions along the Meishan fault, Taiwan: A progress report. *Bull. Seism. Soc. Am.*, **99**(2B), 1468–1474.



| Topic   | Recommendations for seismometer deployment and shielding                                                                                                                                                     |
|---------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Thomas Forbriger, Karlsruhe Institute of Technology (KIT) and Black Forest Observatory (BFO), Heubach 206, D - 77709 Wolfach; E-mail: <a href="mailto:Thomas.Forbriger@kit.edu">Thomas.Forbriger@kit.edu</a> |
| Version | May 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_5.4                                                                                                                                                                    |

## 1 Intention of this information sheet

This information sheet summarizes the most relevant considerations for the proper installation of broadband seismometers. An annotated example of proper shielding is provided as well as references to literature and internet resources.

There is no single solution to the installation of seismic stations. The best solution in your case will depend on the specific conditions of your observational sites and the targeted signals. Comprehensive discussions on seismometer installation and site selection are provided by Wielandt in Chapter 5 and by Trnkoczy et al. in Chapter 7 of this manual.

Temporary field installations for short period observations may put other demands on seismometer installation than permanent installations for broadband observations. For sensitive shortperiod observations you will primarily seek sites without local vibrational sources. High-resolution broadband observations on the other hand require careful thermal shielding and protection against air-pressure variations, since forces due to unwanted sensitivities to temperature and pressure decrease not as rapidly with increasing signal-period as the inertial forces due to ground motion do. For permanent installations you might be able to build a seismometer vault, while you have to make compromises regarding sites for temporary installations. Use the guidelines below when seeking the best solution for your specific case.

## 2 Location: site selection

- Prefer sites which are far away from noise sources like cars, industry, large buildings, tall poles (which can vibrate in the wind), room heaters and air-conditioners (which produce thermal variations), high-current power supply lines (which produce magnetic fields).
- Check whether you need a mains power supply, telephone or computer network connection, free view to the sky for GPS reception or solar panels. Site selection might be limited due to these requirements.
- Prefer subsurface locations (vaults, cellars, tunnels, underground shelters, etc). The overburden will provide thermal shielding. In addition this keeps the instruments at a distance to noise sources like air-pressure and passing cars which generate ground motion by surface loading.
- Avoid locations affected by air drafts. They will undermine your efforts to provide thermal shielding.
- If possible, your seismometer should be installed in a dedicated room all by itself.

- When selecting older cellars or cellars in estate buildings in particular for installation, be alarmed if non of the goods stored therein are placed on the ground directly. There is a potential risk of occasional flooding.

### 3 Protection

In particular for temporary installations, when usually compromises must be made during site selection, you should consider a few additional protection measures:

- **Protection against animals:** Free field installations should be protected against damage by animals. There are known cases of seismic installations being destroyed by cattle in stampede. For some unknown reason sheep and similar animals like to bite signal cables. Mice have been observed to occupy the subsurface cavity around buried seismometers building nests in the thermal insulation.
- **Protection against humans:** While it may appear obvious to protect the installation against vandalism (e.g. solar panels should be considered an attractive swag) also well-meaning persons might endanger the instruments. Loose cables on the floor are a trip hazard. Consider that curious fellows might like to pull at cables entering holes or boxes. Somebody might pull a power supply plug in the need of a free socket in good faith. All technical equipment (recorders, batteries, etc.), except the actual sensor, should be installed in a solid box which can serve as a means of transportation as well as a barrier to protect the installation.

### 4 Coupling

The best seismometer does not serve you well, if it is not rigidly coupled to the ground. You should consider the following:

- The feet of the instrument must have a stable position on the ground. Avoid rough surfaces or surfaces covered with dust or sand. Slightly hit the pier with a hammer to make the feet settle in a stable position. Strong-motion instruments should be bolted to the ground (but not broadband sensors).
- Ensure a direct coupling to the base rock. Do not install seismometers on a buried boulder. It can move independently from base rock and sometimes show resonances.
- Use concrete piers for permanent installations or deployments in unconsolidated sediments.
- In unconsolidated sediments you might prefer to bury the instrument, if the instrument is suitable for this.
- Avoid membranes. Badly coupled floors of cellars or plastic containers can tilt the instrument due to air-pressure variations. For long-period horizontal instruments you might like to install warp-free baseplates (see Chapter 5, section 5.5.3).
- Forces should only be exerted at the feet of the seismometer. Make sure that cables and insulation material cannot exert additional forces on the instrument. The cable should be lead around the seismometer in a big loop and fixed to the ground with additional weights.

## 5 Shielding

### 5.1 Unwanted sensitivities and protective shielding measures

Unfortunately seismometers are not only sensitive to inertial acceleration. You have to shield them against environmental influences which you do not like to observe. Items are sorted by priority:

**Thermal shielding** is a must. Typical coefficients of thermal expansion or temperature coefficients of suspension springs (even for thermally compensated alloys) are in the order of  $10^{-6} \text{ K}^{-1}$  to  $10^{-5} \text{ K}^{-1}$ . To resolve the signal of the permanently excited free oscillations at a level of  $10^{-12} \text{ g}$ , temperature variations of the sensing device must be kept below a level of  $10^{-6} \text{ K}$ . Therefore, wrap the instrument and containers around the instrument with a soft wool-like material or synthetic fleece which does not exert mechanical forces. In installations with high ambient humidity, synthetic fleece or polyester stuffing with stronger fibres should be preferred. Thin fibres can collect condensed ambient humidity which can result in a collapse of the spaces between fibres, thus increasing heat conductivity in the long run. Fill also gaps beneath the instrument. This prevents convective heat transport.

Further wrap the instrument with a reflective foil (available for first-aid kits) to prevent radiative heat transport.

Effective measures for thermal shielding are also subsurface installations, burying the instrument and anything which helps to prevent air drafts.

For temporary installations thermal shielding by covers made from Polystyrene have proven effective (pers. comm. Jörn Groos). Besides their positive thermal effect such covers are easy to install and can be used to protect the sensor mechanically during shipment and when burying the instrument.

**Shielding against air-pressure** variations is essential for long-period observations. Variations of the buoyancy force on the proof-mass of vertical component sensors are caused by variations of air-density due to meteorological air-pressure variations. The amplitude of such forces can easily exceed the amplitude of the tides, if the instrument is not shielded appropriately. The susceptibility to air-pressure for an unshielded seismometer pendulum made of brass and aluminium is approximately  $1500 \text{ nm s}^{-2} \text{ hPa}^{-1}$ . To resolve the signal of the permanently excited free oscillations at a level of  $10^{-12} \text{ g}$ , variations of air-pressure inside the instrument must be kept at a level below  $10^{-5} \text{ hPa}$ . Although broadband seismometers are supplied with an air-tight casing you should install an additional air-tight container with a warp-free design (see below) for highly sensitive installations. Otherwise the deformation caused by variation of air-pressure on the casing of the seismometer can result in tilt. Gravity coupled this way into the horizontal components easily exceeds inertial acceleration at large signal periods.

**Magnetic shielding** is required for high-resolution long-period observations. All temperature compensated suspension springs are susceptible to magnetic fields. For many broadband seismometers variations of the ambient magnetic field must be smaller than  $10^{-2} \text{ nT}$  to allow resolution of the permanently excited free modes. In particular at sites, where man-made magnetic fields due to large DC currents (e.g. tramway supply currents) have to be expected, a passive permalloy or active magnetic shield is advisable.

**Humidity** can cause corrosion and leakage currents. Therefore, keep your instruments dry. Use tight containers, plastic bags and desiccant to protect sensors and electronics. If you install the electronic components in a plastic bag, the excess heat of the electronics can put the surface temperature above the dewpoint.

**Electromagnetic interference** caused by strong fields must be avoided. Maintain a distance between signal cables and AC mains power cables. Use differential signal transmission wherever possible. Twisted pairs of signal cables improve immunity against electromagnetic fields. Avoid loops of signal cables which might act like pickup coils.

**Lightning induced overvoltage.** Long analog cables make your instruments vulnerable to lightning damage. Avoid long signal cables and consider to place your seismometer's feet on glass or perspex plates.

## 5.2 GRSN shielding: An illustrated and annotated case example

The sequence of Figures. 1 to 7 illustrates the shielding method commonly applied in the German Regional Seismic Network (Wielandt and Widmer-Schmidrig, 2002, GRSN). This type of shielding is now commercially available (Stoll, 2008). The GRSN is one of the most sensitive broadband networks and a comparably large number of its stations are able to detect the background free oscillation at a level of  $10^{-11} \text{ ms}^{-2}$  (Kurrle and Widmer-Schmidrig, 2006). The type of shielding used in this network is understood to be one of the essential factors to maintain this high quality.



**Figure 1** The seismometer (an STS-2 in this case; see DS 5.1) is placed on a polished gabbro plate (actually a raw gravestone obtained from a stonemason). The gabbro plate sits on three nuts or lead plates on the actual seismic pier in a subsurface vault. The gabbro plate is part of an air-pressure shield and serves as a practically non-deformable base plate. Electronic cables pass through a hole in the stone which is sealed with glue. The seismometer is aligned to north, the alignment rod points (in case of the STS-2) to east.



**Figure 2** Some kind of synthetic wool, polyester stuffing, or fleece material is gently wrapped around the seismometer. The gap between the gabbro plate and the casing of the instrument is also filled with fleece to prevent air convection. Be careful to keep the thermal insulation loosely fitting. The fleece must not exert forces on the seismometer casing.



**Figure 3** A stainless steel container (actually a large cooking pot) is placed upside down on a rubber gasket on the gabbro plate. It is screwed down onto the plate to provide an air-tight shielding. Air-pressure variations are thus attenuated by at least 40 dB. Pressure acting on the surface of the shielding may deform the steel container but not noticeably the gabbro plate. Deformations of the steel container do not affect the seismometer. This way a stable levelling of the horizontal components is maintained. It is advisable to place some desiccant inside the steel container to capture remaining humidity and reduce the risk of corrosion.





**Figure 4** A second layer of stuffing or fleece is wrapped around the steel container. Other than in this picture it is recommendable to wrap the gabbro plate too. The fleece again serves as a thermal insulation and prevents air drafts from acting directly on the surface of the instrument. Synthetic stuffing fleece with stronger fibres is preferable for this outer layer. The stronger fibres are able to maintain the structure of the fleece even when exposed to high humidity.



**Figure 5** The whole installation is wrapped with a reflective foil which can be obtained as a part of first-aid kits. The reflective foil reduces radiative and convective heat transport and therefore is part of the thermal insulation. Again it is recommended to extend the foil over the gabbro plate down to the surface of the pier. Ensure that no air-draft can pass underneath the foil. Tape it to the pier or put some heavy items on its edges.



**Figure 6** In this example an active magnetic shielding is added to the installation. The cubic frame supports three pairs of coils. A fluxgate magnetometer is used together with an active electronic feedback and the coils in order to maintain a constant magnetic field inside the cube. Such measures are recommended in magnetically polluted environments. In the case of this example the station (STU, Germany) is located in the city area of Stuttgart. The tramway produces large DC currents of varying magnitude in the subsurface. They induce magnetic fields which would be strong enough to limit the seismometer in its long-period seismic detection capability.



**Figure 7:** This picture shows the actual seismometer for station STU, which is associated with the GRSN ([http://www.szgrf.bgr.de/station\\_map.html](http://www.szgrf.bgr.de/station_map.html)) and GEOFON (<http://geofon.gfz-potsdam.de/geofon/>) networks.



## 6 Recommended readings

Scientific literature on technical aspects of seismometer installation is comparably rare. Uhrhammer and Karavas (1997) describe installation considerations at the Berkeley Seismological Laboratory. McMillan (2002) presents those of Albuquerque Seismological Laboratory. Havskov and Alguacil (2004) describe power supply and lightning protection issues besides general sensor installation topics. The effect of air pressure shielding is discussed by Zürn and Wielandt (2007). Wielandt and Widmer-Schmidrig (2002) present installation methods in the GRSN and Klinge et al. (2002) describe noise conditions achieved in this network. Holcomb *et al.* (1998) studied the positive effect of sand in borehole installations. A warp-free baseplate is described by Holcomb and Hutt (1992).

Studies of background noise conditions in existing networks can serve as a reference of what can be achieved in comparable environments. Section 7.2 of Chapter 7 may serve as an example, also for potential site selection. Peterson (1993) and Berger et al. (2004) provide reference levels for low-noise limits obtained in the Global Seismic Network. Widmer-Schmidrig (2003) demonstrates that the NLNM (Peterson, 1993; see Fig. 4.7 in Chapter 4) is biased by the STS-1 (see DS 5.1) self-noise at periods smaller than 1 mHz. Webb (2002) gives a quantitative reference of global noise conditions on land and on the sea floor as well as descriptions of typical unavoidable sources of seismic noise. Groos and Ritter (2009) provide an in-depth study of noise conditions in an urban environment.

To provide optimal conditions for seismic observations you require a basic understanding of non-seismic noise sources and how to avoid them (see, e.g., Chapter 4 as well as in Chapter 7 Fig. 7.5, Section 7.2 and Sections 7.4.3 to 7.3.5 with many Figures). Forbriger et al. (2010) provide a concise summary of considerations for high-resolution broadband installations. They further describe possible limitations due to magnetic field induced noise. Cavity effects which can amplify tilt and strain noise are studied by Emter and Zürn (1985) and Gebauer et al. (2010). Zürn and Widmer (1995) demonstrate the gravitational effect of the atmosphere on long-period vertical component observations and suggest potential countermeasures. Zürn et al. (2007) study tilt-noise induced on horizontal components by surface loading due to barometric pressure.

Do not hesitate to contact other network and station operators. Many of them will be willing to share their experience regarding seismometer installation (in particular if you are willing to share your data). A good point to start are the internet sites of major seismic networks. Some recommended sites are

<http://www.ldeo.columbia.edu/~ekstrom/Projects/WQC.html>

The Waveform Quality Center at Lamont-Doherty Earth Observatory provides routine waveform quality control results and noise level estimates for many stations in the Global Seismographic Network (GSN) which can serve as a reference for comparable installations.

<http://earthquake.usgs.gov/regional/asl/>

Albuquerque Seismological Laboratory provides the USGS Open File Reports on their website (McMillan, 2002, e.g.).



<http://seismo.berkeley.edu/>

UC Berkeley Seismological Laboratory operates the Berkeley Digital Seismic Network and other deployments and has published guidelines for the installation of seismometers (Uhrhammer and Karavas, 1997).

<http://ida.ucsd.edu/index.html>

The Project IDA operates a significant part of the Global Seismic Network. They feature quality control and low noise studies (Berger et al., 2004).

<http://geofon.gfz-potsdam.de/>

Part of Chapter 7 is a field report of the GEOFON network (<http://geofon.gfz-potsdam.de/geofon/manual/welcome.html>).

## Acknowledgements

Jörn Groos generously provided his experience with temporary installations of the Karlsruhe Broadband Array (KABBA). I am grateful to Erhard Wielandt, Walter Zürn, Rudolf Widmer-Schnidrig, and Peter Bormann for fruitful discussions and helpful comments.

## References

- Berger, J., Davis, P., Ekström, G. (2004). Ambient Earth noise: A survey of the Global Seismographic Network. *J. Geophys. Res.*, **109**, B11307, doi: 10.1029/2004JB003408. [http://ida.ucsd.edu/Noise\\_Study/noisestudy.html](http://ida.ucsd.edu/Noise_Study/noisestudy.html)
- Emter, D. and Zürn, W. (1985). Observation of local elastic effects on earth tide tilts and strains. In: J.C. Harrison (ed.), *Earth Tides*, Van Nostrand Reinhold Company, New York, Chapter 14, 309-327.
- Forbriger, T., Widmer-Schnidrig, R., Wielandt, E., Hayman, M., and Ackerley, N. (2010). Magnetic field background variations can limit the resolution of seismic broad-band sensors. *Geophys. J. Int.*, **181**, 303-312; doi: 10.1111/j.1365-246X.2010.04719.x.
- Gebauer, A., Steffen, H., Kroner, C., and Jahr, T. (2010). Finite element modelling of atmosphere loading effects on strain, tilt and displacement at multi-sensor stations. *Geophys. J. Int.*, **181**, 1593-1612.
- Groos, J. and Ritter, J. R. R. (2009). Time domain classification and quantification of seismic noise in an urban environment. *Geophys. J. Int.*, **179**(2), 1213-1231. doi: 10.1111/j.1365-246X.2009.04343.x.
- Havskov, J. and Alguacil, G. (2004). Instrumentation in earthquake seismology. Vol. 22 of Modern Approaches in Geophysics, *Springer*, Dordrecht, Netherlands.
- Holcomb, L. G., and Hutt, R. (1992). An evaluation of installation methods for STS-1 seismometers. *U. S. Geological Survey*, Albuquerque, NM, *Open File Report* 92-302. <http://earthquake.usgs.gov/regional/asl/pubs/>.
- Holcomb, L. G., Sandoval, L., and Hutt, B. (1998). Experimental investigations regarding the use of sand as an inhibitor of air convection in deep seismic boreholes. Open-file report 98-362, U.S. Geological Survey, Albuquerque, New Mexico. <http://earthquake.usgs.gov/regional/asl/pubs/>

- Klinge, K., Kroner, C., and Zürn, W. (2002). Broadband seismic noise at stations of the GRSN. In: M. Korn (ed.): *Ten Years of German Regional Seismic Network (GRSN)*, Wiley-VCH, Weinheim, Germany, 83-101.
- Kurrle, D. and Widmer-Schmidrig, R. (2006). Spatiotemporal features of the earth's background oscillations observed in Central Europe. *Geophys. Res. Lett.*, **33**: L24304; doi: 10.1029/2006/2006GL028429.
- McMillan, J. R. (2002). Methods of installing United States National Seismographic Network (USNSN) stations – a construction manual. *U.S. Geological Survey Open-File Report 02-144*, Albuquerque, New Mexico. <http://earthquake.usgs.gov/regional/asl/pubs/>
- Peterson, J. (1993). Observations and modeling of seismic background noise. *U.S. Geol. Survey Open-File Report 93-322*, 95 pp.; <http://earthquake.usgs.gov/regional/asl/pubs/>.
- Stoll, D. (2008). The GRSN shielding for STS-2. Lennartz electronic GmbH, <http://www.lennartz-electronic.de>
- Uhrhammer, R. A., and Karavas, W. (1997). Guidelines for installing broadband seismic instrumentation. *Technical Report*, Seismic station, University of California at Berkeley, <http://seismo.berkeley.edu/bdsn/instrumentation/guidelines.html>
- Webb, S. C. (2002). Seismic noise on land and on the sea floor. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam*, 305-318.
- Widmer-Schmidrig, R. (2003). What can superconducting gravimeters contribute to normal-mode seismology? *Bull. Seism. Soc. Am.*, **93**(3), 1370-1380.
- Wielandt, E. and Widmer-Schmidrig, R. (2002). Seismic sensing and data acquisition in the GRSN. In: M. Korn (ed.): *Ten Years of German Regional Seismic Network (GRSN)*, Wiley-VCH, Weinheim, Germany, 73-83.
- Zürn, W. and Widmer, R. (1995). On noise reduction in vertical seismic records below 2 mHz using local barometric pressure. *Geophys. Res. Lett.*, **22**(24), 3537-3540.
- Zürn, W. and Wieland, E. (2007). On the minimum of vertical seismic noise near 3 mHz. *Geophys. J. Int.*, **168**, 647-658.
- Zürn, W., Exß, J. H. S., Kroner, C., Jahr, T., and Westerhaus, M. (2007). On reduction of long-period horizontal seismic noise using local barometric pressure. *Geophys. J. Int.*, **171**, 780-796.

| Name     | CALIBRAT                                                                                                                                                               |
|----------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author   | Jens Bribach, GFZ German Research Centre for Geosciences, Telegrafenberg, 14473 Potsdam, Germany, E-mail: <a href="mailto:brib@gfz-potsdam.de">brib@gfz-potsdam.de</a> |
| Versions | October, 2001(DOS); November, 2011 (Windows);<br>DOI: 10.2312/GFZ. NMSOP-2_PD_5.1                                                                                      |

## 1 The CALIBRAT system

**Note:** All programs and sub-programs with names printed below in **bold black** letters can be downloaded via the folder name **calibrat** from the listing of *Download Programs & Files* (see NMSOP-2 item **Overview** on the cover page).

The **calibrat** system consists of three programs:

**response** calculates the response function of a complete signal chain from seismometer/geophone via preamplifier and filter stages to analog or digital recorder. This response is represented as a plot of amplitude/phase versus frequency (Bode-Diagram) or as Poles and Zeros.

**caliseis** calculates missing seismometer parameters by step response, and it designs the electronic scheme of the preamplifier stage as well as the calibration inputs to the seismometer and preamplifier.

**seisfilt** designs single and complex electronic filter stages.

The **calibrat** programs as well as their manual, source code, and examples can be downloaded from the ftp-server: [ftp.gfz-potsdam.de/pub/home/dss/brib/calibrat/](ftp://ftp.gfz-potsdam.de/pub/home/dss/brib/calibrat/) or via hyperlink by right mouse click on **calibrat** in the listing *Download programs and files* (see content overview on the NMSOP-2 cover page).

or via hyperlink from the summary listing of NMSOP-2 *Download Programs & Files*. The folders with programs and program descriptions are listed there in alphabetical order of their names.

## 2 System requirements

The DOS programs are written in Turbo Pascal, and they run on any PC of IBM™ type under DOS 3.0 or higher, also in any DOS window up to Windows XP™. An independent Windows version written in Borland Delphi is under completion.

## 3 Programs

### 3.1 RESPONSE

**response** calculates Amplitude and Phase Response  
of Seismometer-and-Filter Networks  
via Parameters such as Corner Frequency, Damping, Amplification

or via Poles and Zeros  
 or via given RC-Networks applied to Operational Amplifiers

**response** stores the results on disk  
 as Parameters  
 or as Poles and Zeros  
 or as Triples of Frequency, Amplitude and Phase

**response** plots to Screen or Printer  
 Amplitude and Phase  
 versus Period or Frequency or Angular Frequency

### 3.2 CALISEIS

**caliseis** has been developed for Seismometers/Geophones with a magnet-coil transducer.

**caliseis** calculates Seismometer and Geophone Interface (electronic amplifier interface)  
 as Preamplifier, Damping and Calibration Resistor Network  
 via given Parameters  
 or via analog Time Series of the damped Seismometer  
 for given Preamplifier Output  
 and for given Calibration Sources

**caliseis** plots to screen or to Line Printer the electronic scheme  
 of Application of Damping and Calibration Network  
 to Operational Amplifier

**caliseis** prints the electronic scheme parameters  
 as Schedule of Parameters  
 and Schedule of Network Resistors

### 3.3 SEISFILT

**seisfilt** calculates Filter Parameters  
 of single stages of first or second order filters  
 or of Butterworth LOW Passes up to 32nd order  
 or of Bessel LOW Passes up to 12th order

**seisfilt** calculates and prints  
 RC-Filter Networks  
 for Operational Amplifiers

**seisfilt** stores Filter Parameters  
 as '\*.par'

The output files '\*.par' of **seisfilt** are compatible with the input files of the program **response**.

The electronic filter circuits designed by **seisfilt** can be directly connected to the preamplifier circuit designed by **caliseis**.

## 6

## Seismic Recording Systems

Günter Asch

### 6.1 Introduction

During the last ten years, recording devices based on digital technology have completely replaced their old analog predecessors. The latter are costly, require specialized maintenance and consumables, and are incompatible with computer data processing and analysis. They are no longer produced although being still in operation at many older seismological stations and network centers. They are not dealt with in this chapter. Extensive reference to analog photographic and directly visual drum recording and processing of the records is given in the chapters Instruments and Station Operation in Willmore (1979) Manual of Seismological Observatory Practice.

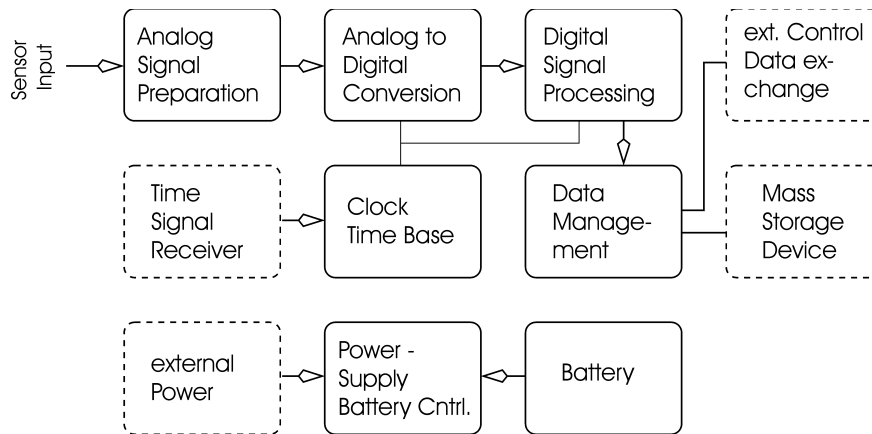
The technological progress in digital signal processing, data storage techniques and highly integrated digital circuits has lead to several instruments being available on the market that all fulfill the basic requirements of a seismic recording instrument and offer several more advanced features as well.

In terms of this chapter, a recording device is an autonomous, self-contained equipment, designed to measure the output signal of a sensor, digitize the signal and record it. In seismological experiments, all three components of ground movement are of interest, whereas in reflection experiments, only the vertical component up to now has been taken into account. Specialized multi-channel recorders with more than 6 channels, preferred in exploration seismics, are not covered here.

Seismic experiments vary from reflection profiles with short recording windows and high sampling rates to the continuous recording of broadband sensors with high resolution at observatories. An instrument well suited for an observatory, may be a bad choice for a wide-angle experiment and vice versa.

This chapter is a short introduction to the principal concept of seismic recording systems and should help the non-technical users to decide which instrument is suitable for their specific requirements. Fig. 6.1 gives an overview of the principal units of a seismic recording system. Each unit will be discussed in detail in the following sections.

## 6. Seismic Recording Systems



**Fig. 6.1** Principal units of a seismic recording device. Dashed boxes relate to optional functions.

## 6.2 Analog signal preparation

### 6.2.1 The Analog Signal Preparation section

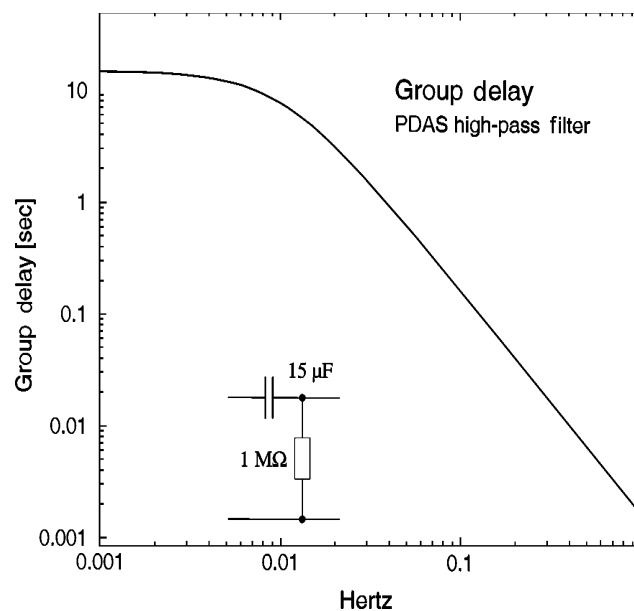
The seismic sensor is connected to the Analog Signal Preparation (ASP) section. This is not only a problem of correct wiring and polarity, but depends strongly on the type of transducer used. If a passive electrodynamic sensor is used, the impedance of the recording system influences the sensitivity and the frequency of the sensor itself. The sensor's response function (see 5.2.4 to 5.2.9 and Scherbaum, 1996) have to be corrected for the input impedance of the recording device. For this type of sensor, one has to think also about the resistivity and capacity of the cable, used to connect the sensor to the recording device, in case it exceeds several tens of meters. For active sensors, i.e., all broadband sensors, the effect of the input impedance can be neglected because it does not influence the sensor's characteristic, however long cables can introduce noise into the system. In general, short shielded cables, a single common analog ground and high quality connectors help to reduce this type of noise problems. The ASP section is also responsible for protecting the sensitive electronics of the recording system against high input electrostatic voltages.

The next step in ASP is the preamplifier which, together with the Analog to Digital Converter (ADC\*), determines the resolution [counts/Volt] of the recording device. The preamplifier has to fulfill several demands, such as linearity with respect to amplitude and phase, low noise, quick recovery from overloading, i.e., no clamping, and in addition, low power consumption. In reality, there is a tradeoff between low noise and low power consumption, and the designer of the preamplifier stage has to find a compromise. In any case, the noise generated by the ASP should be distinctly lower than the least significant bit of the ADC stage. Other requirements are not as critical, but one has to take care in a system with more than one channel, that all the parts in the ASP are identical, so each channel has the same response and sampling is done simultaneously on all channels. The technical approach is explained in section 6.3.4.

-----  
\* Terms typed in Arial letters are explained in more detail in 6.7: Glossary of technical terms and links.

### 6.2.2 Analog filters

Some recorders offer a high-pass filter to remove DC-offset and long-term drifts from the measured signal. These filters are intended to mask temperature and ageing problems, related to electronic components and to the system's specific design. Users should be able to decide whether they wish to activate these filters or not. As an example, the group delay of the optional high-pass filter from a PDAS recorder is given in Fig. 6.2. The high-pass filter is formed by a simple RC-element ( $15\ \mu\text{F} \parallel 1\ \text{M}\Omega$ ) with a time constant of about 15 seconds. Signals with periods of about 3 seconds are delayed by approximately 1 sample, assuming a sampling rate of 100 Hz. For longer periods, the situation becomes worse and it is not acceptable to activate this high-pass filter to record signals from broadband sensors. From the scientific point of view, this filter makes no sense; it simply beautifies the signal and substitutes one problem for another.



**Fig. 6.2** Group delay of the optional PDAS high-pass filter.

Before converting the analog signal into counts, it has to pass a low-pass filter serving as an anti-alias filter. This limits the frequency content of the signal toward higher frequencies. The reason for this fundamental step is described in 6.3.1. In some cases, i.e., continuously integrating converters, this stage can be an integral part of the ADC itself.

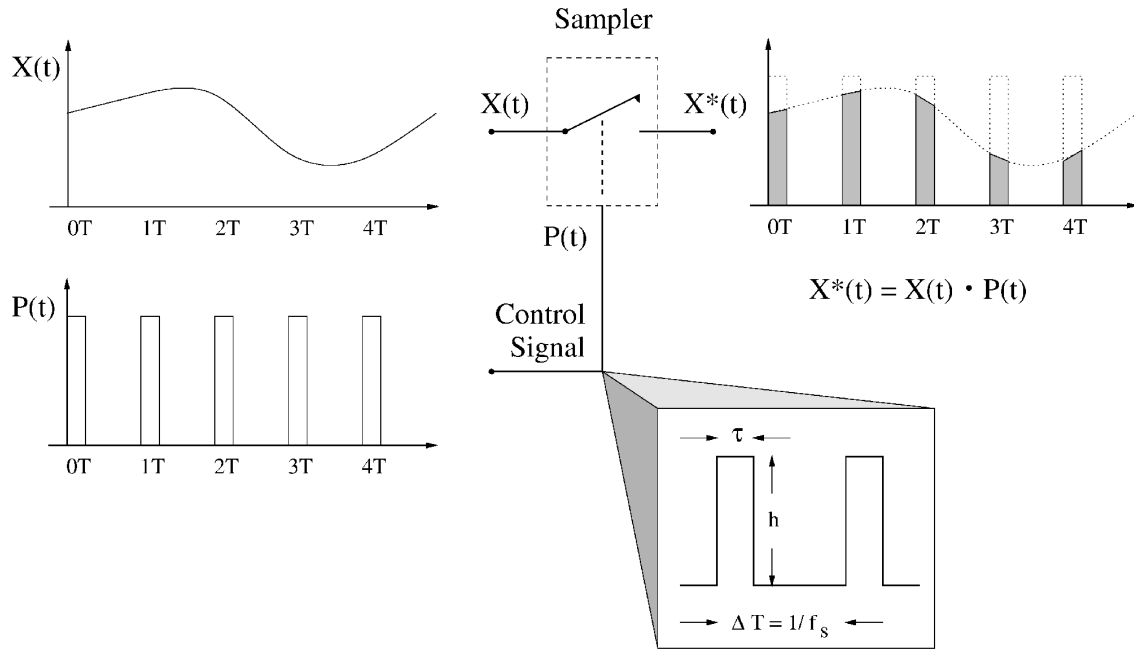
## 6.3 Analog to digital conversion

Conversion of a continuous analog signal into a digital time series is based on quantization with respect to time and amplitude. Therefore it is necessary to understand the influences and limitations introduced by these two different operations. After filtering, the analog signal is sampled and converted into digital values. Since the digital domain consists only of finite length digital words, which have to represent a continuous signal, the conversion step introduces quantization errors.

### 6.3.1 Sampling theorem

The *sampling theorem* describes the effect of quantization at discrete times on an analog signal. It was first published by Shannon (1949). A simple model is given by a switch, closed periodically for a certain time. At first, this sounds quite simple but in fact, this simple device is a modulator, performing the multiplication of two signals. Fig. 6.3 shows that the output signal is the control signal of the switch, multiplied by the input signal. The control signal is a train of impulses, therefore it is periodic and can be expressed by a Fourier series:

$$P(t) = \sum_{n=-\infty}^{\infty} h \cdot \frac{\tau}{\Delta T} \cdot \frac{\sin(\pi n f_s \tau)}{\pi n f_s \tau} \cdot e^{i2\pi n f_s t}. \quad (6.1)$$



**Fig. 6.3** Relation between input-, output- and control signal of a sample device.

The variables  $h$ ,  $\tau$ ,  $\Delta T$ , and  $f_s$  are explained in Fig. 6.3, and  $\sin(\pi f_s \tau) / \pi f_s \tau$  is the Fourier transform of the  $n^{\text{th}}$  boxcar's impulse response. If we look at the sampler as an ideal device,  $\tau$  approaches 0 and therefore  $\lim_{\tau \rightarrow 0} \sin(\pi f_s \tau) / \pi f_s \tau = 1$ , i.e., a so-called Dirac impulse. The energy of a single impulse is given by  $A = h \cdot \tau$ , so the output signal is described by

$$X^*(t) = X(t) \cdot \underbrace{\frac{A}{\Delta T} \sum_{n=-\infty}^{\infty} e^{i2\pi n f_s t}}_{P(t)} \quad (6.2)$$

Due to the periodic character of the control signal  $P(t)$ , we get an infinite number of impulse responses, separated from each other by  $\Delta T = 1/f_s$  with  $f_s$  as the sampling frequency. This changes the frequency content of the measured signal, and the spectra of the input- and output signal are no longer identical. The Fourier transform of  $X^*(t)$



$$X^*(f) = \frac{A}{\Delta T} \int_{-\infty}^{\infty} (X(t) \cdot \sum_{n=-\infty}^{\infty} e^{i2\pi n f_s t}) \cdot e^{-i2\pi f t} dt \quad (6.3)$$

results in

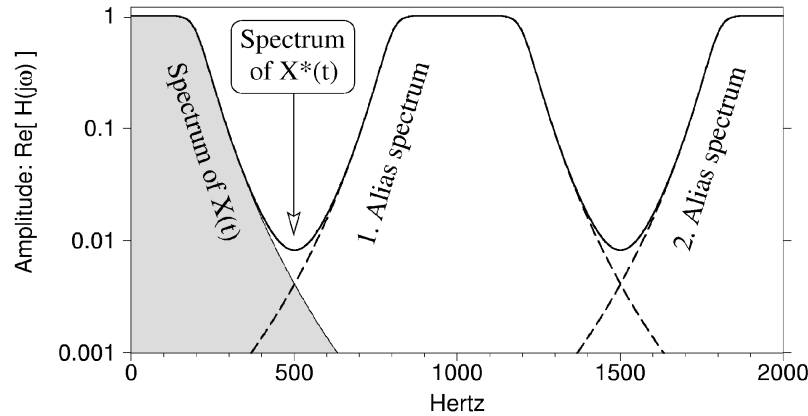
$$X^*(f) = \frac{A}{\Delta T} \sum_{n=-\infty}^{\infty} X(f - n f_s). \quad (6.4)$$

In other words, the spectrum of the input signal is transformed into a sequence of an infinite number of spectra. All these spectra, except the one of the order 0, are called the alias spectra of  $X^*(f)$ .

To illustrate this behavior, a concrete example is given in Fig. 6.4 which demonstrates the situation with the PDAS recording system. Sampling is done with a primary sampling rate of 1 kHz. The anti-alias filters are designed as Butterworth low-pass filter of an order of 6 (36 dB/octave). The corner frequency of this filter is set at 200 Hz. Its transfer function is given by

$$H(j\omega) = \frac{1}{\sqrt{1 + (j \frac{\omega}{\omega_0})^{2n}}} \quad (6.5)$$

with  $\omega_0 = 2 \cdot \pi \cdot 200$  Hz and  $n = 6$  is the order of the Butterworth filter.



**Fig. 6.4** Alias spectra in a PDAS recording system after sampling.

The consequence of aliasing in this example is that all signals with frequencies above the so-called Nyquist frequency  $f_N = f_s / 2$  are mirrored into the spectrum of  $X(t)$  (Eq. 6.2). A signal with a frequency of 600 Hz can not be distinguished from a 400 Hz signal, and a signal with 1 kHz is seen as DC with an amplitude, depending on its amplitude and the phase difference between the signal and control signal  $P(t)$ . That is why we have to assure, by applying an anti-alias low-pass filter to the analog input time series, that its high-frequency amplitudes are drastically reduced and thus do not mirror much energy into the alias spectra. Actually, the output of the sampler  $X^*(t)$  is the summing curve of all spectra, as shown in Fig. 6.4. The spectra in Fig. 6.4 were calculated on the assumption that the PDAS recording system was fed with a white noise signal.

## 6. Seismic Recording Systems

In field seismology, very little data is collected with a sampling rate of 1 kHz. In local and regional experiments, 100 Hz sampling rate is sufficient, and for teleseismic studies even 20 Hz data are commonly used. Why does the PDAS sample at such a high frequency? One answer is that the first alias spectrum is shifted far away from the frequencies in which we are interested. A second answer is given in the next section.

### 6.3.2 Oversampling

There are several ways to describe how oversampling increases resolution. The following is adopted from a technical report published by Texas Instruments (1998). The maximum error of an ideal quantizer is  $\pm 0.5$  LSB (least significant bit). Since the input range of a  $n$ -bit ADC is divided into  $2^n$  discrete levels, each represented by an  $n$ -bit binary word, the ADC input range and the word width  $n$  are a direct measure of the maximum absolute error. In addition, the quantization step can be analyzed in the frequency domain. The number of bits representing the digital value determines the signal-to-noise ratio (SNR). Therefore, by increasing the signal-to-noise ratio, the effective resolution of the conversion will be increased. Assuming the input signal is an ever-changing signal (always true in seismology), the error caused by quantization from an ideal ADC can be viewed as a white noise signal that spreads the energy uniformly across the whole bandwidth from DC to one half of the sampling rate. The RMS amplitude of the quantization error  $U_{neff}$  is given by Tietze and Schenk (1990)

$$U_{neff} = \frac{U_{LSB}}{\sqrt{12}} \quad (6.6)$$

For a full-scale sinewave signal, the RMS amplitude for an  $n$ -bit ADC is given by:

$$U_{seff} = \frac{1}{\sqrt{2}} \cdot \frac{1}{2} \cdot 2^n \cdot U_{LSB} \quad (6.7)$$

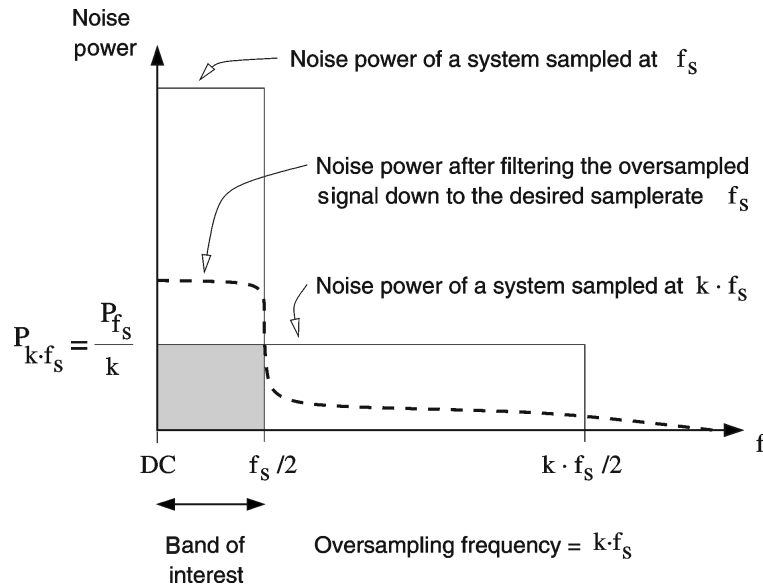
where the factor  $1/2$  is caused by the signal having  $\pm$ maximum amplitude while the converter is from  $0 - 2^n$ . The  $1/\sqrt{2}$  comes from conversion to RMS. On the basis of Eqs. (6.6) and (6.7) the signal-to-noise ratio can be calculated

$$\begin{aligned} SNR [dB] &= 20 \log_{10} \frac{U_{seff}}{U_{neff}} \\ &= 20 \log_{10} 2^n \cdot \frac{\sqrt{12}}{2 \cdot \sqrt{2}} \\ &\approx n \cdot 6 [dB] + 1.8 [dB] \end{aligned} \quad (6.8)$$

With an ideal ADC, the quantization noise power  $P = U_{neff}^2$  is uniformly distributed across the spectrum between DC and half the sampling rate. This quantization noise power is independent of the sampling rate. If we use higher sampling rates, the noise power is spread over a wider range of frequencies. Therefore the effective noise power density at the band of interest is lower at higher sampling rates. Fig. 6.5 illustrates the effective noise power reduction in the frequency band of interest at a rate of oversampling of  $k$  and the resulting sampling rate of  $k f_s$ . One has to use digital low-pass filters to remove all frequencies above  $f_s/2$ . The effective resolution is determined by the quality of the digital filter. The remaining noise power beyond  $f_s/2$  is a measure for the quantization noise and therefore responsible for a decrease in the signal-to-noise ratio.

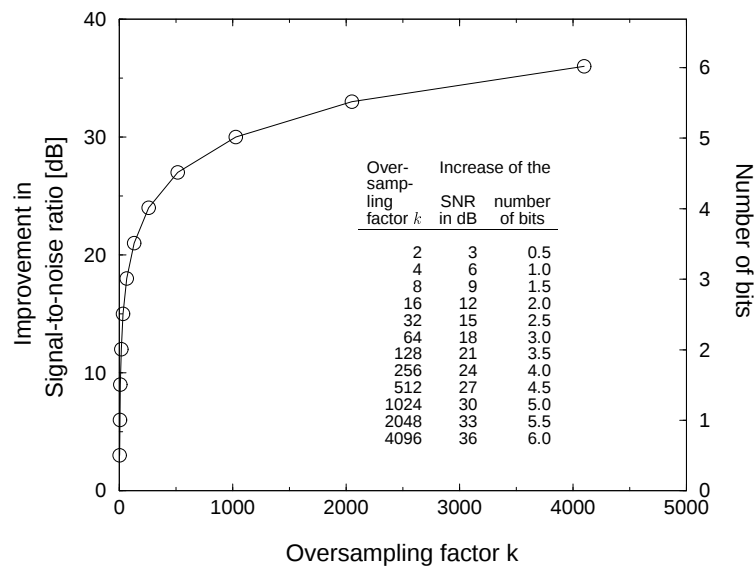
In an oversampled system, the sampling rate is decimated - after or during filtering - by a factor of  $k$ . In such a case, an ideal low-pass filter and decimator will reduce the quantization noise also by the factor of  $k$ . Since the signal at the band is not affected by the filter, this leads to an enhancement of the signal-to-noise ratio. The formula for the improved signal-to-noise ratio is:

$$SNR [dB] = 20 \log_{10} \left( \frac{U_{seff}}{U_{neff}} \cdot \sqrt{k} \right) \quad (6.9)$$



**Fig. 6.5** Noise power reduction in an oversampled system.

The increase of the signal-to-noise ratio and the corresponding increase in the number of bits is shown in Fig. 6.6.



**Fig. 6.6** Improvement of the signal-to-noise ratio with oversampling.

### 6.3.3 Digital filters

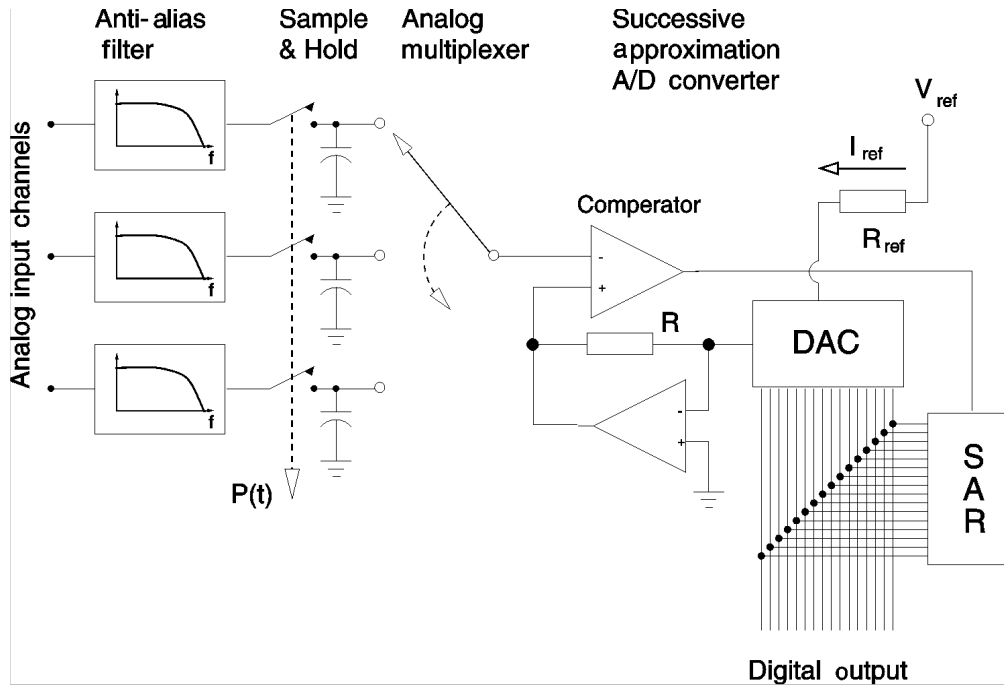
The concept of oversampling is strongly related to digital filters. In principal, there are two concepts available, filtering a digital data stream in the time domain. We distinguish between IIR and FIR filters. Both have their specific advantages and drawbacks and are discussed in detail by Buttkus (1991) and Scherbaum (1996). The IIR concept is adopted from analog filter design techniques and utilizes polynomial approximation to the desired transfer function. The name of this type of filter is given by the recursive design of the underlying algorithm. The FIR design is a non-causally approach, applying the time domain representation of the desired transfer function (Laplace transformation of the filter's impulse response; see 5.2.2) to the digital data stream. Both types have their specific benefits and disadvantages. On the one hand IIR filters are fast to compute, on the other hand they are not guaranteed to be stable with increasing order and, as their analog counterparts, introduce a phase shift depending on the frequency of the input signal. FIR filter are easy to implement, always stable and produce no phase shift (if expressed symmetrically). This results in a constant group delay and gives rise to small amplitude precursory artefacts at the output. A compromise getting the best from both approaches is the minimum-phase filter. A brief discussion about this problem and how to deal with is given by Scherbaum (see [http://lbutler.geo.uni-potsdam.de/FIR/fir\\_daaf.htm](http://lbutler.geo.uni-potsdam.de/FIR/fir_daaf.htm)).

All of the digital recorders available use FIR filters to decimate the digital data streams, and most of them are zero-phase filters. At the moment, only Earth Data and Kinematics (for the Quanterra records) offer minimum-phase decimation filters for their products.

### 6.3.4 Analog to Digital Converter (ADC)

The first generation of seismic recorders mainly utilized Successive Approximation Register ADCs (SAR). The concept is based on a DAC combined with a comparator and a shift register in a feedback loop.

It takes  $n$  steps to convert one sample with a binary resolution of  $n$ -bit. The right-hand part of Fig. 6.7 shows a block diagram of an SAR-ADC. In operation, the system enables the bits of the DAC one at a time, starting with the most significant bit (MSB). As each bit is enabled, the comparator gives an output signifying that the input signal is greater or less in amplitude than the output of the DAC. If the DAC output is greater than the input signal, the bit is reset, i.e., turned off. The system does this with the MSB first, then with the next most significant bit, etc. After  $n$ -steps, all the bits of the DAC have been tried, and the conversion cycle is completed. Fig. 6.7 shows the typical components and the signal flow for this type of ADC, as used for example, in the PDAS recorder. The anti-alias filter is followed by a Sample & Hold device. Using one common control signal  $P(t)$  for all Sample & Hold devices results in sampling at the same time of all input channels. Thus only one ADC device is necessary for multiple analog input channels, reducing power consumption and cost.

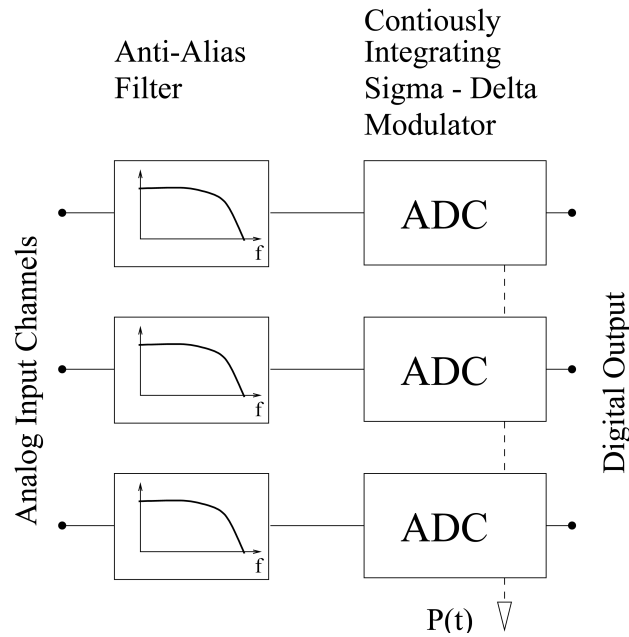


**Fig. 6.7** ADC concept used in the first generation of digital recording systems with a 16 bit converter.

This concept, however, has some severe disadvantages. During the conversion cycle, the input voltage must be kept stable below the resolution of the ADC. For this, and the parallel sampling of all input channels, the Sample & Hold unit and the analog multiplexer are introduced. But there is no ideal sampler with  $\lim_{\tau \rightarrow 0} \sin(\pi f_s \tau) / \pi f_s \tau = 1$  as required in Eq. (6.1). Also, the capacitor that holds the input voltage is not an ideal analog storage medium. All this together adds noise to the processed signal. But the most significant problem of this concept is the gap introduced between two Sample & Hold cycles, owing to the input signal being disconnected from the ADC.

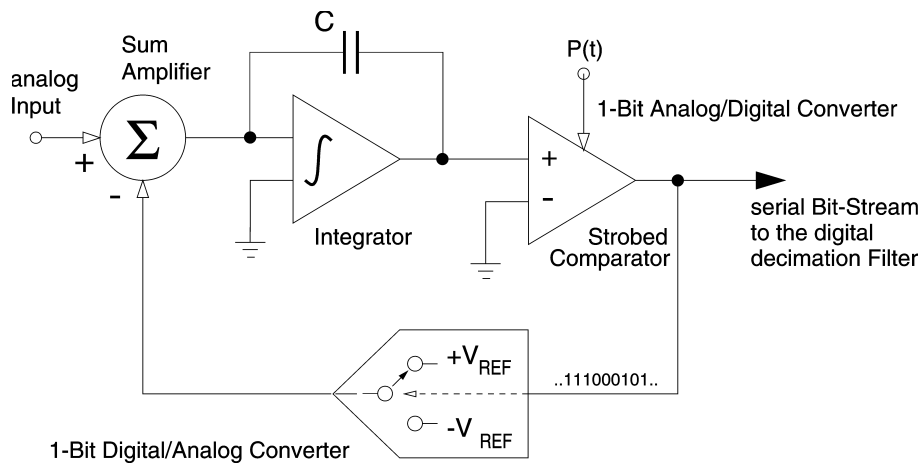
Therefore small variations of the input signal are only taken into account if they are significant with respect to the resolution of the ADC stage. The noise reduction achieved by oversampling with this type of converter is proportional to the square root of the oversampling factor, as shown in Eq. (6.9) and Fig. 6.6.

In modern data acquisition systems, there is no clear separation any more of the **Analog to Digital Conversion** and the **Digital Signal Processing** stage, as shown in Fig. 6.1. The ADC itself does a lot of digital signal processing to achieve high resolution, high signal-to-noise ratio, and high dynamics. The used techniques are based on continuous integration and oversampling in combination with carefully designed digital low-pass filters. In most modern seismic recording systems, a Delta-Sigma-Modulator is used as ADC. A simplified block diagram of the signal flow is given in Fig. 6.8. The concept of continuously integrating the analog signal avoids gaps in the sampling process and takes into account variations of the input signal even below the resolution of the ADC. The drawback is that one has to have a separate unit for each input channel, but this is more than compensated for by having abolished the Sample & Hold- and multiplexer devices.



**Fig. 6.8** ADC concept used in most modern digital recording systems.

The scheme of a Delta-Sigma-Modulator is given in Fig. 6.9. The negative feedback from the output along the 1-bit DAC and the sum amplifier at the input is performed at a high sampling rate, transferring the quantization noise into the stop band of the digital low-pass filter used for decimation.



**Fig. 6.9** First order Delta-Sigma Modulator.

The principal elements in the block diagram of the Delta-Sigma-ADC shown in Fig. 6.9 are :

- continuously sampling integrator;
- 1-Bit A/D converter (strobed comparator);
- 1-Bit D/A converter (feedback);
- sum amplifier;
- digital low-pass filter (not shown in Fig. 6.9);

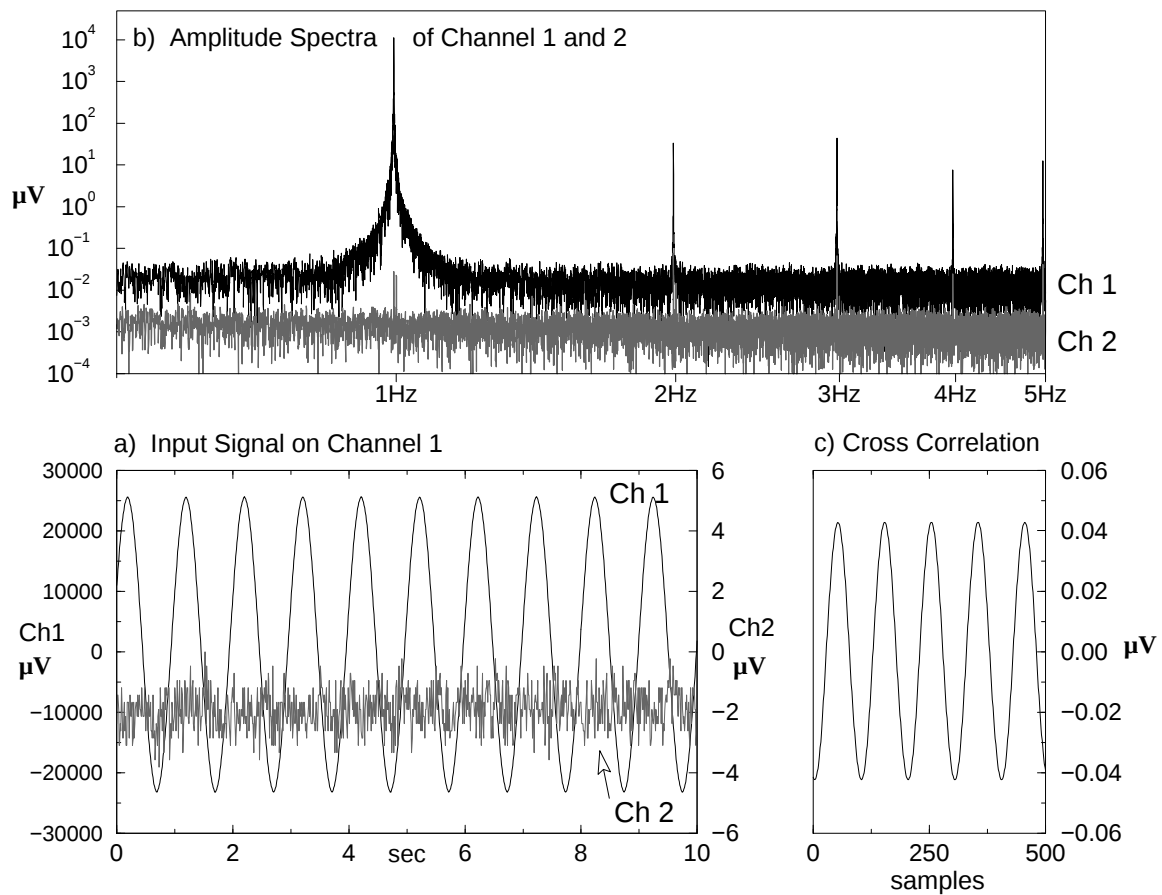
In operation, the sampled analog signal is fed to the sum amplifier, along with the output of the 1-bit DAC. The integrated difference signal is fed to the strobed comparator, whose output samples the difference signal at a frequency (the actual sampling frequency) many times that of the analog signal frequency. The output of the comparator provides the digital input for the 1-bit DAC, so the system functions as a negative feedback loop which minimizes the difference signal by tracking the input. The integrator is continuously fed with the differential signal and there are no gaps in the analog input signal, as introduced by sample and hold devices. The digital information representing the analog input voltage is coded in the polarities of the pulse train appearing at the output of the comparator. It can be retrieved as a parallel binary data word applying a digital filter operator. In general Delta-Sigma ADCs are described by the order of the integrator, which is in fact an analog low-pass filter. Fig. 6.9 shows a simple first order integrator. In real world systems, an order of four is used. This reflects a compromise between oversampling factor and stability of the modulator limited only by the performance of the analog part of the Delta-Sigma-Modulator.

### 6.3.5 Noise test

The overall noise level of a recording system can be tested by shortening the input and recording for an adequate period of time. The variation of the measured values reflects the overall noise of the recording device. This test also helps to check temperature influences and find problems in the instruments design. Loading one channel with a defined signal and cross correlating with the others, gives a detailed insight into the system's real performance.

Fig. 6.10 gives an example measured with a PDA recorder. Channel 1 was fed with a  $50\text{mV}_{pp}$  sinusoidal signal with a period of one second. Channel 2 was shortened by a  $2.5\text{k}\Omega$  resistor to simulate the output impedance of a standard passive geophone. All channels were recorded with 100 Hz sampling rate. The lower left section a) of Fig. 6.10 shows the measured signals on both channels with the left ordinate giving the  $\mu\text{Volts}$  of the input signal on channel 1 and the right ordinate the  $\mu\text{Volts}$  of the signal recorded on channel 2.

The two amplitude-spectra in Fig. 6.10 b) show a zoomed section between 0.5 to 5 Hz from a  $2^{18}$ -point FFT, calculated from a 45 minute measured record. The signals at 2, 3, 4 and 5 Hz are overtones of the signal from the signal generator. They are the result of the non-ideal sinusoidal shape of the signal generator's output, used in this test. The mean level of the spectrum of channel 1 is about one decade above the spectrum of channel 2. This shift mainly reflects the difference in resolution in the gain-ranged mode, that we used in this example. In the record of channel 1, the resolution is 30 nV/count whereas channel 2 is resolved with 4 nV/count. But one can clearly see a peak at 1 Hz in the spectrum of channel 2. Its amplitude is about 6 decades below that of the input signal on channel 1. The cross correlation between the two channels over 500 samples is shown in Fig. 6.10 c). The peak-to-peak amplitude is 0.082  $\mu\text{Volts}$ . From this, the cross coupling can be calculated as  $20 \times \log(82\text{nV}/50\text{mV}) = -115 \text{ dB}$  for this specific instrument. This is a fairly good separation of the input channels for field installations if one keeps in mind that 3-component signals measured (not only) in seismology are highly correlated.

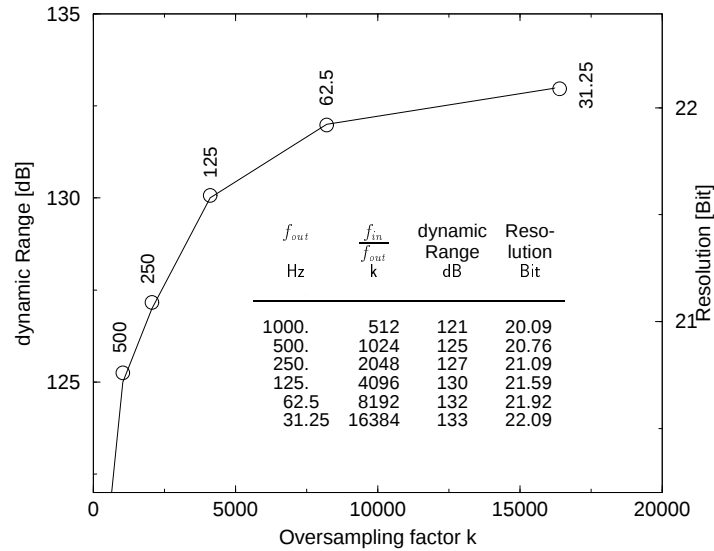


**Fig. 6.10** Signal with 50 mV<sub>pp</sub> amplitude and 1 Hz of channel 1 of a PDA recorder and the cross coupled signal on channel 2. Shown are the a) measured signals, b) their amplitude-spectra and c) their cross correlation.

### 6.3.6 The Crystal chip set

Crystal's CS5321 analog modulator and the CS5322 digital filter function together as a high resolution ADC. Therefore they are widely used in geophysical applications. The Reftek and Güralp recording systems, as well as Nanometrics's Orion are based on this chip set. The CS5322/CS5321 combination performs sampling, A/D conversion and anti-alias filtering. The CS5321 utilizes a fourth order Delta-Sigma-Modulator architecture to produce highly accurate conversions at low power dissipation (< 100 mW). It provides an oversampled serial bit stream at 128 kBit/second ( $f_{in} = 512$  kHz, 4<sup>th</sup> order oversampling architecture) to the CS5322 FIR decimation filter. From the manufacturer's data sheet, one can compile the characteristics of the CS5322/CS5321 modulator and filter combination as shown in Fig. 6.11.





**Fig. 6.11** Dynamic range of the CS5322/CS5321 chip set ( $f_{in} = 512$  kHz).

The analog input is resolved between 20 and 22 bits, depending on the oversampling ratio  $k$ . The resolution in Fig. 6.11 is limited by the modulator, not by the filter. The CS5322 provides the digital decimation filter for the CS5321 modulator output. It is not a general purpose DSP but consists of a multi-stage FIR-filter. The decimation factor by which the oversampling frequency is reduced, is selectable from  $64\times$  to  $4096\times$ . The data at the output of the digital filter is represented in a 24-bit serial format.

The -3 dB bandwidth of each decimation rate is approximately 82% of the Nyquist frequency. The filter achieves a minimum of 130 dB signal reduction at the Nyquist frequency for all filter selections. Tab. 6.1 gives an overview of the relation of the user selectable output sampling rate  $f_{out}$  with a Reftek data logger and the associated -3 dB bandwidth of the output spectra. Due to the zero-phase FIR design of the digital decimation filter CS5322, phase shift does not depend on frequency. This results in a constant factor for the group delay.

**Tab.6.1** Selectable sampling rates, -3 dB bandwidth, and associated group delay in a Reftek system.

| $f_{out}$<br>Hz | $f_{in}$<br>kHz | $\frac{f_{in}}{f_{out}}$<br>k | -3 dB<br>Point<br>Hz | Group<br>Delay<br>sec |
|-----------------|-----------------|-------------------------------|----------------------|-----------------------|
| 1000            | 512.0           | 512                           | 412.2                | 0.029                 |
| 500             | 512.0           | 1024                          | 206.0                | 0.058                 |
| 250             | 512.0           | 2048                          | 103.0                | 0.116                 |
| 125             | 512.0           | 4096                          | 51.5                 | 0.232                 |
| 200             | 409.6           | 2048                          | 82.4                 | 0.145                 |
| 100             | 409.6           | 4096                          | 41.2                 | 0.290                 |
| 50              | 409.6           | 8192                          | 20.6                 | 0.580                 |
| 25              | 409.6           | 16384                         | 10.3                 | 1.160                 |
| 40              | 327.68          | 8192                          | 16.5                 | 0.725                 |
| 20              | 327.68          | 16384                         | 8.2                  | 1.450                 |
| 10              | 81.92           | 8192                          | 4.1                  | 2.900                 |
| 5               | 81.92           | 16384                         | 2.1                  | 5.800                 |

## 6. Seismic Recording Systems

The CS5322 is realized as a cascade of three symmetrical FIR filters, which produce by design a constant delay of the output signal of  $29/f_{out}$  (see the manufacturer's data sheet for details of the design). The resulting group delay is given in the last column of Tab. 6.1. The software of the recording system has to shift the time stamp of a sample at the output of the filter stage by this amount of time, which the DSP needs to calculate the filtered output sample. But one has to take into account that the filter settles to full accuracy only after all filter stages have been completely recalculated, i.e., after two times the group delay, according to 57 output words. A transient signal like a step appears at the output of the DSP 29 words later. But to settle to full accuracy it will take another 28 steps. This example demonstrates that the field of application for Delta-Sigma converters is to track slowly changing signals, but not to record transient signals like spikes and steps.

### 6.3.7 The Quanterra family

The main impact in modern broadband observation was the availability of the Wielandt/Streckeisen (1972) sensor, the presentation of the doctoral thesis of Steim (1986), and the cooperation of Wielandt and Steim (1986). Steim introduced a 24-bit digitizer (US Patent 4866442) which even today gives outstanding performance. It is the de-facto standard in leading broadband experiments such as IRIS GSN, TERRAScope and GEOFON, just to name a few. Steim's digitizer is a variant of the Delta-Sigma-Modulator described in section 6.3.4, but is built up from discrete electronic parts and circuits. The main difference with respect to Fig. 6.9 is the use of a 16-bit ADC instead of the strobed comparator. The digital decimation filter is fed with a 16 bit data stream at 20 kHz. The feedback loop within the Delta-Sigma-Modulator is realized as a 1-bit stream, similar to the design shown in Fig. 6.9.

The main drawback of the original Quanterra data loggers is their rather high electric power consumption. By no means is it a portable instrument; it is designed for the installation in an observatory. These days Kinematics offers an array of Quanterra products for different kinds of application, including portables.

## 6.4 Time base

The usual thing to say about timing on all recording devices is that they allow synchronizing the internal time base against an external clock signal. With the availability of relatively cheap GPS receivers, the timing problem has been solved in a global sense. This type of time signal receivers provides a stable clock reference in the order of  $\mu$ seconds along with time, date and geographic position information, which allow synchronized data acquisition all over the world.

Moreover, the way a recording device synchronizes the internal data streams against the external clock signal differs, depending on the design goals. In active seismic field experiments, predefined time windows are recorded. The time slots vary from some tens of seconds in reflection - up to some tens of minutes in wide angle experiments. Also, in recent seismological experiments a triggered recording mode was mainly used due to limited storage capacities and computer facilities. In this mode, recording windows of several tens of minutes to hours are produced. Synchronizing a clock at the beginning of a recording window and keeping it running for several minutes up to hours will normally not produce significant drift errors. The situation becomes more complicated, if one has to synchronize a continuous

stream of data against an external clock signal. If the basic design is not done carefully, life can become a nightmare for people evaluating the data.

The standard procedure sounds quite simple. An internal clock signal is compared against an external time signal, normally the 1 pps (pulse per second) signal derived from a GPS receiver's output. From the difference in time, a control signal is derived and the internal clock is adjusted accordingly. Normally this is done by a phase locked loop in combination with a voltage controlled oscillator. All systems distinguish between two operation modes. If the difference is bigger than a selectable amount of time, the adjustment is done instantly. If everything works perfectly, the so-called jam set mode only occurs during the initializing phase of the system. While switching on the recorder, it updates the absolute time and date by reading the GPS input, and synchronizes the internal time base. During normal operation the time base advance and retard control provides the capability to adjust the internal oscillator without disturbing data acquisition. Adjustments in the order of  $\mu$ seconds per millisecond of the time base are allowed to smoothly synchronize the internal timing clocks with an external source or to simply adjust the internal clocks for known drift rates.

There are important differences in the technical realization of the clock control, resulting in quite different behaviors of the data acquisition systems. Selecting the start time of a recording window for a PDAS, an Orion, or a Güralp, results in digital data streams, starting exactly at the full second of the selected time. A Reftek or a Quanterra will start somewhere in the vicinity of the desired start time. The difference in design is quite small, but of great consequence. In an experiment with more than one instrument of these two types, each recorder samples at the same sampling rate, but at different times. A sample, related to midnight, is randomly distributed over two days on different recorders. In the case of a Reftek or a Quanterra, the ADC/DSP timing signal is derived from an internal oscillator, only providing the clock signal for the ADC/DSP unit and not synchronized by the external clock. Synchronizing is done at the output of the digital decimation filter. The PDAS, Orion, and Güralp recorders are so-called mono-oscillator systems. All timing signals are derived from a master oscillator, providing a 1 pps signal, which controls the timing of the whole recording system. This internal 1 pps signal is adjusted to zero phase with respect to the synchronizing external 1 pps clock signal and forces synchronizing of the ADC, i.e., at the input of the decimation filter. With this timing concept, all instruments in an array or on a profile, take their samples at the same time, starting at full seconds. This simplifies data acquisition of continuous data streams to a great extent.

## 6.5 Data management

### 6.5.1 Storage media

In field experiments with continuously recording instruments, one collects about 30-50 MBytes of data per day at a 3-component station with 100 Hz sampling rate. During the last years, hard disks have overcome other mass storage technologies in field recording. They have become reliable, are quite cheap, robust if switched off, their power consumption is moderate, and access is much faster than on other mass storage media. The big advantage of hard disks against other magnetic- and opto-magnetic media is the fact that they are encapsulated and insensitive in rough and dusty environments. For connecting the hard disk to the recorder, many systems support SCSI. This bus system is an accepted and well defined

## 6. Seismic Recording Systems

industry standard, available on most platforms. Up to 7 devices can be connected in parallel on a 8-bit SCSI bus and the technical specification even allows hot pluggable devices, i.e., the operator can change the hard-disc without interrupting continuous recording. With an Orion or a Güralp system, it is possible to change the hard disk without interrupting the recording process. On a PDAS this will not only result in a corrupted file system, but one also has a good chance to kill the SCSI controller chip. The only problem with SCSI hard-discs is that they are developing more and more to the high performance server market. In this market segment, low-power (and low price) is not a basic design goal. All the hard-discs fulfilling the requirements for seismic recording systems, were designed for laptops, where IDE is the standard system interface. But to swap IDE devices in a running computer system for reading or writing, ranges from tricky to impossible. So we are still waiting for the manufacturer of seismological recording systems providing us with a hot pluggable mass storage device based on low-power IDE drives and equipped with a TCP/IP connection, which would allow access to a practically unlimited number of storage media at the same time, with all the benefits of a network connection.

### 6.5.2 Data formats, compression and metadata

Even a hard disk, designed for random access, can be used as a sequential block device, just like a tape drive. The data stream is subdivided into blocks of a fixed length, including the header which defines the type of data and holds the date and time stamp of the first sample. Besides the raw data streams, state of health and status information are generated. Additional channels containing information about internal voltages, temperatures, and information related to the synchronization against the external clock are recorded.

Each subsequent block of data is written to the hard disk. No file or directory structure is necessary and it is quite easy to implement compression algorithms or ring buffer structures. Steim (1986) introduced a widely used algorithm to compress integer time series without loss of information. Blocks of data are organized in frames with a fixed length. Only the difference with respect to the first sample in the frame is stored. An associated table, occupying a 2-bit/sample within the frame, holds the information about the significant word length of the stored differential value (8-, 16- or 32-bit). The compressional rate depends on the difference between two subsequent samples, or in terms of seismology, on the noise situation of the measured signal. In general the compression rate achieved with this so-called STEIM1 algorithm is in-between 1/3 and 2/3. Steim extended this scheme introducing 4 bit differences in the STEIM2 algorithm and gaining a compression rate up to 30% against STEIM1. The Quanterra, the Orion, and the Reftek compress their data with the STEIM1 algorithm, whereas Güralp has its own compression method, but there is no principal difference.

If the block header holds two additional pointers to its predecessor and successor, this is called a tagged file system. In fact, it allows only sequential access to the data of a specific stream. Some systems like the Orion reserve a certain amount of disk space for each raw data stream. This saves the two pointers and in addition, allows an easy implementation of ring buffers. After a ring buffer is filled, the oldest data are subsequently overwritten. This user selectable behavior utilizes continuous data acquisition with a fixed length of time history. If an event in which one is interested occurs, the length of the ring buffer gives the time in which one has to access the station and download the data. Sampling 3 channels at a rate of 100 Hz, a 4 GByte hard disc will provide recording capacity for about 80 days.

To read the data into a computer system, one has only to be able to access the physical raw data device and to know the logical structure of the data frames. The big disadvantage of the frame structure with compressed data is that there is no possibility to address an item directly as recorded at a specific time. One has to read all subsequent block headers from the beginning until finally accessing the data. This storage and compression concept in seismic recording systems is optimized with respect to disk space. This is the best solution for acquiring but a bad choice for processing the data. So the first step after reading the data into a computer system, is to convert and organize it within a file system. There are several different operating systems around, and each operates with several different versions or types of file systems. If a recording system relies on a specific file system, this can become a problem for the user, who has to organize the data on a different operating system.

File systems introduce an additional abstract layer to the data model and also have their specific limitations, starting with the number of characters in a file name, up to the maximum length of a single file. The PDAS is the only recorder mentioned here writing MS-DOS<sup>TM</sup> data files directly and organizing the data streams in a hierarchical directory structure. The Orion also utilizes the same file system. If the host system is able to mount an MS-DOS<sup>TM</sup> file system, the ring buffers are visible as ordinary files.

The conclusion of all the different data formats on different recording systems is that each one needs its special treatment. The only common level at the moment is the SCSI hardware, but even this will change in future. All systems have their own way to store the data logically. Even if Steim's compression algorithm is used to write the raw data, the block header structures may differ. But this is not the real annoyance. There is no common agreement on what type of additional information is or should be recorded by the system.

Each manufacturer has his own ideas, how to synchronize against the external clock signal and how to correct the drift of the time base, and of course, how to report this. So, from here on, no general recipe for converting raw data to the user's file system can be given. Also, no recipe is available on how to treat timing errors from different data loggers (there are systems around, even reporting *unknown error*).

The user's file system normally depends on the software being run to process the data. This is the scientific part of the game and may vary from project to project. To buy an instrument from a manufacturer who is supporting only one specific operating system and withholding information about the low-level data structures, may be a bad investment.

Unfortunately, the approach to unify the low level data formats for seismic recording systems, the SUDS-2 format initialized by Ward (1992), failed because of the complexity of the abstract data model and the non-agreement on a common platform for Unix and PC-based systems. Thus, the problem handling the metadata is still unsolved. All the data describing the instrumentation and the site location related to the stored waveform are called metadata. For archiving and exchange of waveform data, there are several data formats in use, mainly, the different variants of SEG-Y and GCF (pure ASCII code), which are independent of the hardware platform and even email-proof on 7-Bit mail servers. But the most complete and widely accepted standard is the SEED format and its M-SEED variant. SEED is the only standard holding the most important metadata and the waveform data within one single file, a so-called SEED volume. M-SEED is a subset of the SEED definition, holding the waveform-data only. Thanks to the excellent software library QLIB2, M-SEED has become an easy-to-use and platform-independent data format.

## 6. Seismic Recording Systems

The software packages provided by the manufacturers are mainly reduced to set-up, quick lock, and quality check functions and are normally not intended for scientific evaluation of the data. The software should also provide converters to more common formats mentioned above. Here the GEOFON, the ORFEUS, and the PASSCAL home pages are ideal starting points to find all kinds of useful software and information.

### 6.6 Conclusions and final remarks

There is no "best" recording system available. It always depends on several aspects, which instrument to choose - and the weights are different from project to project. The requirements for a temporary installed, portable network are completely different from a permanent setup in an observatory, connected to the power supply system and the Internet. Beyond this, Delta-Sigma-Modulators with 130-140 dB and GPS-timing are state of the art, Web-hosted communication via Internet is coming up, and prices, hopefully, down. Within this chapter several products from various manufacturers are named. They are only used as examples to show specific differences in the technical realization of seismic recording systems. In fact, there are more products available but this chapter has no intention to give an overview of the market. If one particular product is not mentioned here, it is not an opinion in terms of its quality.

### 6.7 Glossary of technical terms and links

**ADC** : Analog to Digital Converter. A device that converts data from analog to digital form.

**CS5321/CS5322** : A 24-bit word-length variable bandwidth ADC. The CS5321 is a Sigma-Delta-Modulator which functions together with the CS5322 digital filter as a high resolution ADC. The CS5322/CS5321 combination performs sampling, AD conversion, and anti-alias filtering. The circuits are manufactured by *Crystal* (<http://www.crystal.com/>).

**DAC** : Digital to Analog Converter. A device which takes a digital value and outputs a voltage which is proportional to the input value.

**DSP** : Digital Signal Processor. The *big four* programmable DSP chip manufacturers are Texas Instruments, with the TMS320 series of chips; Motorola, with the DSP56000, DSP56100, DSP56300, DSP56600, and DSP96000 series; Lucent Technologies (formerly AT&T), with the DSP1600 and DSP3200 series; and Analog Devices, with the ADSP2100 and ADSP21000 series. For further information have a look at [www.bdti.com/pocket/dsp\\_guide.htm](http://www.bdti.com/pocket/dsp_guide.htm).

**Earth Data** : Earth Data Limited specializes in the research, design and manufacture of data acquisition, telemetry and 24-Bit portable recording systems ([www.kenda.co.uk/edata](http://www.kenda.co.uk/edata)) .

**FIR** : Finite Impulse Response filter, also named acausal- or, in their symmetrical realization, zero-phase-filters. For a brief description of digital filters (see Scherbaum,1996; and <http://lbutler.geo.uni-potsdam.de/service.htm>).

**Gain-ranged Mode**: A Gain Ranging Amplifier is scaling the input signal to fit the ADC's digitizing window based on signal level. This method increases the dynamic range, but not the resolution of a recording system. The resolution is limited to something less than

the number of bits available from the ADC, due to noise introduced by the uncertainties of the different gain steps. Gain-ranged mode was used in older 16-bit ADC based systems to achieve dynamic ranges up to 130 dB.

**GEOFON** : The GEOFON program of the GeoForschungsZentrum Potsdam presently operates, together with its partner organizations, 40 permanent and a varying number of longterm temporary broadband seismological stations. 30 permanent stations are located in Europe and the Mediterranean. Most of these stations are equipped with Streckeisen STS-2 very broad band seismometers and adequate Quanterra Q380 or Q4120 dataloggers (see [www.gfz-potsdam.de/geofon/index.html](http://www.gfz-potsdam.de/geofon/index.html)).

**GPS** : Global Positioning System. There are a lot of good sites to start from, for example [www.skydiversdepot.com/gps2.htm](http://www.skydiversdepot.com/gps2.htm).

**Güralp** : Güralp Systems Ltd (GSL). Manufacturer of seismometers and data acquisition systems ([www.guralp.demon.co.uk](http://www.guralp.demon.co.uk)).

**GSN** : The IRIS Global Seismographic Network. The goal of the GSN is to deploy 128 permanent seismic recording stations uniformly over the Earth's surface. These stations continuously record seismic data from very broad band seismometers at 20 samples per second. It is also the goal of the GSN to record data with a dynamic range of 140 db (24 bit digitizers). ([www.iris.washington.edu/GSN/index1.htm](http://www.iris.washington.edu/GSN/index1.htm))

**IDE** : Integrated Drive Electronics. Originally called IDE, the ATA interface was invented by Compaq around 1986. Standardized by the ANSI group X3T10 (who named it *Advanced Technology Attachment* (ATA)). Ratification in November 1990.

**IIR** : Infinite Impulse Response filter, also named causal-filter. All analog filters are of this type. For brief discription of digital filters see Scherbaum (1996).

**IRIS** : Incorporated Research Institutions for Seismology. 1200 New York Ave. NW, Suite 800, Washington, DC 20005 ([www.iris.edu](http://www.iris.edu)).

**Kinemetrics** : Kinemetrics Inc. Manufacturer of seismic sensors and data acquisition systems. (see [www.kinemetrics.com](http://www.kinemetrics.com)).

**Lennartz** : Lennartz electronic GmbH. Manufacturer of seismic sensors and data acquisition systems (see [www.lennartz-electronic.de](http://www.lennartz-electronic.de)).

**MARSLite** : The MARS recorders are an array of portable data loggers developed and manufactured by Lennartz electronic GmbH. Recent products are the MARSLite and M24 data loggers.

**Nanometrics** : Manufacturer of data acquisition systems. Nanometrics also offers complete network solutions based on satellite communication (see [www.nanometrics.ca](http://www.nanometrics.ca)).

**ORFEUS** : Observatories and Research Facilities for European Seismology. ORFEUS is the European non-profit organization that aims at coordinating and promoting digital, broadband seismology in Europe (see [orfeus.knmi.nl](http://orfeus.knmi.nl)).

**ORION** : Digital recording system, manufactured by Nanometrics.

**PASSCAL** : Program for the Array Seismic Studies of the Continental Lithosphere. PASSCAL Instrument Center, New Mexico Tech, 100 East Road, Socorro, NM 87801, U.S.A (see [www.passcal.nmt.edu/passcal/resources.htm](http://www.passcal.nmt.edu/passcal/resources.htm)).

## 6. Seismic Recording Systems

**PDAS** : Portable Data Acquisition System. Solid first generation digital recording system manufactured by Geotech/Teledyne, 3401 Shiloh Road, Garland, Texas 75041 (see [www.geoinstr.com](http://www.geoinstr.com)). This data-logger is no longer available on the market, but still widely used.

**QLIB2** : Software library available as source code (C/Fortran) from Doug Neuhauser ([doug@seismo.berkeley.edu](mailto:doug@seismo.berkeley.edu)) to read and write MSEED data files on different hardware platforms. Jewel found in the PASSCAL software distribution (Ver. 1.9) under contrib/mseed/qlib2. Unfortunately the PASSCAL routines do not profit from this library, because they used their own.

**Quanterra** : Advanced broad band remote data acquisition system. Developed by Stein (1986), it is the defacto standard in global broadband seismology. These days Kinemetrics is selling an array of Quanterra products with different specifications.

**Reftek** : Refraction Technology, Inc. 2626 Lombardy Lane, Suite 105, Dallas, TX 75220, U.S.A. (see [www.reftek.com](http://www.reftek.com)).

**SAR-ADC** : Successive Approximation Analog to Digital Converter.

**SCSI** : Small Computer System Interface. Bidirectional, parallel interface to connect up to 7 (15 with wide SCSI) external devices to a computer. ANSI-X-3T9.2 *American National Standards Institute* (see [www.ieee.org/index.html](http://www.ieee.org/index.html)).

**TCP/IP** : Transmission Control Protocol over Internet Protocol. The de facto standard Ethernet protocols. TCP/IP was developed by DARPA for internet working and encompasses both network layer and transport layer protocols. While TCP and refer to the entire DoD protocol suite based upon these, including telnet, FTP, UDP and RDP.

**TERRAscope** : is a very broadband seismographic network in Southern California. Each station consists of a Wielandt/Streckeisen seismograph, a strong-motion sensor, and a barograph. The data is recorded on-site using the 24-bit Quanterra datalogging system and collected by the Caltech Seismo Lab. ([www.gps.caltech.edu/terrascope/TerraInfo.html](http://www.gps.caltech.edu/terrascope/TerraInfo.html)).

## Acknowledgments

Reviews and suggestions by P. Bormann, J. Bribach, J. Havskov and K.-H. Jäckel have been taken into account to complement the initial draft of this paper.

## Recommended overview readings (see References under Miscellaneous in Volume 2)

Buttkus (1991)  
Havskov and Alguacil (2004)  
Scherbaum (1997)  
Scherbaum (2001)  
Shannon (1949)  
Ward (1992).  
Willmore (1979)



# Chapter 7

## Site Selection, Preparation and Installation of Seismic Stations

(Amended version, September 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch7)

Amadej Trnkoczy<sup>1)</sup>, Peter Bormann<sup>2)</sup>, Winfried Hanka<sup>2)</sup>,  
L. Gary Holcomb<sup>3)</sup>, Robert L. Nigbor<sup>4)</sup>, Masanao Shinohara<sup>5)</sup>,  
Hajime Shiobara<sup>5)</sup> and Kiyoshi Suyehiro<sup>6)</sup>

<sup>1)</sup> Formerly Kinematics; Bovec, Slovenia; E-mail: amadej.trnkoczy@siol.net;

<sup>2)</sup> GFZ German Research Center for Geosciences, Telegrafenberg, 14473 Potsdam, Germany;  
E-mail: [pb65@gmx.net](mailto:pb65@gmx.net) and [hanka@gfz-potsdam.de](mailto:hanka@gfz-potsdam.de);

<sup>3)</sup> Formerly U.S. Geological Survey, Albuquerque Seismological Laboratory; now: 3723 Espejo NE,  
Albuquerque, NM 87111-3430, United States; E-mail: [gary.holcomb@huskers.unl.edu](mailto:gary.holcomb@huskers.unl.edu);

<sup>4)</sup> Department of Civil Engineering, University of California at Los Angeles, Los Angeles, CA 90095, USA;  
E-mail: [nigbor@ucla.edu](mailto:nigbor@ucla.edu);

<sup>5)</sup> Earthquake Research Institute, the University of Tokyo, 1-1-1, Yayoi, Bunkyo-ku, Tokyo 113-0032, Japan,  
E-mail: [mshino@eri.u-tokyo.ac.jp](mailto:mshino@eri.u-tokyo.ac.jp) and [shio@eri.u-tokyo.ac.jp](mailto:shio@eri.u-tokyo.ac.jp);

<sup>6)</sup> Integrated Ocean Drilling Program Management International, Inc., Tokyo University of Marine Science and  
Technology Office of Liaison and Cooperative Research, 3rd Floor 2-1-6, Etchujima, Koto-ku, Tokyo 135-  
8533 Japan, E-mail: [ksuyehiro@iodp.org](mailto:ksuyehiro@iodp.org)

|                                                                                | <b>page</b> |
|--------------------------------------------------------------------------------|-------------|
| <b>7.1 Factors affecting seismic site quality and site selection procedure</b> | <b>5</b>    |
| <b>(A. Trnkoczy; version 2002)</b>                                             |             |
| 7.1.1 Introduction                                                             | 5           |
| 7.1.2 Offsite studies                                                          | 5           |
| 7.1.2.1 Definition of the geographic region of interest                        | 6           |
| 7.1.2.2 Seismo-geological considerations                                       | 7           |
| 7.1.2.3 Topographical considerations                                           | 8           |
| 7.1.2.4 Station access considerations                                          | 9           |
| 7.1.2.5 Evaluation of seismic noise sources                                    | 9           |
| 7.1.2.6 Seismic data transmission and power considerations                     | 12          |
| 7.1.2.7 Land ownership and future land use                                     | 12          |
| 7.1.2.8 Climatic considerations                                                | 12          |
| 7.1.3 Field studies                                                            | 13          |
| 7.1.3.1 Station access verification                                            | 14          |
| 7.1.3.2 Local seismic noise sources and seismic noise measurements             | 14          |
| 7.1.3.3 Field study of seismo-geological conditions                            | 16          |
| 7.1.3.4 Field survey of radio frequency (RF) conditions                        | 16          |
| 7.1.3.5 Shallow seismic profiling                                              | 16          |
| 7.1.4 Using computer models to determine network layout capabilities           | 17          |
| <b>7.2 Investigation of noise and signal conditions at potential sites</b>     | <b>19</b>   |
| <b>(P. Bormann; version 2002)</b>                                              |             |
| 7.2.1 Introduction                                                             | 19          |
| 7.2.2 Reconnaissance noise studies prior to station site selection             | 20          |
| 7.2.2.1 Offsite assessment of expected noise levels and measurement            |             |

|            |                                                                                                                                                   |           |
|------------|---------------------------------------------------------------------------------------------------------------------------------------------------|-----------|
|            | of instrumental self-noise                                                                                                                        | 20        |
|            | 7.2.2.2 Sensor installation, measurements and logbook entries in the field                                                                        | 22        |
|            | 7.2.2.3 Case study of noise records in the frequency range $0.3 \text{ Hz} < f < 50 \text{ Hz}$                                                   | 24        |
| 7.2.3      | Comparison of noise and signals at permanent seismological stations                                                                               | 29        |
|            | 7.2.3.1 Introduction                                                                                                                              | 29        |
|            | 7.2.3.2 Data analysis                                                                                                                             | 30        |
|            | 7.2.3.3 Results                                                                                                                                   | 31        |
| 7.2.4      | Searching for alternative sites in a given network                                                                                                | 35        |
|            | 7.2.4.1 Geological and infrastructure considerations                                                                                              | 35        |
|            | 7.2.4.2 Recording conditions and data analysis of temporary noise measurements for alternative permanent broadband stations                       | 36        |
|            | 7.2.4.3 Results of noise and signal measurements at BRNL and RUE                                                                                  | 37        |
|            | 7.2.4.4 Results of noise and signal measurements at HAM and BSEG                                                                                  | 39        |
|            | 7.2.4.5 Causes of spectral noise reduction at RUE and BSEG and conclusions                                                                        | 42        |
| <b>7.3</b> | <b>Data transmission by radio-link and RF survey (A. Trnkoczy; version 2002)</b>                                                                  | <b>43</b> |
|            | 7.3.1 Introduction                                                                                                                                | 43        |
|            | 7.3.2 Types of RF data transmission used in seismology                                                                                            | 43        |
|            | 7.3.3 The need for a professional radio frequency (RF) survey                                                                                     | 45        |
|            | 7.3.4 Benefits of a professional RF survey                                                                                                        | 46        |
|            | 7.3.5 Radio-frequency (RF) survey procedure                                                                                                       | 47        |
|            | 7.3.6 The problem of radio-frequency interference                                                                                                 | 49        |
| <b>7.4</b> | <b>Seismic station site preparation, instrument installation and shielding</b>                                                                    | <b>50</b> |
|            | 7.4.1 Introduction and general requirements (A. Trnkoczy; version 2002)                                                                           | 50        |
|            | 7.4.2 Vault-type seismic stations (A. Trnkoczy; version 2002)                                                                                     | 51        |
|            | 7.4.2.1 Controlling environmental conditions                                                                                                      | 52        |
|            | • Mitigating temperature changes                                                                                                                  | 53        |
|            | • Thermal tilt mitigation                                                                                                                         | 58        |
|            | • Lightning protection                                                                                                                            | 58        |
|            | • Electro-Magnetic Interference protection                                                                                                        | 59        |
|            | • Water protection                                                                                                                                | 60        |
|            | • Protection from small animals                                                                                                                   | 61        |
|            | 7.4.2.2 Contact with bedrock                                                                                                                      | 61        |
|            | 7.4.2.3 Seismic soil-structure interaction and wind-generated noise                                                                               | 62        |
|            | 7.4.2.4 Other noise sources                                                                                                                       | 63        |
|            | 7.4.2.5 Electrical grounding                                                                                                                      | 64        |
|            | 7.4.2.6 Vault construction                                                                                                                        | 65        |
|            | 7.4.2.7 Miscellaneous Hints                                                                                                                       | 66        |
|            | • Vault cover design                                                                                                                              | 66        |
|            | • Alternative materials                                                                                                                           | 66        |
|            | • Mitigating vandalism                                                                                                                            | 66        |
|            | • Fixing seismometers to the ground                                                                                                               | 66        |
| 7.4.3      | Seismic installations in tunnels and mines (G. Holcomb; version 2002)                                                                             | 67        |
| 7.4.4      | Parameters which influence the very long-period performance of a seismological station: examples from the GEOFON Network (W. Hanka; version 2002) | 68        |
|            | 7.4.4.1 Introduction                                                                                                                              | 68        |
|            | 7.4.4.2 Comparison of instrumentation and installation                                                                                            | 69        |

|                                                                                               |     |
|-----------------------------------------------------------------------------------------------|-----|
| • Which seismometer to choose?                                                                | 69  |
| • Installation of an STS1/VBB                                                                 | 70  |
| • Installation of an STS2                                                                     | 71  |
| 7.4.4.3 Comparison of vault constructions, depth of burial,<br>geology and climate            | 72  |
| • Tunnel vaults                                                                               | 73  |
| • Shallow vaults                                                                              | 74  |
| • Surface vaults in moderate climate                                                          | 76  |
| • Surface vaults in arctic climate                                                            | 77  |
| 7.4.4.4 Conclusions                                                                           | 77  |
| 7.4.5 Broadband seismic installations in boreholes ( <b>L. G. Holcomb;<br/>version 2002</b> ) | 79  |
| 7.4.5.1 Introduction                                                                          | 79  |
| 7.4.5.2 Noise attenuation with depth                                                          | 80  |
| 7.4.5.3 Site selection criteria                                                               | 81  |
| 7.4.5.4 Contracting                                                                           | 82  |
| 7.4.5.5 Suggested borehole specifications                                                     | 82  |
| 7.4.5.6 Instrument installation techniques                                                    | 84  |
| 7.4.5.7 Typical borehole parameters                                                           | 86  |
| 7.4.5.8 Commercial sources of borehole instruments                                            | 87  |
| 7.4.5.9 Instrument noise                                                                      | 88  |
| 7.4.5.10 Organizations with known noteworthy borehole experience                              | 89  |
| 7.4.6 Borehole strong-motion array installation ( <b>R. L. Nigbor; version 2002</b> )         | 90  |
| 7.4.6.1 Introduction                                                                          | 91  |
| 7.4.6.2 Borehole array planning                                                               | 93  |
| • Location                                                                                    | 93  |
| • Geologic implications                                                                       | 94  |
| • Coupling and retrievability issues                                                          | 95  |
| • Sensor orientation                                                                          | 96  |
| • Systems issues                                                                              | 97  |
| 7.4.6.3 Borehole preparation                                                                  | 97  |
| • Planning                                                                                    | 97  |
| • Selection of drilling contractor                                                            | 99  |
| • Permits                                                                                     | 99  |
| • Drilling                                                                                    | 100 |
| • Geotechnical sampling                                                                       | 101 |
| • Casing                                                                                      | 103 |
| • Grouting                                                                                    | 104 |
| 7.4.6.4 Geotechnical/Geophysical measurements                                                 | 104 |
| • Literature search                                                                           | 105 |
| • Pre-installation geophysical studies                                                        | 106 |
| • Lithology logging                                                                           | 106 |
| • Laboratory testing of soil samples                                                          | 106 |
| • Borehole geophysical measurements                                                           | 107 |
| 7.4.6.5 Installation procedure                                                                | 109 |
| • Sensor installation                                                                         | 109 |
| • Orientation                                                                                 | 110 |
| • Operational checkout                                                                        | 110 |
| • Evaluation period                                                                           | 110 |

|                                                                                                                        |     |
|------------------------------------------------------------------------------------------------------------------------|-----|
| • Coupling/Locking                                                                                                     | 110 |
| • Documentation/Reporting                                                                                              | 111 |
| 7.4.6.6 Costs                                                                                                          | 111 |
| 7.4.6.7 Special references and websites related to strong-motion<br>Installations, measurements, data analysis and use | 111 |
| • Other sources                                                                                                        | 112 |
| <b>7.5 Marine seismic observation (M. Shinohara, K. Suyehiro, and<br/>H. Shiobara, version 2011)</b>                   | 112 |
| 7.5.1 Deep ocean environment, required logistics, sensors and data recording                                           | 112 |
| 7.5.1.1 Introduction                                                                                                   | 112 |
| 7.5.1.2 Access                                                                                                         | 112 |
| • The ship                                                                                                             | 113 |
| • Ship capability                                                                                                      | 113 |
| • Applying for and sharing ship time                                                                                   | 114 |
| • Legal matters                                                                                                        | 114 |
| • Environment                                                                                                          | 115 |
| 7.5.1.3 Sea water and unconsolidated sedimentary layer                                                                 | 115 |
| 7.5.1.4 Noise                                                                                                          | 115 |
| 7.5.1.5 Sensors                                                                                                        | 117 |
| • Geophones                                                                                                            | 118 |
| • Active type (moving coil)                                                                                            | 118 |
| • Active type (liquid)                                                                                                 | 118 |
| • Servo-type accelerometer and MEMS                                                                                    | 118 |
| • Broadband sensors (CMG-1T, CMG-3T)                                                                                   | 118 |
| • Leveling of sensors                                                                                                  | 120 |
| 7.5.1.6 Data recording                                                                                                 | 120 |
| • The necessity of low-power compact systems                                                                           | 120 |
| • A/D conversion                                                                                                       | 120 |
| • Data storage media (Tape, MO, HD, Memory)                                                                            | 121 |
| 7.5.2 Ocean Bottom Seismographs (OBS)                                                                                  | 121 |
| 7.5.2.1 Introduction                                                                                                   | 121 |
| 7.5.2.2 Pop-up type OBS                                                                                                | 122 |
| • Short-term OBS                                                                                                       | 122 |
| • Long-term OBS                                                                                                        | 123 |
| • Broadband OBS                                                                                                        | 124 |
| • Ocean bottom accelerometer (OBA)                                                                                     | 125 |
| 7.5.2.3 Cabled OBS system                                                                                              | 125 |
| 7.5.2.4 Ocean floor borehole system                                                                                    | 126 |
| 7.5.2.5 Dealing with multiple-OBS long-term data sets                                                                  | 127 |
| 7.5.2.6 State of OBS activities                                                                                        | 128 |
| 7.5.3 Examples of sea floor seismic observations                                                                       | 129 |
| 7.5.3.1 Local earthquakes                                                                                              | 129 |
| 7.5.3.2 Teleseismic events                                                                                             | 131 |
| 7.5.3.3 Tsunami events                                                                                                 | 134 |
| 7.5.3.4 Technical specifications examples                                                                              | 134 |
| • Specification of ST-OBS                                                                                              | 134 |
| • Specification of LT-OBS                                                                                              | 134 |
| • Specification of BBOBS                                                                                               | 135 |
| • Specification of OBA                                                                                                 | 135 |

|                                      |     |
|--------------------------------------|-----|
| <b>Acknowledgments</b>               | 135 |
| <b>Recommended overview readings</b> | 135 |
| <b>References</b>                    | 136 |

## **7.1 Factors affecting seismic site quality and site selection procedure (A. Trnkoczy; version 2002)**

### **7.1.1 Introduction**

Seismic site selection is not often given the amount of study it requires. The capacity of any new seismic network to detect earthquakes and to record representative event waveforms will be governed by the signal and noise characteristics of its sites, no matter how technologically advanced and expensive the equipment used. If seismic noise at the sites is too high, many of the benefits of modern, high dynamic-range equipment will be lost. If the noise contains excessive spikes or other transients, or if man-made seismic noise is present, high trigger thresholds will be needed and result in poor network detectability. If a station is situated on soft ground, very broadband (VBB) or even broadband (BB) recording can be useless and short-period (SP) signals may be unrepresentative due to local ground effects. If the network layout is inappropriate, the location of seismic events will be inaccurate, systematically biased, or even impossible. A professional site-selection procedure is therefore essential for the success of any new seismic station or network.

It is best to begin the process of site selection by choosing, generally, two to three times as many potential sites as will finally be used. One can then study each one and choose the sites that meet as many desired criteria as possible. One may even model the performance of a few most-likely network layouts and, by comparing the results, be able to make an informed decision about which layout will best record and locate seismic events.

All parameters relevant to the site selection process are discussed here and the process is demonstrated by seeking the best locations for a six-station network around a nuclear power plant. The main goals of this particular project (Trnkoczy and Živčić, 1992) were to monitor local seismicity with a high network detectability and the ability to accurately locate local events. Thus, the placement of short-period seismometers and of surface seismic vaults were mainly, but not solely, considered.

### **7.1.2 Offsite studies**

The site selection procedure includes off-site studies and fieldwork. Off-site, or "office" studies are relatively inexpensive. They should be performed first. One can study maps and gather information about the potential sites from local and regional authorities. Once we have gathered all this information, it is likely that many potential sites will be eliminated for one reason or another. This will minimize future fieldwork and its associated costs.

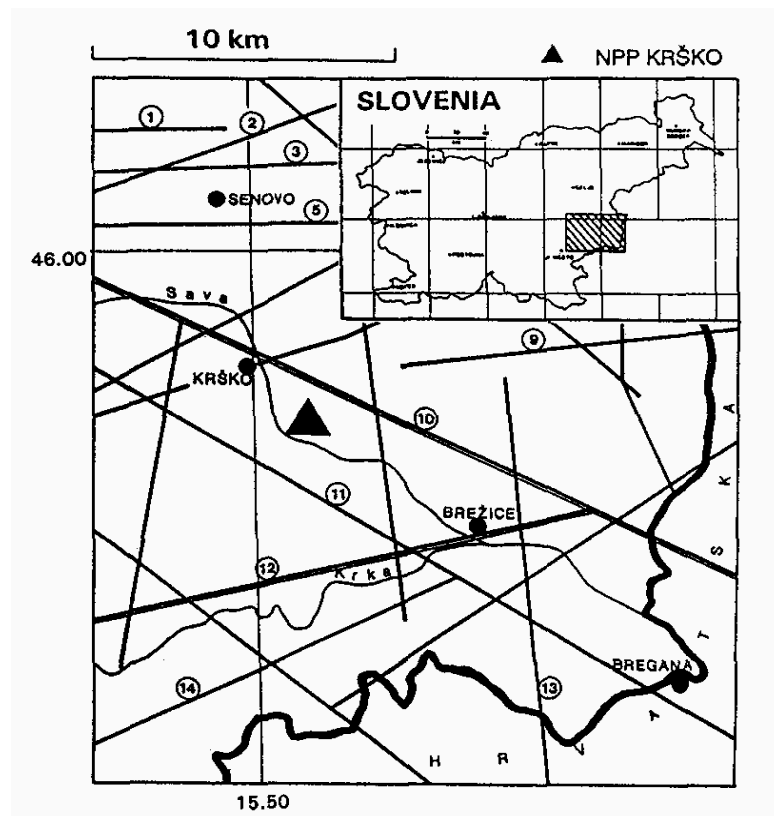
A list of parameters usually included in the off-site study includes:

- geographic region of interest
- seismo-geological conditions
- topographic conditions
- accessibility

- seismic noise sources in the region
- data transmission and power considerations
- land ownership and future land use issues
- climatic conditions

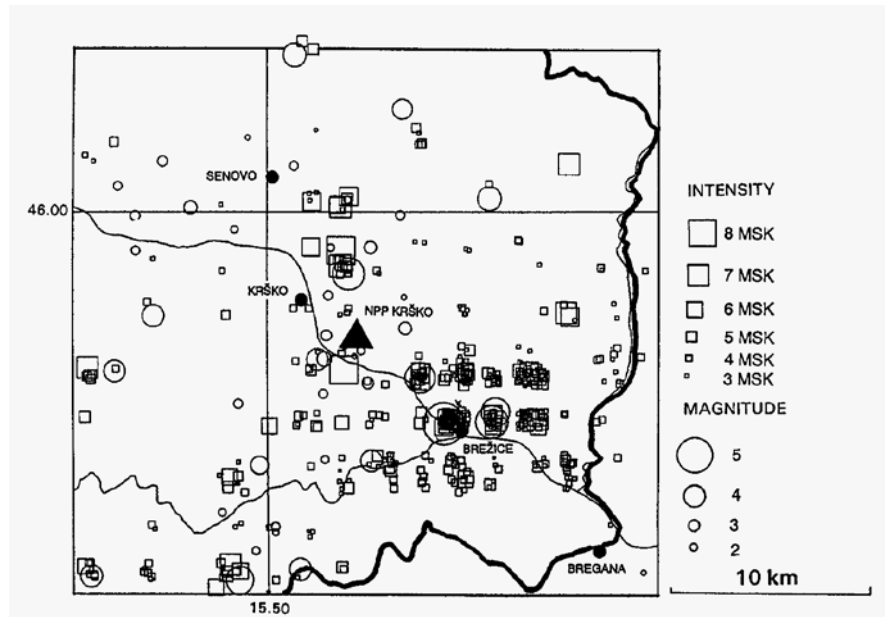
#### 7.1.2.1 Definition of the geographic region of interest

The first step is defining the goals and the geographic region of interest taking both socio-economic and seismic information into account. If the main goal of the new seismic network is monitoring of the general seismicity in an entire country, this stage is greatly simplified. For other projects, one has to examine all the known major geologic faults from geological maps with a view to assess their neotectonic activity and potential, identify seismotectonic features from seismotectonic maps, if available, and compile all available information about the seismicity in the area. One has also to compile historical and instrumentally recorded events in the broader region from earthquake catalogs and other sources. The results of such a study are shown in the following figures for an area in Slovenia. Fig. 7.1 shows the broader region chosen for our example and the main geological faults within it.

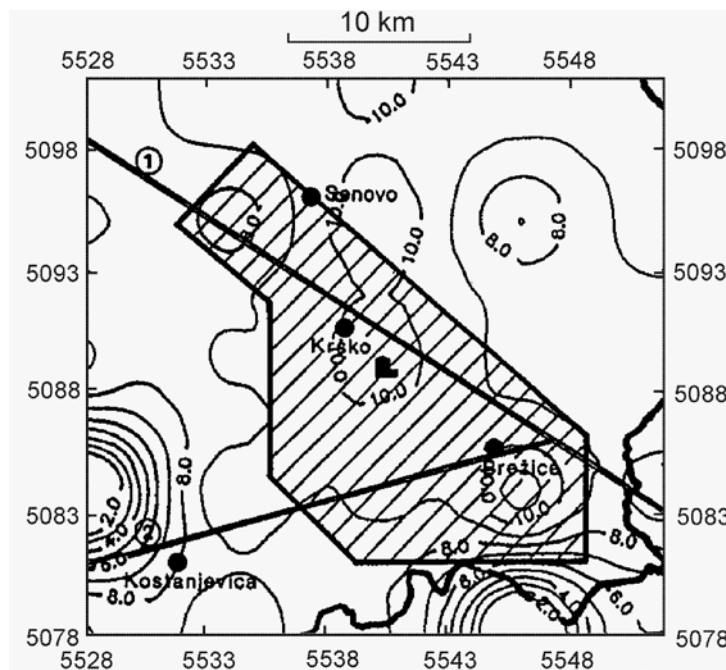


**Fig. 7.1** The broader region chosen for the network in Slovenia and the main geological faults in this area (9 - Artice fault, 10 - Brestanica fault, 11 - Sava fault, 12 - Podbočje fault, 13 - Brežice fault, 14 - Orehovec fault).

Fig. 7.2 shows the distribution of earthquake epicenters as taken from seismic catalogs while Fig. 7.3 depicts the isolines of seismic energy release during the time-span of the catalogs and the hatched area finally chosen for the detailed study.



**Fig. 7.2** Earthquakes in the wider region under investigation in Slovenia. The data were compiled from all available earthquake catalogs.

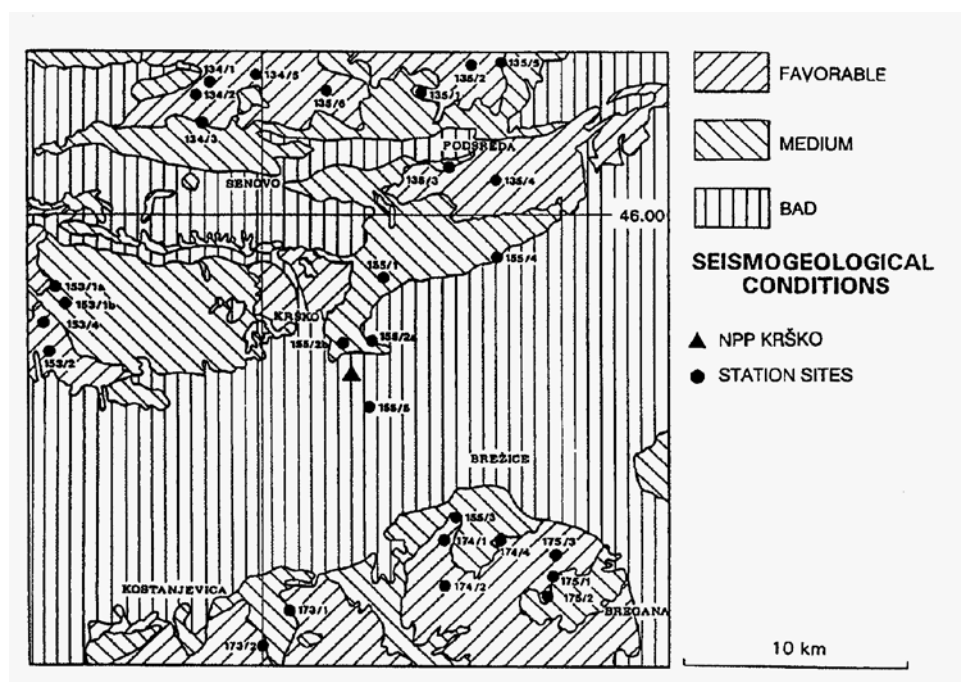


**Fig. 7.3** Final choice of the area to be studied in detail by the seismic network (hatched area). Also shown are the isolines of released log-seismic energy during the time-span of the catalogs (in J/km<sup>2</sup>).

#### 7.1.2.2 Seismo-geological considerations

The underground conditions at a station influence both the seismic signal and the noise conditions and thus have a significant bearing on the potential sensitivity of a seismic station. Usually, the higher the acoustic impedance of the bedrock, the smaller the seismic noise and the higher the maximum possible gain of a station. Therefore, for each new seismic network,

one should at least prepare a map showing simplified seismo-geological conditions. One may then infer a related map in terms of acoustic impedance or bedrock quality grades with respect to their suitability for the installation of seismic recording sites. Fig. 7.4 shows an example for the region under study while Tab. 7.1 gives an example of how bedrock “quality” grades may be classified.



**Fig. 7.4** Subdivision of the region shown in Fig. 7.2 into three grades of bedrock quality. The dots mark the positions of considered station sites.

**Tab. 7.1** Classification of different types of outcropping geological formations in “quality” categories (according to R. Vidrih, personal communication 2001). Grade 5 is the best rock for seismic recordings and grade 1 is the worst.

| Grade | Type of sediments/rocks                                                                                                   | S-wave velocity   |
|-------|---------------------------------------------------------------------------------------------------------------------------|-------------------|
| 1     | Unconsolidated (Alluvial) sediments (clays, sands, mud)                                                                   | < 100 – 600 m/s   |
| 2     | Consolidated clastic sediments (sandstone, marls); schist                                                                 | 500 – 2100 m/s    |
| 3     | Less compact carbonatic rocks (limestone, dolomite) and less compact metamorphic rocks; conglomerates, breccia, ophiolite | 1800 – 3800 m/s   |
| 4     | Compact metamorphic rocks and carbonatic rocks                                                                            | 2100 – 3800 m/s   |
| 5     | Magmatic rocks (granites, basalts) ; marble, quartzite                                                                    | 2500 - > 4000 m/s |

Note: Shear-wave velocities given by engineers (e.g. Ambraseys et al., 1996) in relation to the category “bedrock” (> 750 m/s) are significantly smaller than for competent hard rock due to near surface weathering (see 7.1.3.3) and the consideration of very short wavelengths only.

### 7.1.2.3 Topographical considerations



The topography in the vicinity of a potential site has to be considered. Extremely steep mountain slopes or deep valleys may unpredictably and unfavorably influence seismic waveforms and signal amplitudes. In addition, mountain peaks are usually much more susceptible to wind-generated seismic noise, lightning strikes, and perhaps icing of the communications equipment. Therefore it is wise to avoid such locations, if possible. Sites in moderately changing topography are preferable.

The topography also has to be considered for radio-frequency (RF) telemetry networks. Establishing RF links is much simpler if hill-top sites are selected, but it is important not to let this consideration compromise the seismological considerations. (See IS 7.2 Using existing communication tower sites as seismic sites.)

#### **7.1.2.4 Station access considerations**

Seismic stations are generally located in remote areas, as far as possible away from any human activity. This can often result in relatively difficult access. Public roads do not (or should not) reach most good seismic stations and walking a considerable distance, or the use of off-road vehicles, is more or less inevitable. Inexperience in site-selection often leads to too much compromise in this respect. One needs to find a reasonable trade-off between remoteness and ease of access. Stations which are too difficult to access are expensive to establish and maintain. In consequence, they often suffer from inadequate maintenance and long repair times.

Road maps and 1:25,000 scale topographic maps usually allow an approximate estimate of the difficulties and time needed to access any potential sites. In mountainous regions both the distance from the nearest road accessible by vehicle and the elevation difference between the site and the last point accessible by vehicle are important. One should allow between 15 and 30 min of cross-country walking time for each km of distance (25 to 50 min for each mile), depending on vegetation cover, and between 20 and 30 min for each 100 m (300 feet) of height difference. Stations which require more than half an hour of cross-country walking are rare. However, one has sometimes to accept longer walking distances, particularly if RF telemetry is involved.

Seismic stations are frequently set up at existing meteorological stations. This often happens in countries which are not experienced in seismometry and especially when meteorological institutions are appointed to maintain seismic installations. Such combination of stations or network operations are not advisable, since seismological and meteorological site selection criteria are very different.

#### **7.1.2.5 Evaluation of seismic noise sources**

An assessment of man-made and natural seismic noise sources in the region from maps is only the first stage of a proper seismic noise study. It should always be followed by field measurements of the noise. Nevertheless, road and railway traffic, heavy industry, mining and quarry activities, extensively exploited agricultural areas, and many other sources of man-made seismic noise around the potential sites, along with natural sources like ocean and lake shores, rivers or waterfalls can be evaluated in a qualitative manner from the maps and by inquiry of local authorities. Willmore (1979) gives valuable information about the recommended minimum distances between the site and these types of noise sources.

Distances are given for three different sensitivities of a seismic station, two different geological conditions and both high and low seismic coupling between noise source and station site. The table is reproduced in IS 7.3 along with instructions for its use. An example for its application for the station Loma Palo Benito is given in Fig. 7.5. Nearly all the minimum distance requirements for recordings with a gain around 1 Hz of between 50,000 and 150,000 are fulfilled (the distance to the lake shore is an exception). Six criteria are not fulfilled for a gain of 200,000 or more (see shaded cells).

Note that the above guidelines were designed for 1960's technology (analog paper seismographs). They are most applicable for seismic signal frequencies above 0.1 Hz; i.e., for the medium- and high-frequency range of seismic signals. Seismic noise at lower frequencies is mainly influenced by seismo-geological and climatic conditions (see 7.1.2.8) at the recording site and much less by the seismic noise sources dealt with in the table.

| STATION SITE NAME: Loma Palo Bonito<br><br>COORDINATES:<br>N 18° 46' 58.4"<br>W 70° 13' 20.1" |   | SITE #:7                              |      |       | DATE OF VISIST:<br>02/14/1998 |      |      | ACTUAL<br>DISTANCE |
|-----------------------------------------------------------------------------------------------|---|---------------------------------------|------|-------|-------------------------------|------|------|--------------------|
|                                                                                               |   | HARD ROCK<br>GRANITE, ETC.            |      |       | HARDPAN<br>HARD CLAY, ETC.    |      |      |                    |
|                                                                                               |   | RECOMMENDED MINIMAL DISTANCES<br>[KM] |      |       |                               |      |      |                    |
|                                                                                               |   | A                                     | B    | C     | A                             | B    | C    |                    |
| 1. Oceans, coastal mountains systems                                                          |   | 300                                   | 50   | 1     | 300                           | 50   | 1    | 75                 |
| 2. Large lakes                                                                                |   | 150                                   | 25   | 1     | 150                           | 25   | 1    | 22                 |
| 3. Large dams, waterfalls                                                                     | a | 40                                    | 10   | 1     | 150                           | 25   | 5    | 22                 |
|                                                                                               | b | 60                                    | 15   | 5     | 50                            | 15   | 10   |                    |
| 4. Large oil pipelines                                                                        | a | 20                                    | 10   | 5     | 30                            | 15   | 5    |                    |
|                                                                                               | b | 100                                   | 30   | 10    | 100                           | 30   | 10   |                    |
| 5. Small lakes                                                                                | a | 20                                    | 10   | 1     | 20                            | 10   | 1    | 20                 |
|                                                                                               | b | 50                                    | 15   | 1     | 50                            | 15   | 1    |                    |
| 6. Heavy machinery, reciprocating machinery                                                   | a | 15                                    | 3    | 1     | 20                            | 5    | 2    | 25                 |
|                                                                                               | b | 25                                    | 5    | 2     | 40                            | 15   | 3    |                    |
| 7. Low waterfalls, rapids of a large river, intermittent flow over large dams                 | a | 5                                     | 2    | 0.1   | 15                            | 5    | 1    |                    |
|                                                                                               | b | 15                                    | 3    | 1     | 25                            | 8    | 2    | 6                  |
| 8. Railway, frequent operation                                                                | a | 6                                     | 3    | 1     | 10                            | 5    | 1    | 40                 |
|                                                                                               | b | 15                                    | 5    | 1     | 20                            | 10   | 1    |                    |
| 9. Airport, air traffic                                                                       |   | 6                                     | 3    | 1     | 6                             | 3    | 1    |                    |
| 10. Non-reciprocating machinery, balanced industrial machinery                                | a | 2                                     | 0.5  | 0.1   | 10                            | 4    | 1    | 25                 |
|                                                                                               | b | 4                                     | 1    | 0.2   | 15                            | 6    | 1    |                    |
| 11. Busy highway, large farms                                                                 |   | 1                                     | 0.3  | 0.1   | 6                             | 1    | 0.5  | 2.3                |
| 12. Country roads, high buildings                                                             |   | 0.3                                   | 0.2  | 0.05  | 2                             | 1    | 0.5  | 2.0                |
| 13. Low buildings, high trees and masts                                                       |   | 0.1                                   | 0.03 | 0.01  | 0.1                           | 0.1  | 0.05 | 0.03               |
| 14. High fences, low trees, high bushes, large rocks                                          |   | 0.05                                  | 0.02 | 0.005 | 0.06                          | 0.03 | 0.01 | 0.02               |

#### Legend:

A - Seismic station with a gain of 200,000 or more at 1 Hz

B - Seismic station with a gain from 50,000 to 150,000 at 1 Hz

C - Seismic station with a gain of approximately 25,000 at 1 Hz

a - Source and seismometer on widely different formations or that mountain ranges or valleys intervene

b - Source and seismometer on the same formation and with no intervening alluvial valley or mountain ranges

**Fig. 7.5** Minimum recommended noise-source-to-station-site distances according to Willmore (1979) and actual distances for the seismic station Loma Palo Bonito, which is placed on hard granite rock. Shaded cells indicate that for these criteria the conditions for a class A site are not fulfilled.

Nowadays, with the ready availability of seismic recorders with a large dynamic range, it would be preferable to express the seismometer gain classes A – C in terms of the achievable minimum resolution of ground displacement or velocity amplitudes above the noise level at about 1 Hz. These would be approximately  $< 5$  nm or  $< 30$  nm/s, respectively, for class A and about 2-4 times and  $> 8$  times larger for classes B and C.

For each potential site in a network, one should determine, using maps, the actual distances of the site from relevant seismic noise sources (the extreme right column) and compare them with the recommended minimum distances. The sites which satisfy all or most of the recommendations are the best. Note, however, that local seismic noise sources like trees, buildings, fences, would require on-site evaluation. This information can be added to the table later during fieldwork.

Once we have gathered this information for all the potential sites in a network, we can draw a map, similar to that in Fig. 7.6, where all the potential sites and minimum recommended distances from known seismic noise sources are shown. The latter is achieved by drawing circles around point noise sources and bands of appropriate width along roads or railways. This gives a good overview of all the noise sources at once and helps us to see which ones and how many of them influence a particular potential seismic site.



**Fig. 7.6** Map of the seismic network region with all potential station sites (full dots) and known seismic noise sources (roads, railway, cities, villages, industrial facilities, quarries, etc) with circles of minimum recommended distances drawn around them for the case of gain 25.000 for SP seismic stations at 1 Hz set on hard clay, hardpan and similar ground, i.e., case

C b (i.e., source and seismometer on same formation and with no intervening alluvial valley or mountain range) according to Willmore (1979).

#### **7.1.2.6 Seismic data transmission and power considerations**

For radio-telemetry networks we must consider the topography within the entire network in order to design the data transmission links. Topographic maps (1:50.000 or 1:25.000) are best for this purpose. We look for a topography which enables reliable direct radio frequency (RF) links from the remote stations to the central recording site, or the minimum number of RF repeaters if topography and/or distance do not allow direct connection. More information is given in section 7.3.

If telephone lines are used for seismic data transmission, we must first check for line availability and the distances over which new lines would have to be installed. This information can be obtained from local telephone companies. New phone lines are often a significant proportion of the total cost of site preparation.

The next question concerns the power supply. If mains power is not available on site, we need to calculate the distance over which new power lines would have to be laid and the likely costs. If this is not possible, or the cost is too great, the cost for solar panels has to be evaluated.

#### **7.1.2.7 Land ownership and future land use**

During planning of a new network it is very important to clarify the ownership of the land being considered for a station and any plans for its future use. It makes no sense to undertake extensive studies if one is actually unable to use certain sites because of property ownership issues or if it appears that future development will make the site unsuitable for a seismic station. This information should be gathered from local (land ownership) and regional (future land use) public offices and authorities.

If the land is privately owned, one should contact the owner as soon as possible and make every effort to agree on a renting or purchasing contract to the satisfaction of both parties. It is very important to have "friends" rather than "enemies" around the seismic stations. In many countries this may be very important for the security of the installed equipment.

#### **7.1.2.8 Climatic considerations**

Several climatic parameters can influence seismic site selection. Regional or national meteorological surveys can provide this information. It can also be found in yearly or longer-term bulletins, which are published by nearly every meteorological institution. In developing countries it is sometimes not easy to get complete information. However, we do not need precise values for these parameters and even rough estimates can help in site selection and design of seismic shelters.

The following climatic parameters are important:

- The minimum and maximum temperatures at a site determine how much thermal insulation will be needed for the seismic vault and instruments. Temperatures below

zero degrees Celsius may cause icing of antennae. Special shielding is often required in high mountains and polar regions.

- We need to know the frequency and maximum wind speeds at sites. Wind is a major source of seismic noise, so sites with less wind are preferable to sites placed on windy mountain ridges.
- Solar data is needed to determine the minimum size required for solar panels, if they are required to provide power. The number of sunny days in the worst month and/or the longest expected uninterrupted cloudy period in a year can serve as a measure.
- The frequency and amount of precipitation (total precipitation per year and maximum precipitation per hour) will determine protection measures required to keep the vaults dry.
- In colder climates, annual snowfall levels determine how accessible a station will be during the winter, the waterproofing measures required and – if used – the optimum installation angle and size for solar panels.
- Protection against lightning is very important and has significant financial consequences. One needs to decide on what protection equipment is necessary using information on the observed frequency of lightning. Alternatively, one has to calculate how much lightning damage is likely if protection measures are not implemented. The best method for this is to obtain isokeraunic isolines, which are related to the probability of a lightning strike. This data is rarely available and it is often easier to obtain less specific but more generally available meteorological parameters – such as the annual number of days with severe thunderstorms in the area. Lightning usually varies enormously from one region to another and also varies locally, depending on the topography. Serious consideration of these parameters and the knowledge of local people on these issues are definitely worthwhile.

### 7.1.3 Field studies

Field studies are the next step in the site selection process. Expect to make several visits to each potential site. A seismologist familiar with seismic noise measurements, a seismo-geologist, and a communications expert (if a telemetry network is considered) should all visit the sites. You should allow between one and three days per site to accomplish the fieldwork. This assumes that all pertinent maps and information are available in advance and the logistics are well organized. Much also depends on a country's infrastructure and the size of the network. If the network will use RF telemetry, add an extra 20% to the time for topographical profiling and RF link calculations.

If site selection is purchased as part of the services provided by an equipment manufacturer, see IS 7.1 for a summary of the information that should be provided to them.

In general, experts visiting the sites should:

- verify the ease (in any weather) of access to the site;
- search for very local man-made seismic noise sources which might influence the site, but may not be indicated on maps (see text to Fig. 7.7);
- perform seismic noise measurements;
- study the local seismo-geological conditions;
- investigate the local RF data transmission conditions (if applicable);
- verify availability of power and telephone lines.

#### **7.1.3.1 Station access verification**

Station access should generally be possible throughout the year. However, a few days of inaccessibility due to snow or high water per year can normally be tolerated. This can be checked by talking to local people.

If non-public dirt roads are used to access the site, we need to ask about the future of these roads since roads built and owned by private, military, or forest authorities are sometimes abandoned. If there is no guarantee that such roads will be maintained in future, it is better to reposition the seismic site.

#### **7.1.3.2 Local seismic noise sources and seismic noise measurements**

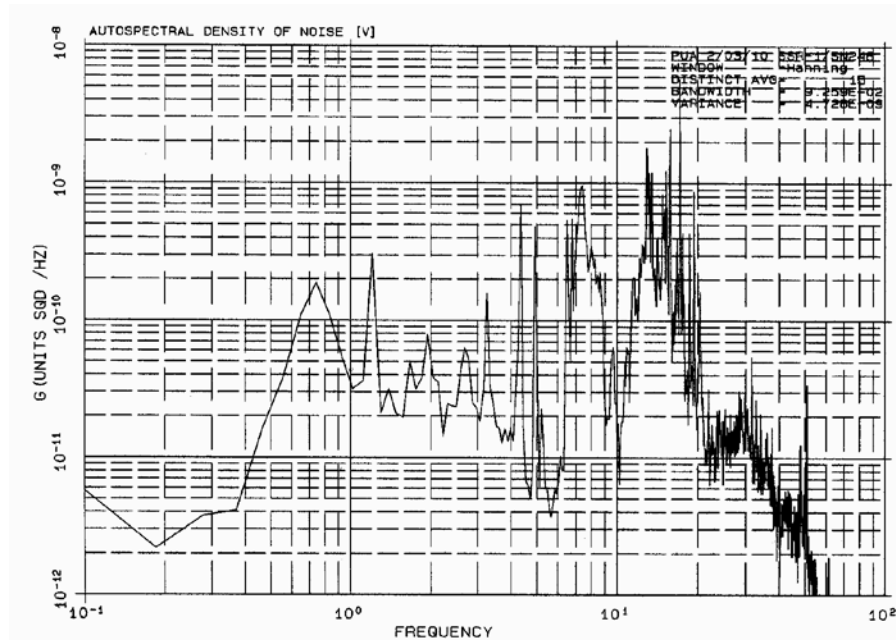
During fieldwork, one should explore the vicinity of the potential site for local sources of seismic noise, usually man-made, which may not be resolvable from the available maps. A single small private "industrial" facility too close to the site may ruin its seismic noise performances completely. Local people are the best source of information.

Measuring seismic noise at the site is an important task. Seismic noise varies greatly depending on the season of the year, weather conditions, and innumerable daily occurrences. Seasonal variability of seismic noise has mainly natural causes and is clearly developed for periods,  $T$ , greater than 2 s. The variation may be as large as 20 dB at the spectral peak for ocean-storm microseisms close to  $T = 7$  s. In contrast, high-frequency noise is mostly man-made (traffic, machinery), often with a pronounced diurnal variation of the order of 10 to 20 dB. In order to accurately record all these factors, it is best to take measurements at each site over a long period of time; long enough to record a number of earthquakes. These will allow a comparison of the sites based on signal-to-noise ratio, which is the main guiding parameter for the quality of a site.

Sufficiently long measurements are often not performed for financial reasons. In such cases, some measurements are much better than none at all. Short-term measurements can not provide complete information about the noise levels at a site, but they are still very useful to identify man-made noise sources and to assess the daily noise fluctuations in the important frequency range for small local and teleseismic events (i.e. from 0.5 Hz to 20 Hz). It is important that any short-term measurements (say of 15 min duration) are carried out during specific times when maximum and minimum noise conditions are expected.

To assess the potential influence of long-term natural seismic noise variation, we should also obtain noise data from existing seismic stations in the region. If there are none of these, we have to set up one or more temporal reference stations which are not moved from site to site. By comparing noise records taken at the same time at the reference station(s) and the potential new site locations we can, at least with respect to the long-period natural seismic noise, assess the representativeness of the noise data sampled at the potential sites by scaling it to the reference site(s). This assures that any variations in natural seismic noise levels over time will not affect the comparison of different potential sites.

Records of seismic noise are usually presented as noise spectra. These can reveal more information about the type and importance of various seismic noise sources around the site than the corresponding time-domain records alone. A typical noise spectrum is shown in Fig. 7.7. We can easily see high levels of man-made seismic noise (frequencies around 15 Hz). Spectral spikes from 3 to 5 Hz shown in this spectrum originate from heavy machinery working in a quarry at 4 km distance.



**Fig. 7.7** Typical seismic noise spectrum (ground velocity power density in  $\text{m}^2/\text{s}^2/\text{Hz}$ ) at a potential seismic station site showing man-made seismic noise generated by a nearby city and heavy machinery working in a 4 km distant quarry.

However, noise spectra should never be determined without prior inspection of the original time domain records which have to be cleaned of unrepresentative spurious or transient events. Also the analysis of noise conditions should never be made on the basis of the calculated spectra alone but always in conjunction with the related time-domain records. Examples are given in sub-Chapter 7.2.

The data requirements for noise analysis depend on the type of station to be installed. For short-period stations, use noise records that are at least two minutes long to allow calculation of stable seismic noise spectra in the frequency range from 0.1 to 50 Hz. For broadband stations, use noise records that are at least twenty minutes long for noise spectra calculations from 0.01 to 50 Hz. The sampling rate should be at least 100 Hz in both cases. In order to reduce any bias due to diurnal noise variations, the measurements at the various sites should be taken at about the same time of the day. Whenever possible, use identical equipment and processing methods at all potential sites and at the reference station(s). This greatly simplifies the normalization procedure. More information about seismic noise and its measurement is given in sub-Chapter 7.2

It should be mentioned here that the assessment of seismic noise for a Very-Broad-Band (VBB) seismic station requires much more effort. Days or even months of measurement are often required to get a full picture of the seismic noise conditions at the potential site (see

Uhrhammer et al., 1998). A quiet short-period station site is not necessarily a good long-period noise site. Seismic noise may behave differently in the different frequency ranges.

### **7.1.3.3 Field study of seismo-geological conditions**

A seismo-geologist should study the geology to determine its local complexity. Uniform local underground conditions are preferred for seismic stations. The seismo-geologist should also verify the actual quality of bedrock as compared to that given in geologic maps and try to estimate the degree of weathering that local rocks have undergone. This can sometimes give a rough estimate of the depth required for the seismic vault to place the seismometers on unweathered bedrock. Unfortunately, it is often highly unreliable to judge the required vault depth in this way. At most sites only shallow seismic profiling, drilling, or actual digging of the vault can reliably reveal how deeply the rock is weathered and how deep the seismic vault must be. If shallow profiling is planned (see 7.1.3.5 below), the seismo-geologist should precisely determine the position of the profiles.

If there are local sources of high-frequency seismic noise around the site, a seismologist should carefully assess, both by inspection and measurement, to what extent they might affect recordings at the site. If the noise sources and the site are located on the same rock or soil formation, one can expect a high degree of seismic coupling between the noise source and the station. On the other hand, when the noise sources and the station are located on different geological formations with a significant impedance contrast between them, the seismic coupling is rather weak. In this case even nearby noise sources might not disturb seismic records much. The station BRG in Germany is a striking example. This is one of the best stations in the German Regional Seismograph Network (GRSN). The station is located in the middle of a busy resort town, next to a main road built on the aggraded bank of a rushing creek. The seismographs have been placed 150 m away from the road in an abandoned mining gallery which was driven horizontally from the road level into an outcropping Devonian hornschist rock cliff. Thus, the seismic sensors are well decoupled from the nearby-generated traffic noise. The site quality of BRG would correspond to B in Fig. 7.5.

### **7.1.3.4 Field survey of radio frequency (RF) conditions**

A communications expert visiting the site should examine potential obstacles to radio-wave transmission. He or she should also examine the immediate topography surrounding the site because frequently it can not be resolved from 1:50,000 scale maps, normally used in RF topographical profiling. This study needs to define the minimum required antenna height for reliable data transmission. For more information see sub-Chapter 7.3.

### **7.1.3.5 Shallow seismic profiling**

Shallow seismic profiling is usually the last step in the site selection process. It is probably the most expensive step and has usually to be contracted out to a seismic-engineering company. It should be done only at the most likely and most important sites. Shallow refraction profiles yield quantitative parameters on the rheological quality of the bedrock and enable determination of the depth of weathering. The results can determine the best position of the seismic vault as well as its required depth. One should use two approximately perpendicular



profiles, each about 100 meters long, in order to determine the seismic wave velocity (for P and/or S waves, depending on the type of source used) down to a depth of 20 to 30 meters. This is enough even for the deepest seismic vaults considered. If the seismometer is to be installed in a borehole, seismic profiling needs to penetrate to depths of about 100m, the typical maximum borehole depth.

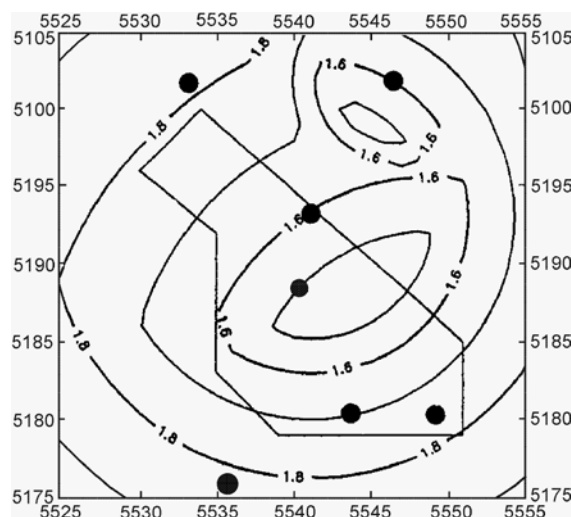
If seismic profiling is not included in the site evaluation, most likely for financial reasons, unexpected results may occur when digging the seismic vault. One should dig until reaching bedrock and that can sometimes be unexpectedly deep. One needs to anticipate that vaults will have to be repositioned and re-dug if weathered bedrock happens to be extremely thick. This often makes the relatively high cost of profiling a wise investment. The same argument applies to boreholes, although it is easier and less costly to deepen or move a borehole than it is for a vault.

#### 7.1.4 Using computer models to determine network layout capabilities

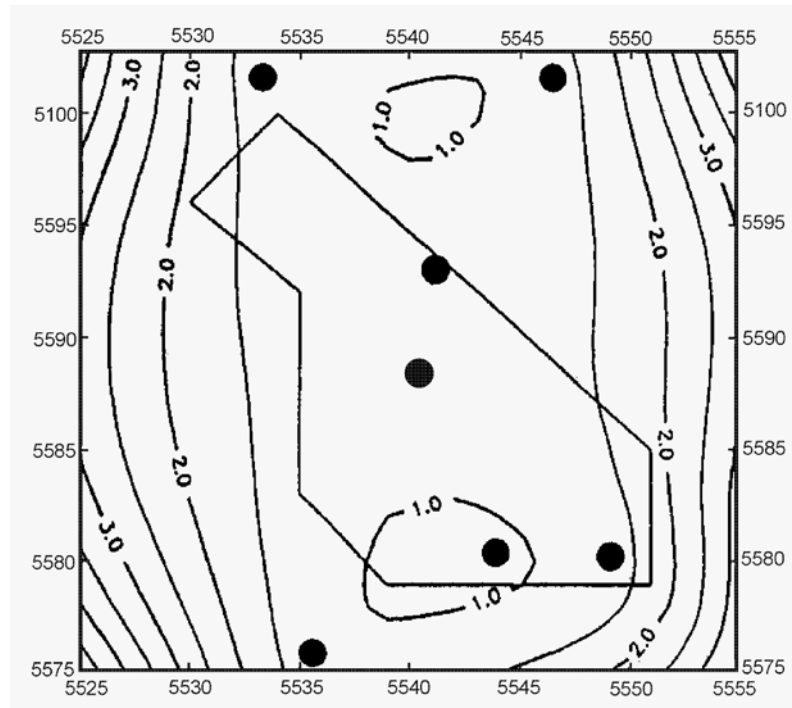
Once we have decided on the final number of seismic stations and are very close to the final layout of the system, meaning that we have chosen two or three possible network layouts, the next useful step is to make a computer model of the network. The modeling should answer the question: Which particular network layout performs best for different aspects of network performances? One can then use these results to choose the best possible network layout for particular requirements. Among the parameters one may wish to study are:

- network detectability in terms of the spatial distribution of minimum magnitude of events which can still be recorded with a given signal-to-noise ratio (Fig. 7.8);
- precision (i.e., calculated accuracy) of event epicenter determinations in the region (Fig. 7.9);
- precision of event hypocenter determination in the region (Fig. 7.10);
- maximum magnitude of events that can be recorded without clipping (this requires an assumed gain and dynamic range of the recording equipment to be used in the network).

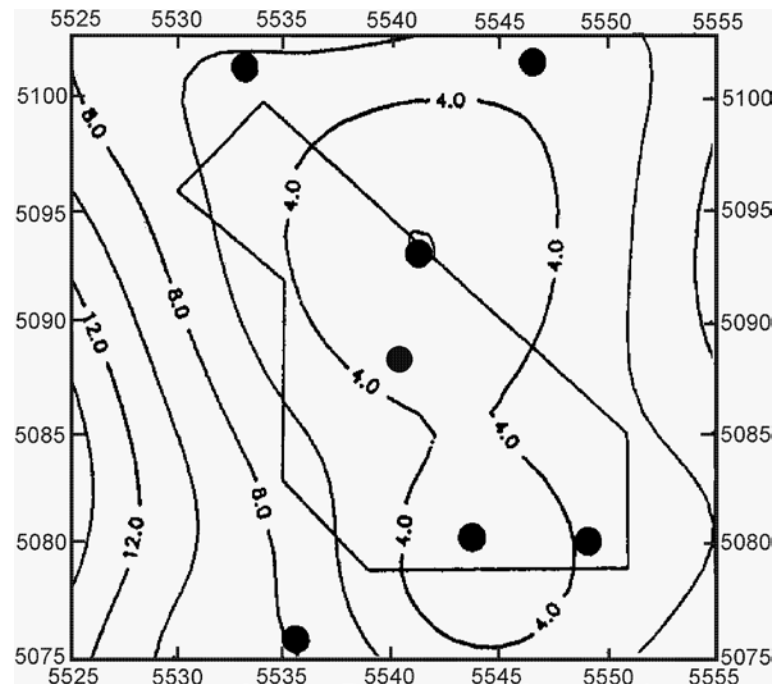
Note that optimal configurations for event location are often not optimal for source mechanism determination, tomographic studies or other tasks (Hardt and Scherbaum, 1994).



**Fig. 7.8** An example of computer modeling of network capabilities. Isolines of minimum magnitude of events detected at 5 seismic stations (from six in the network) with a signal-to-noise-ratio  $>20$  dB are shown for the best of the alternative network layouts.



**Fig. 7.9** An example of computer modeling of network capabilities. Isolines of uncertainty of epicenter determination in km ( $\pm 1$  standard deviation) are shown for the best of the alternative network layouts.



**Fig. 7.10** An example of computer modeling of network capability. Isolines of uncertainty of hypocenter determination in km ( $\pm 1$  standard deviation) are shown for the best of the alternative network layouts.

Several methods for direct computer calculation of optimal network configuration (layout) are described in the literature (e.g., Kijko, 1977; Rabinowitz and Steinberg, 1990; Steinberg et al., 1995). However, practical limiting conditions with respect to infrastructure, topography and accessibility usually outweigh such theoretical approaches. "Optimal" layouts calculated with these methods are rather sensitive to initial conditions such as the predicted gain of stations. This often renders results of questionable value. However, some of these programs may be of help in deciding whether to add or remove stations to an existing network (e.g., Trnkoczy and Živčić 1992; Hardt and Scherbaum 1994; Steinberg et al. 1995; Bartal et al. 2000).

A more detailed discussion of these programs is beyond the scope of this Manual. Simple methods usually suffice for our purposes because we want to compare results for various layout options. Determination of network performances in an absolute sense requires a more sophisticated approach. One program which works rather well for relative performance and which can be made available on request is described in IS 7.4: "Detectability and earthquake location accuracy modeling of seismic networks". The program is based on a simplified and uniform attenuation law for seismic waves in a homogeneous half space or in a single- or double-layer ground model. The software uses estimated uncertainties in the P- and S- wave velocities in the model and in the P- and S-phase readings. The software requires as an input the predicted sensitivity of the seismic stations in the network based on measured seismic noise amplitudes at the sites.

No matter what modeling work is carried out, choosing a seismic network layout always involves making good, educated guesses based on experience.

## **7.2 Investigation of noise and signal conditions at potential sites** (P. Bormann; version 2002)

### **7.2.1 Introduction**

The general factors affecting seismic site quality and suitable site selection procedures have been discussed above. This sub-Chapter discusses specifically the instrumental measurement of seismic noise and signals for optimal site selection, discusses specific features of noise records and spectra from different noise sources and gives recommendations for carrying out such measurements. In the following we discriminate between:

- reconnaissance noise studies prior to station site selection;
- comparison of noise and signal conditions at existing permanent stations;
- searching for alternative sites in a given network.

Examples for each case are based on data from noise surveys in Iran and Germany.

Many sites in a wide area usually have to be inspected and measured during reconnaissance noise studies, sometimes covering an entire country. Carrying out such a survey within a

reasonable time and reasonable cost often dictates making measurements with short-period instruments. These are easily and quickly deployed, require less care than long-period or broadband sensors for thermal shielding and underground tilt stability, and yield stable records immediately after installation and useful high-frequency spectra from a few minutes of recording. Many potential sites can then be measured within a day and thus quickly give a

good idea of their suitability depending on surface geology, topography, distance from potentially disturbing noise sources, etc.

However, short-term measurements using short-period seismographs do not allow judgement of the level of long-period noise ( $T > 3$  s). They are also not very suitable for assessing seasonal or diurnal variation of seismic noise. Furthermore, it is highly unlikely that during the short time windows of measurements, any signals from real seismic events will be recorded which would allow comparison of signal-to-noise-ratios (SNR) at different sites. This is important because sites with the lowest noise are not necessarily the sites with the best signal-to-noise ratio. Signal amplitudes may vary by a factor of three or more, depending on local conditions (see Figs. 4.34 to 4.36).

Nevertheless, short-period and short-term noise measurements are sufficient to get an idea of the high-frequency ( $f > 0.3$  Hz) background noise and to assess the potential influence of various types of man-made noise sources. It is also possible to assess the daily noise variability and to scale and compare measurements at the more remote sites by using a reference station at the nearest main source of man-made noise (town, factory, railway line, high way, etc.), which records throughout the investigation. In this way, we can get a reliable idea of the relative suitability of different potential sites for the frequency range of small local, regional and teleseismic events ( $0.3 \text{ Hz} < f < 30 \text{ Hz}$ ).

Existing permanent recording sites with stable recording platforms and reasonable shielding against environmental influences allow long-term comparative measurements of both seismic noise and signals in a much broader frequency band. These will give a more reliable assessment of the suitability of sites for event detection and location and also for a variety of other seismological tasks, such as source mechanism studies, tomographic studies of the Earth's structure or the use of very long-period normal modes.

If some of the sites within a seismic network are significantly noisier than others, one should look for alternative sites. For a broadband network, the measurements at alternative sites must be made with the same type of broadband sensors and with every precautions for stable installation and appropriate shielding against wind, weather and direct sunshine. The recording time at each site should be long enough to ensure proper stabilization of the sensor after installation (a few hours to days). Additional days or weeks of recording are needed for assessing diurnal noise variability and relative SNRs for local and teleseismic events.

## **7.2.2 Reconnaissance noise studies prior to station site selection**

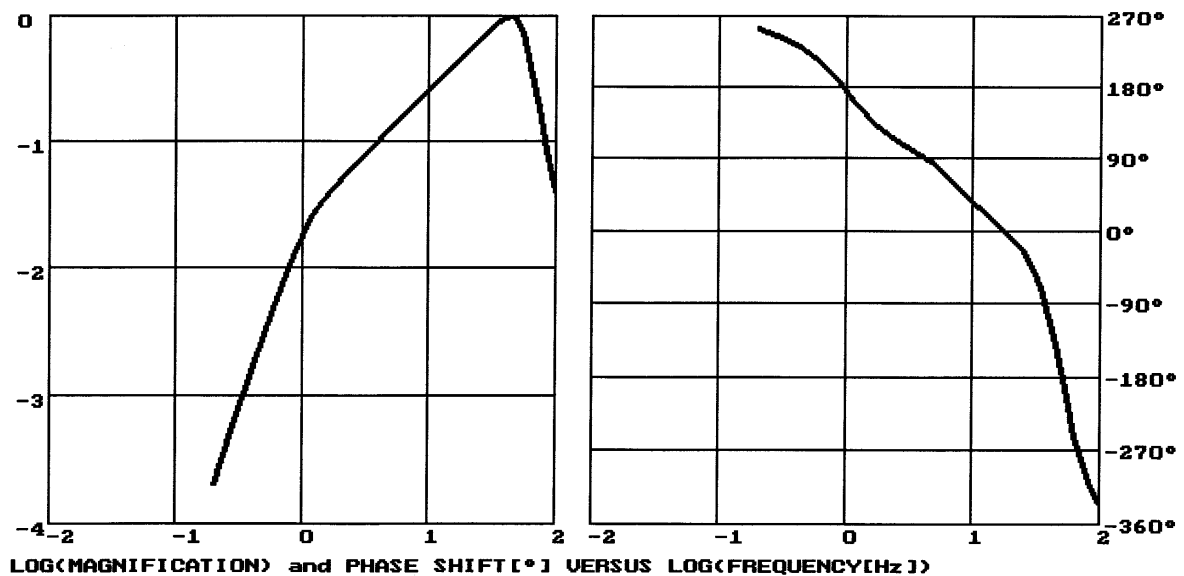
### **7.2.2.1 Offsite assessment of expected noise levels and measurement of instrumental self-noise**

Field measurements should always be preceded by offsite studies (see 7.1.2). They help locate the most promising sites and most likely noise sources, help speed up the measurements and reduce the risk of unwanted surprise in the field and final assessment.

When geologic, environmental, climatic, settlement and infrastructure conditions indicate that sites may have very low levels then only high-performance short-period seismographs with very low instrumental self-noise should be used for noise measurements (see 5.6). The level of self-noise should be measured before going into the field and compared with the global

New Low Noise Model (NLNM) (see Fig. 5.21). The seismometer noise should be at least 6 dB below the minimum local seismic noise for the entire pass band of the sensor. The signal pre-amplification has to be set high enough to ensure that very low-level ambient noise is well resolved. The resolution of the data acquisition unit should be set at about 18 dB (3bits) below the minimum local seismic noise over the pass band of the seismograph. Clearly, the frequency response of the seismograph must be known or has to be determined beforehand (see 5.7 and 5.8 and well as the exercises EX 5.1 to 5.5).

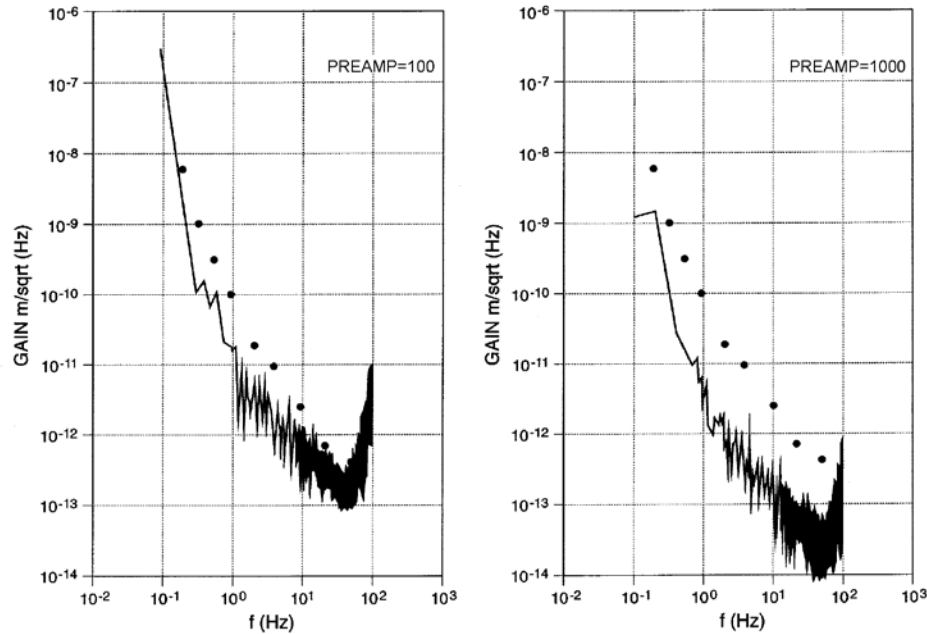
Fig. 7.11 shows the combined frequency response of an SS-1 seismometer and an SSR-1 recorder used in field measurements for site selection in NW Iran.<sup>1)</sup> The sampling rate was 200 Hz using a 6<sup>th</sup> order low-pass filter with corner frequency  $f_c = 50$  Hz in order to avoid spectral aliasing (see 6.3.1). The filter reduces the seismograph gain between  $f_c$  and the Nyquist frequency  $f_{Ny}$  (half of the sampling frequency) in such a way that very small seismic background noise signals no longer may be resolved above the least-count digitizer noise. Correcting the noise spectrum for the decrease in seismograph gain for  $f > f_c$  results in an *apparent* increase of noise power between  $f_c$  and  $f_{Ny}$ . This is clearly to be seen in Fig. 7.12. Here, therefore, we consider only noise spectra up to 1/4 or 1/2 of the sampling frequency.



**Fig. 7.11** Amplitude and phase response curves for the seismometer-recorder combination SS-1/SSR-1 as used in field measurements in NW Iran (see Figs. 7.12 to 7.21). The response is proportional to velocity between about 1 and 50 Hz.

<sup>1)</sup> The data in Iran have been collected as part of a joint project between the International Institute of Earthquake Engineering and Seismology (IIEES) and UNDP (Ref. No.

IRA/90/009). The data relates to the seismic noise measurements at potential station sites for the Iranian National Seismic Network (INSN). The authors thank Prof. M. Ghafory-Ashtiany, President of IESSS, for the technical and staff support provided and for his kind permission to publish part of the data in this Manual.



**Fig. 7.12** Comparison of the average values (●) of the ground displacement spectrum of seismic noise recorded at the quietest site found during a noise survey in NW Iran with the equivalent displacement spectrum of the combined instrumental self-noise of the Kinemetrics SS-1 seismometer and SSR-1 data logger at a pre-amplification level of 100 times (left) and 1000 times (right). One order of magnitude difference in the amplitude spectra corresponds to 20 dB difference in the respective power spectra. In this case only the higher pre-amplification allows the resolution conditions to be met at the quietest sites in the area under investigation. Current data loggers using 24 bit digitizers connected to high output seismometers generally cover the dynamic range down to and below the low noise model. Nevertheless the use of lower output seismometers, e.g. short-period geophones, still require switchable pre-amplifications in the range of 10 to 100.

#### 7.2.2.2 Sensor installation, measurements and logbook entries in the field

Potential measurement and reference sites should be pre-selected before going into the field, based on geologic and road maps and taking into account other significant aspects or findings from preceding offsite studies. The selections may be changed during the field inspection. Essential points to be considered in field studies have already been outlined in section 7.1.3. The following complementary rules should be observed:

- take daily synchronization of the internal clocks of the data loggers used in the field and at the reference site if they have no common time reference such as GPS-controlled clocks.
- keep a log-book;

- note carefully all relevant features which characterize the measurement sites (local geology and topography, compact or weathered rock outcrop, soil type, vegetation cover, distance to settlements or industry, main roads, power lines);
- note the environmental conditions during measurement (weather, wind, rain, insolation) and the occurrence time of any transient events that might have influenced the noise record (e.g., wind gusts or cars, trains or people passing by at what distance);
- mark the position of any measurement site in your road and/or geological map;
- take representative photographs of each measurement site and sensor installation;
- whenever possible, position the seismometer directly on a flat outcropping rock surface and level it with its three adjustment screws. Unusually-long adjustment screws can be fitted to help level the sensor on rough rock surfaces (proper counter-locking of the screws ensures sensor stability);
- in the case of well-binding (clayish) soil, screw long-leg tripod adjustments directly into the soil. Alternatively, position the seismometer on a thick solid rock plate placed firmly on the ground after removing any loose gravel or vegetation (Fig. 7.13). This may be the only reliable solution when making three-component noise measurements if three individual sensors are used requiring identical installation conditions. It may also work well on rough rock surfaces as long as a nearly horizontal stable three-point support of the plate can be found. A rock plate is not necessary if the three components are mounted in the same package, e.g. for Mark L4C-3D seismometers (see DS 5.1).



**Fig.7.13** Temporary three-component reference installation in NW Iran on a leveled marble plate placed on unconsolidated ground. Two other measurement points on the horizon using outcropping hard rock are marked. The noise at the latter sites was close to the NLNM.

- in the case of wind, rain or snowfall, try to find shielding on the lee-side of a rock-face (Fig. 7.14) or bury the sensor in the ground and cover it with a tightened sheet or blanket or with a box;

- if test measurements show that noise levels are comparable for all three components in the area under investigation it is sufficient to continue the survey using only vertical component recordings. This is usually the case in isotropic noise fields, i.e. in the absence of distinct localized noise sources.
- set up at least one continuously-recording reference station in the study area in order to assess the influence of diurnal noise variations on the measurements made at different sites and at different times of the day. The reference station can be used to scale the noise records at the other sites (see Fig. 7.15).



**Fig. 7.14** Hiding with the noise recording equipment on a windy day with snowfall in a small cave on the lee-side of a rock cliff. Surface recordings under adverse weather conditions of the noise level at this site in NW Iran were close to the best sites in good weather.

- take comparative measurements at different distances are recommended to assess the reduction of noise with distance from transient sources (such as nearby road or railway traffic). Measurements on different soil conditions may also be needed if the noise also depends on the lateral impedance contrast of adjacent rock formations (see Figs. 7.16 and 7.17).
- if, at low-noise sites, the ground displacements are of the order of nm ( $10^{-9}$  m), do not stand or walk close to the sensors during the recordings. Stay at least 10 m away, remain sitting down, and keep absolutely quiet (see Fig. 7.18).
- stay several hundreds of meters away from large power lines or transformer houses. Otherwise you may get strong induction currents in the seismometer's measurement coils or record 50 to 60 Hz vibrations that are typical of large transformers or heavily loaded power lines (see Fig. 7.19).

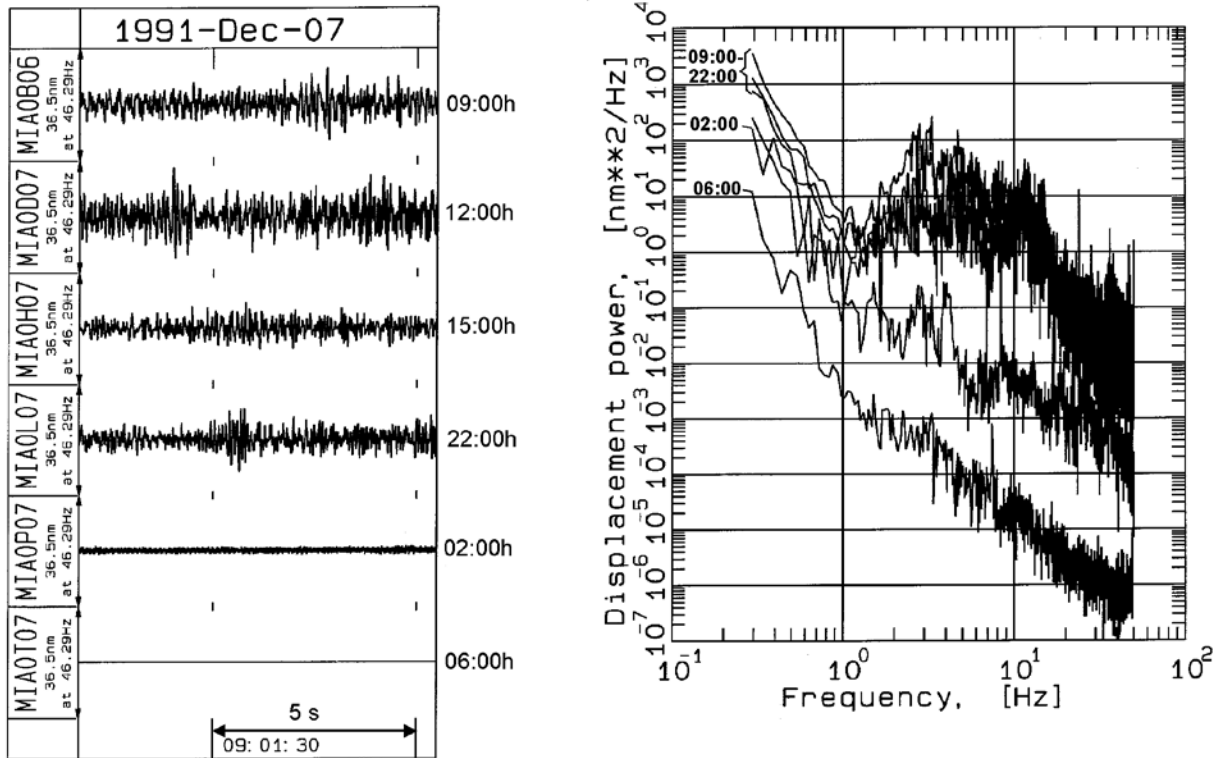
### 7.2.2.3 Case study of noise records in the frequency range $0.3 \text{ Hz} < f < 50 \text{ Hz}$

Fig. 7.15 shows the daily noise variation at a reference site in a town in NW Iran. The noise between night and day time varies by about 20 to 30 dB around 1 Hz and by about 50 dB

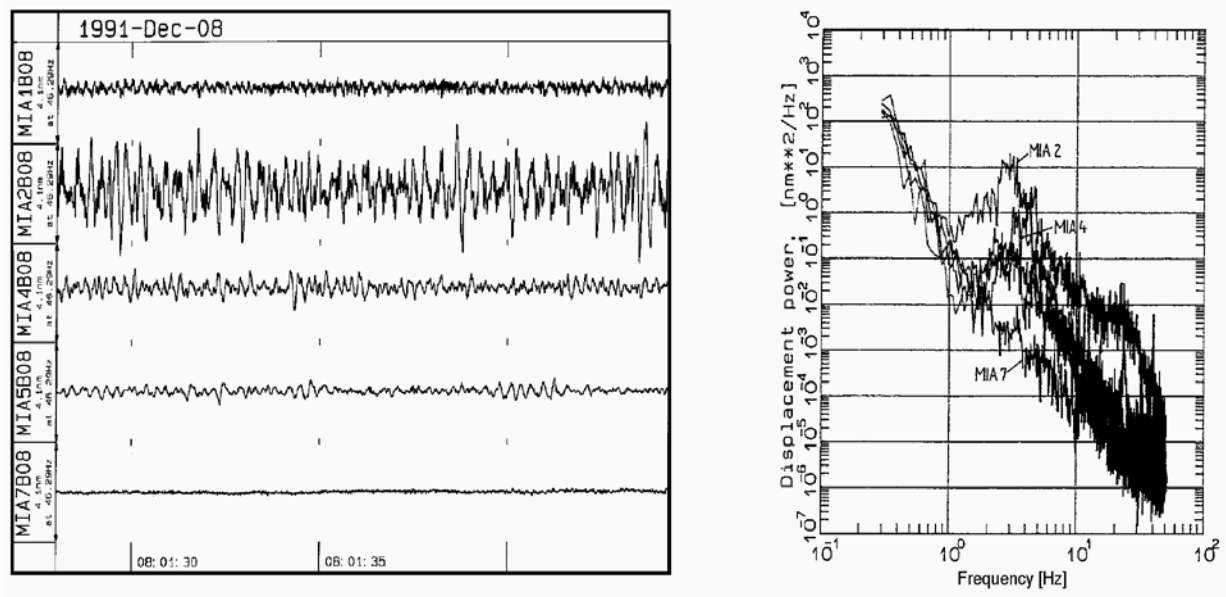


around 10 Hz because of the site's proximity to a main road and poor underground conditions. Fig. 7.16 shows the large dependence of noise records and spectra on the geological underground conditions and remoteness from villages and traffic roads in the area around one of the reference stations in NW Iran .

Note that noise spectra should not be determined unless the related time domain records have been inspected and any non-representative spurious or transient events have been removed. The analysis and assessment of noise conditions should never be made on the basis of the calculated spectra alone but always in conjunction with the related time-domain records.

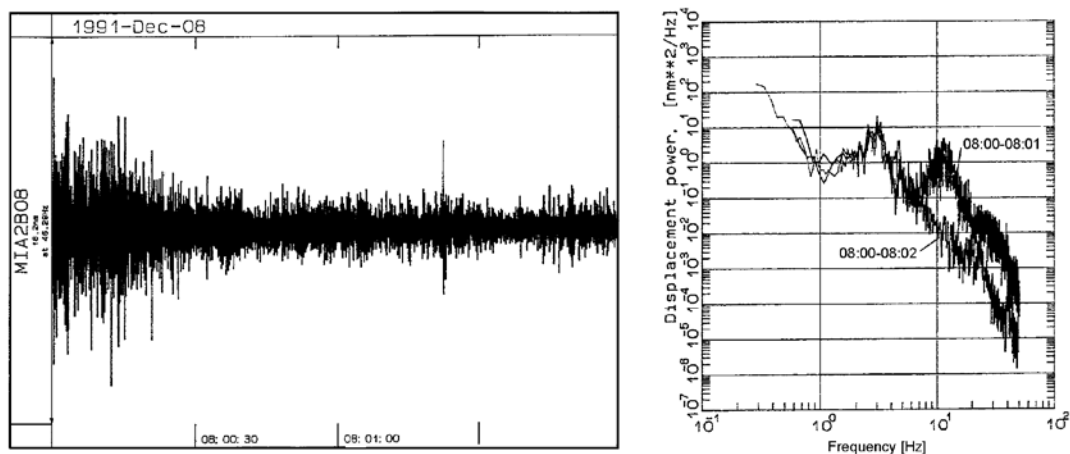


**Fig. 7.15** Comparison of relatively quiet sections of vertical component noise records (left; without strong transients) and related power spectra (right) at a reference site in a town in NW Iran. The measurements were made at different times of the day.



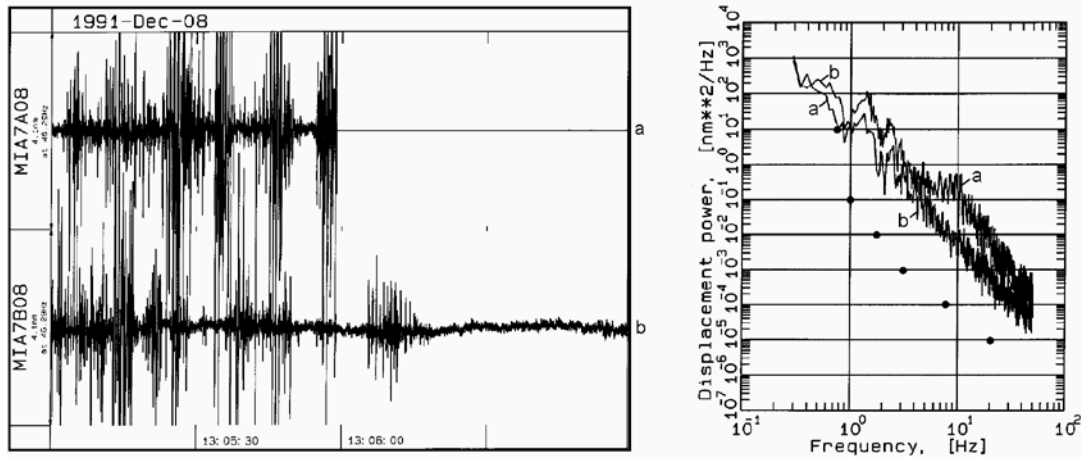
**Fig. 7.16** Noise recordings (left) and related power spectra (right) at different sites in NW Iran. From top to bottom: 1) unconsolidated Miocene terrace, 2) unconsolidated Alluvial valley fill, about 2 km away from the main road, 3) as for 2) but some 5 km away from main road; 4) outcropping volcanic hard rock near the road in a valley (with no nearby traffic at the time of measurement, 5) volcanic hard rock surface near a mountain pass road. The noise at MIA 7 is around 1 Hz very close to the global New Low Noise Model (see 4.1) and at 10 Hz only about 14 dB above it.

Fig. 7.17 shows a two minute noise record (left) and the related power spectra (right). The large amplitudes at the beginning are due to a truck and car passing by on the bumpy country road at some 100 to 400 m distance (documented by photograph and time check). Accordingly, for frequencies  $f > 7$  Hz, the noise power of the first minute of the record is 10 to 20 dB higher than for the background noise after the transient is over.



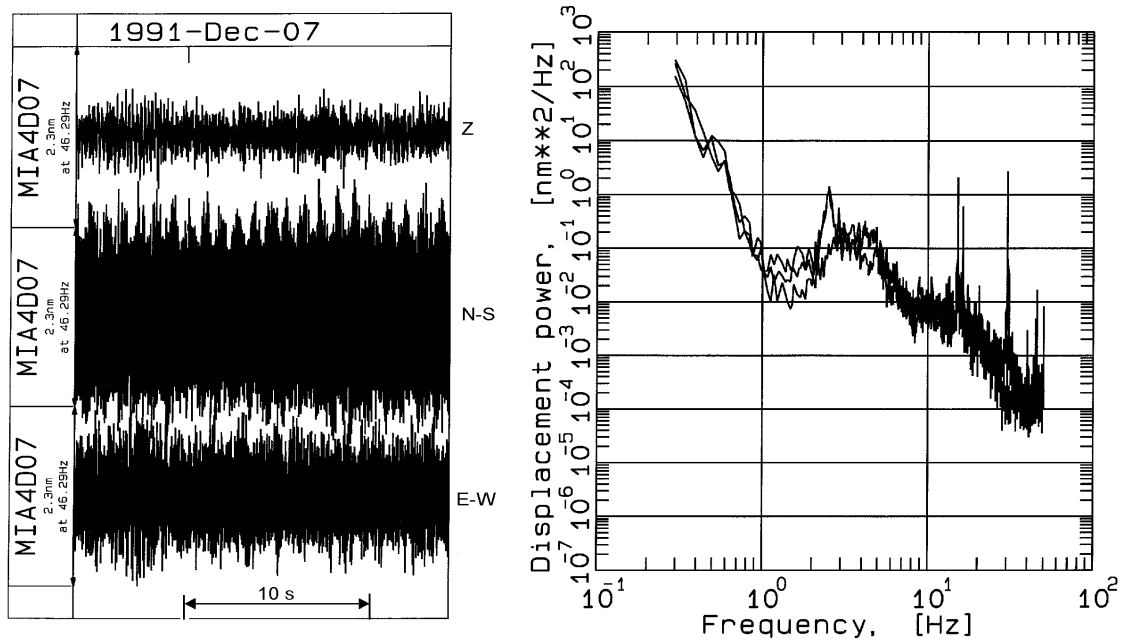
**Fig.7.17** Noise record and related spectra for the first minute (transient) and second minute (background noise). The transient is due to a truck passing by at several 100 m distance from the recording site.

Fig. 7.18 shows a recording at a remote low-noise hard rock site. The first segments are very noisy because people were "stretching their legs" only a few meters away from the sensors. This man-made noise stopped abruptly at 13:06:15 hours when they were asked to sit down and not move. Comparing the related noise power spectra for these two different record segments shows amplitudes 20 to 30 dB lower for the unspoilt ambient noise. Therefore, all members of a noise measurement crew must be instructed to stay away from the sensors and keep very quiet during measurements.



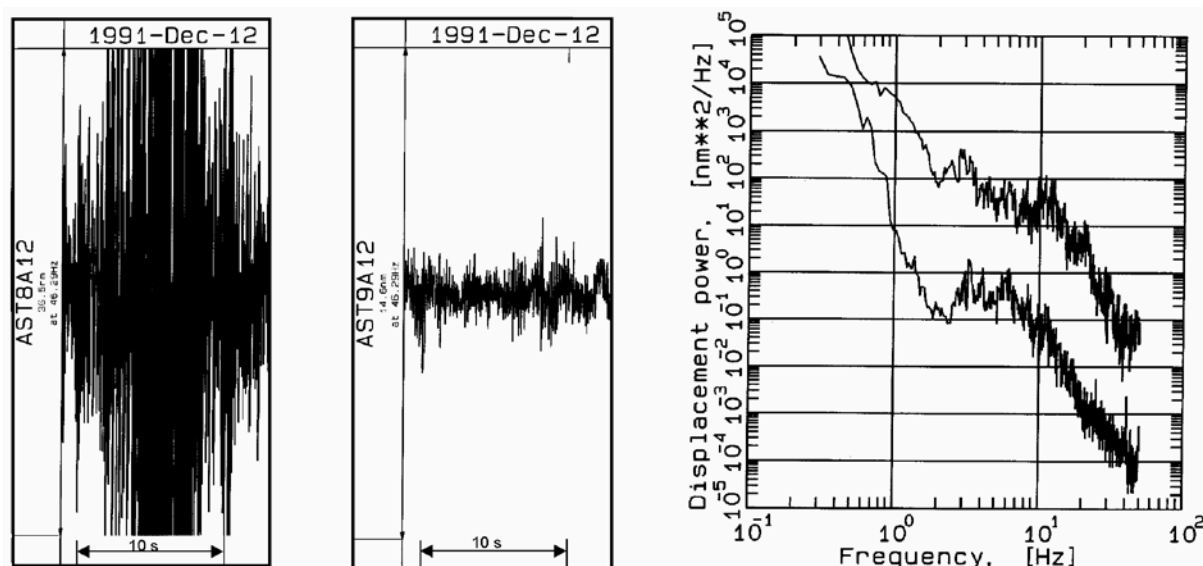
**Fig. 7.18** Noise records (left) and related power spectra (right) at a remote low-noise hard rock site in NW Iran. The large, impulse-like amplitudes in the first part of the record are due to the movement of team members near to the sensors. Note the much lower noise (dots in the spectrum) after they were asked to "sit down and be quiet".

Measurements near power lines and transformer houses likewise may significantly spoil the records. The recordings shown in Fig. 7.19 were made near a quiet countryside village. For frequencies below 13 Hz, the noise amplitudes are roughly the same in the vertical and horizontal components. At higher frequencies, surprisingly, the horizontal records are extremely noisy. The related power spectra show strong, almost monochromatic, noise peaks around 13, 30 and 50 Hz in the horizontal components. (Note that the spectral calculation stopped at the seismometer's upper corner frequency of 50 Hz; see Fig. 7.12). According to the notebook entry and site photograph the record was made only about 30 m away from a transformer house and power line. The strong monochromatic high-frequency noise peaks are probably due to strong electromagnetic induction in the horizontal measuring coils by the AC current frequency of 50 or 60 Hz and its lower harmonics (30 and 13 Hz). However, experience at other sites shows that large transformers and heavily loaded power lines may also vibrate at 50-60 Hz.



**Fig. 7.19** Noise records and related power spectra near to a transformer house and power line. Note the monochromatic spectral lines around 13, 30 and 50-60 Hz, either induced by the AC current frequency and its lower harmonics and/or caused by the vibration of the transformer.

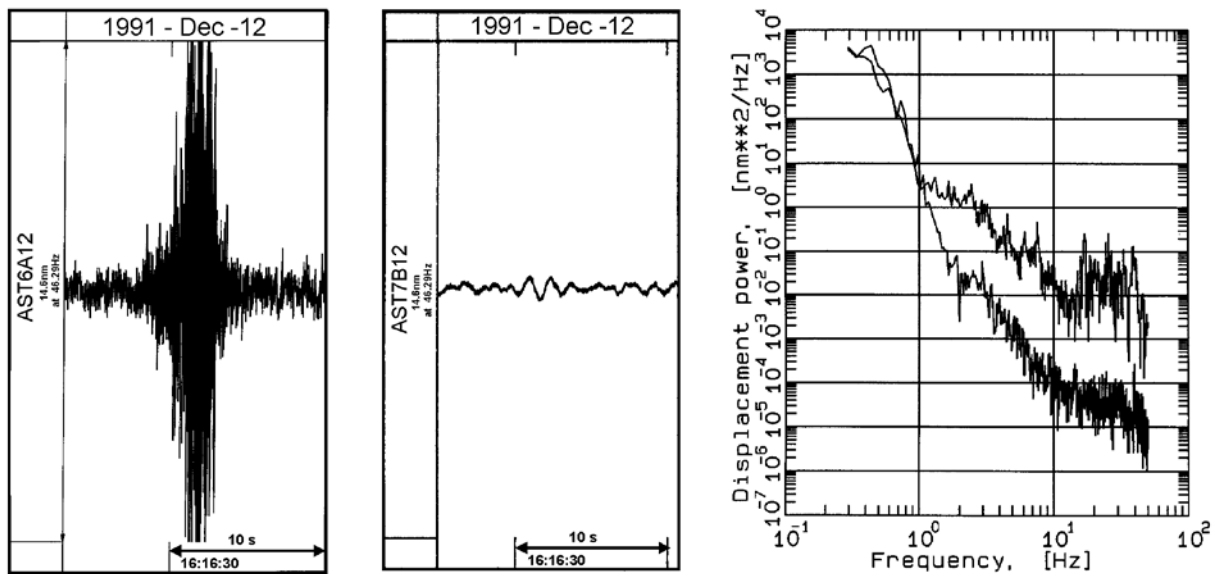
Another experiment demonstrates the attenuation of truck-traffic noise with distance from the road and the influence of the acoustic underground impedance on the recorded spectra. In two different cases, one sensor was placed at the foot of an asphalt-covered road embankment while the other one was installed about 1 km away from the main road in the countryside. In the first case, the underground consisted of wet alluvial coastal plane deposits; in the second case, outcropping competent Cretaceous tuffaceous sandstone, i.e. a rock with a much higher acoustic impedance. The recordings were made simultaneously and the time segments analyzed when a heavy truck was passing by on the main road. On the wet alluvium the vibrations caused by a truck were recorded on the road embankment with very strong amplitudes for almost 30 seconds. Frequencies between 0.3 Hz and 20 Hz were strongly excited. Although power spectral amplitudes at 1 km distant were generally 20 to 30 dB lower, high frequencies were still clearly visible in the record and the spectrum (Fig 7.20).



**Fig. 7. 20** Comparison of seismic records and related noise spectra made at the time of passing of a heavy truck: Left: record near the road embankment; middle: record made about 1 km away from the road in the countryside (middle); right: noise power density spectra. Underground: wet Alluvial coastal plain deposits. (Note that the noise amplitudes in the left panel have been reproduced with only 40 % of the magnification in the central panel).

In contrast, Fig. 7.21 shows the records made at another section of the road embankment consisting of broken rock overlaying outcropping competent rock. A strong increase in noise amplitudes above the general background level was observed for about 5 s only, i.e., when the truck was close to the site. The general noise level, even at the time of the passing truck, was 20 to 30 dB lower than on the alluvial embankment. Also, at the broken/compact rock road embankment, spectral amplitudes for frequencies between 0.3 and 1.5 Hz were about the same as 1 km away in the side valley on the outcropping compact sandstone. On the other hand, high frequency amplitudes generated by the truck are no longer visible in the record at the hard rock site 1 km away from the main road and reduced by 20 to 30 dB in the power spectrum.

In summary, these examples show what one can expect for noise reduction with distance from main traffic roads or other sources of man-made noise, and their dependence on underground conditions. This may help guide reconnaissance field measurements for appropriate and accessible sites. The examples also illustrate the usefulness of comparing noise records in the time domain with the related power spectra in order to better identify the kind of noise sources and understand their appearance in the records.



**Fig. 7.21** As Fig. 7.20, except that records were made near a broken rock embankment of a main road and on outcropping compact tuffaceous sandstone, 1 km away in a side valley, respectively.

## 7.2.3 Comparison of noise and signals at permanent seismological stations

### 7.2.3.1 Introduction

Existing permanent seismological stations have historically been established by different institutions for different reasons and have often been installed under different underground and environmental conditions. The stations were usually operated independently, each reporting their own data readings to national or international data centers. Modern methods of data communication make it easy to link these stations, to merge them into virtual networks (see 8.4.3), to exchange waveform data in real time and to perform joint data analysis at local, national or regional data centers. The overall network performance and quality of results strongly depends on the local conditions at the individual stations. One crucial parameter is the detection threshold. This is mainly (but not exclusively) controlled by the noise conditions at the sites. High noise conditions at some stations reduces their contribution to event detection, discrimination and location accuracy of the network, may bias average network magnitude estimates and may result in inhomogeneous completeness and accuracy of earthquake catalogs. Therefore, when setting up new seismic networks or linking already existing stations into a network, a priority task should be to investigate and compare the signal-to-noise conditions at the various stations, and to find alternatives for inferior sites. Such decisions may have far-reaching consequences and involve significant cost and so should not be based on just a few short-term noise measurements in a limited frequency band. Noise measurements should be taken over at least several days, but preferably over weeks or even months, in order to get a clear understanding of the diurnal and seasonal variability of seismic noise in the full frequency band of interest for the operation of the network. Moreover, one should determine the signal-to-noise ratio (SNR) for events from different distance and azimuth ranges and compare this at existing and possible alternative sites. It is vital that all records should be made with equipment having an identical instrument response.

This is demonstrated using data from the German Regional Seismic Network (GRSN) (see Fig. 8.15). Originally the GRSN consisted of 12 sites in western Germany. Several permanent stations in eastern Germany were subsequently added to the network. The GRSN now consists of 16 digital broadband stations equipped with STS2 seismometers (see DS 5.1), 24-bit data loggers and a seismological data center at the Gräfenberg BB array center (GRFO) in Erlangen. The network covers the whole territory of Germany with a station-spacing between 80 km and 240 km. The stations are located in very different environments: e.g., near the Baltic Sea coast (HAM and LID, now BSEG; RGN); up to distances of about 700 km away from the coast (FUR); within cities (BRNL, HAM) or up to about 10 km away from any major settlement, industry or busy roads. The underground varies from outcropping Paleozoic hard rocks in Hercynian mountain areas (BFO, BRG, CLL, CLZ, GERES, MOX, TNS, WET), sedimentary rocks in areas of Paleozoic (BUG, IBBN) or Mesozoic platform cover (GRFO, STU) to unconsolidated Pleistocene (glacial) deposits (BRNL, HAM, FUR, LID, RGN). The seismometers are installed either at surface level (CLL, CLZ, HAM, IBBN, RGN, WET), in shallow vaults just a few meters below the ground surface (BUG, FUR, GSH, TNS), in boreholes (GRFO, 116 m), or in bunkers, tunnels or abandoned mines between 20 and 162 m below surface (STU, MOX, BSEG, BRG, RUE, BFO). More details about these stations and their equipment can be found on the Internet at [http://www.szgrf.bgr.de/station\\_map.html](http://www.szgrf.bgr.de/station_map.html), and <http://www.seismologie.bgr.de>.

Seismic background noise at GRSN stations varies in a wide range between the upper and lower bounds of the new global noise model (see Fig. 7.27). The noise conditions at the GRSN have been investigated in detail in the frequency range from  $10^{-2}$  to 40 Hz by Bormann et al. (1997).

### 7.2.3.2 Data analysis

Continuous recordings at all stations were systematically screened at different times of the day (0, 6, 12 and 18 hrs UT) in order to reveal diurnal variations and their site dependence. Records were also monitored throughout the year in order to identify periods of minimum and maximum noise level and their seasonal variations. Respective record sections and related power spectral densities (PSD) were plotted together and checked for transient signals from seismic or other spurious events.

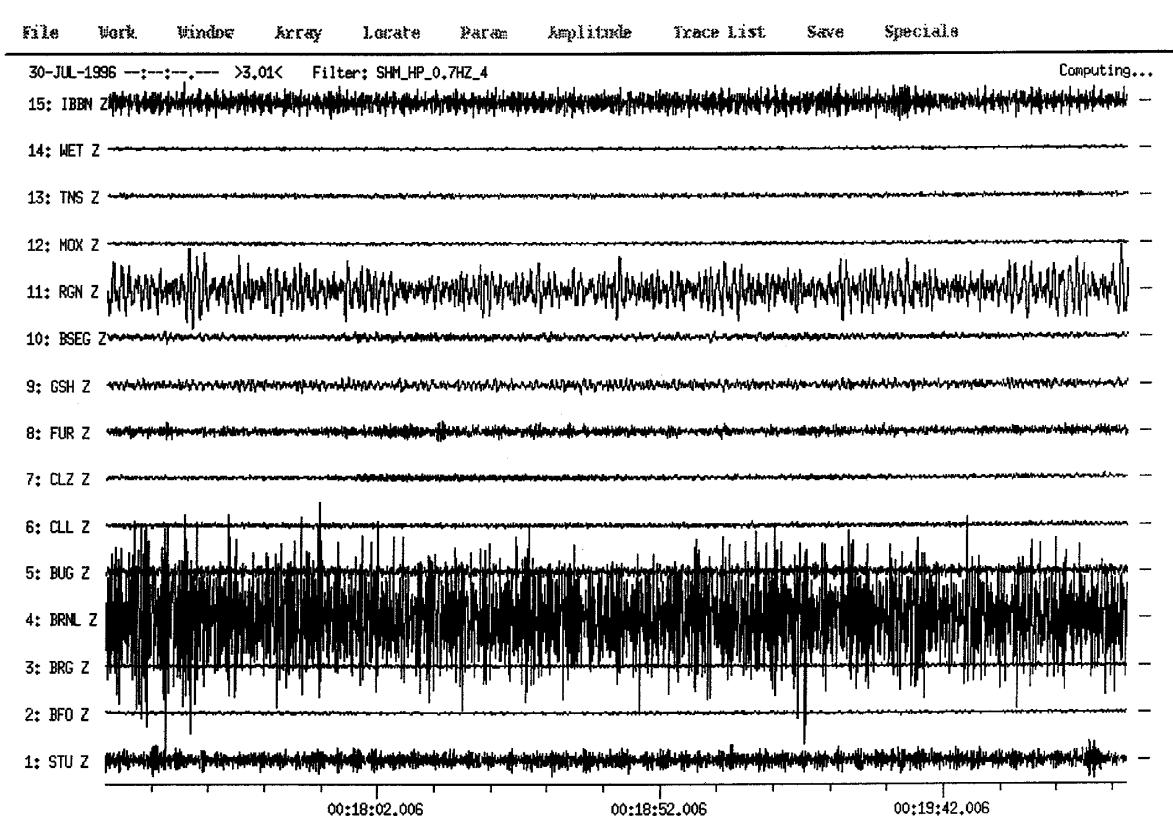
Data of the GRSN are acquired at a sampling rate of 80 Hz for most stations and 20 Hz at the more noisy stations. For most of the routine noise analysis, the 80 Hz data were re-sampled at 20 Hz. The Power Spectral Density (PSD) was calculated using a subroutine from the program SEIS89 (Baumbach 1999). It implements, in a somewhat modified form, an algorithm recommended as a standard for the calculation and presentation of noise spectra by the Ad Hoc Group of Scientific Experts (1991). The modification allows the use of segments of data larger than 512 samples, thus permitting the analysis of more long-period noise. The digital time series containing background noise are divided into a number of half-overlapping record segments, normally of 4096 samples. The power spectra are then calculated for each segment (after removing the mean and tapering the ends of each segment with a sine-cosine window) and then averaged over eight segments in order to reduce the variance of the PSD estimate. Accordingly, the presented power spectra are representative for noise records of about 15.4 min duration in case of 20 s.p.s. and of about 3.8 min duration for 80 s.p.s.. All spectra are corrected for the instrument response. The power spectra are presented in units of displacement power spectral density in  $\text{nm}^2/\text{Hz}$ . A lower frequency limit is imposed such that the longest period which can be analyzed using this procedure is one sixth of the segment length.

According to Fig. 7.48 in section 7.4.4, STS2 seismographs have a self-noise which is below the global New Low-Noise Model between about  $10^{-3}$  Hz and 10 Hz. According to Wielandt and Zürn (1991), they can resolve the noise at BFO, which is one of the quietest seismic stations in Germany, for frequencies below 30 Hz. Thus, instrumental and/or digitization noise can potentially affect the noise estimates at the best sites only at frequencies above and below this range.

Essential results of the analysis are presented below. Figs. 7.22 - 7.37 are reproduced from Journal of Seismology, Vol. 1, 1997, pp. 357-381, "Analysis of broadband seismic noise at the German Regional Seismic Network and search for improved alternative station sites" by P. Bormann, K. Wylegalla and K. Klinge, Figures 2, 4, 6-7, 9, 11-15, 17-20 and 22; © 1997( with kind permission from Kluwer Academic Publishers).

### 7.2.3.3 Results

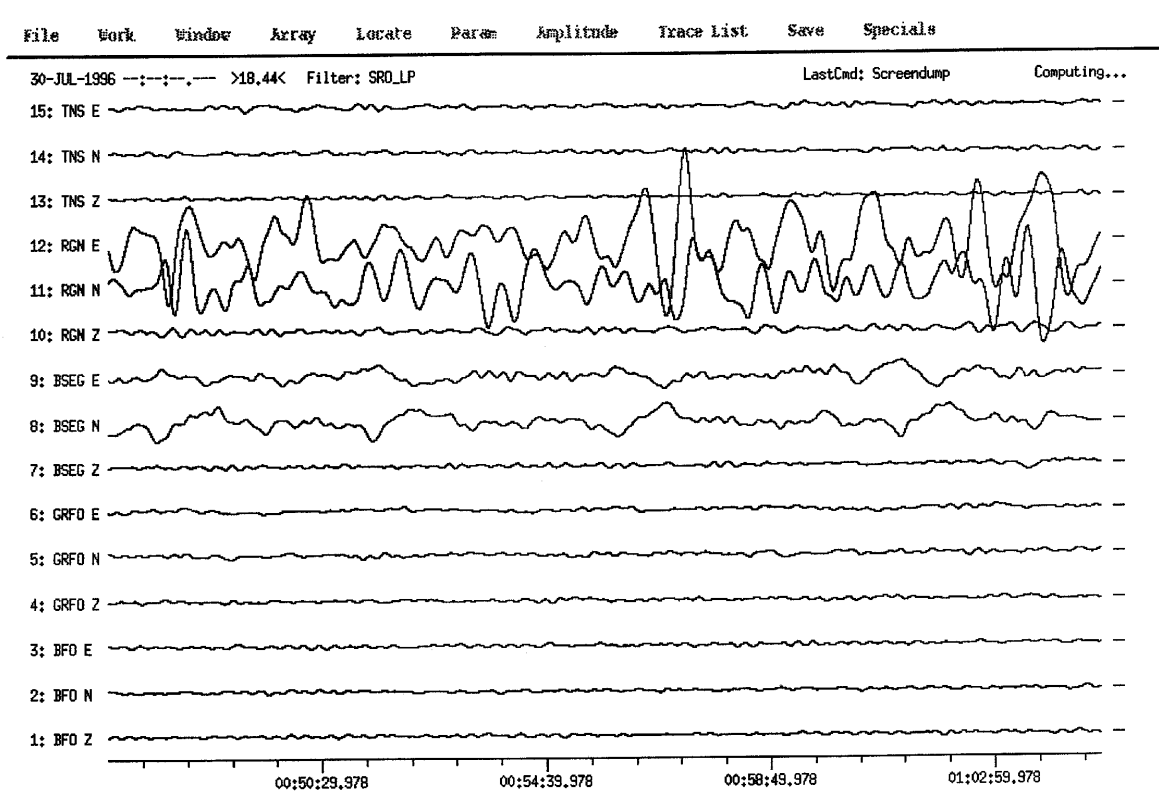
Fig. 7.22 shows an example of high-pass filtered short-period Z-component records of seismic background noise from 15 stations of the GRSN. Amplitudes differ by more than one order of magnitude. Noise amplitudes on vertical and horizontal recordings were about the same at any given station. Therefore, only spectra from Z-component records are considered. In long-period records, however, horizontal noise is sometimes significantly larger (e.g., for stations RGN and BSEG in Fig. 7.23), due to the high tilt sensitivity of long-period horizontal seismometers (see 5.3.3).





**Fig. 7.22** High-pass filtered ( $f_c = 0.7$  Hz) Z-component noise records of GRSN stations on July 30, 1996, at night time (from Bormann et al., 1997).

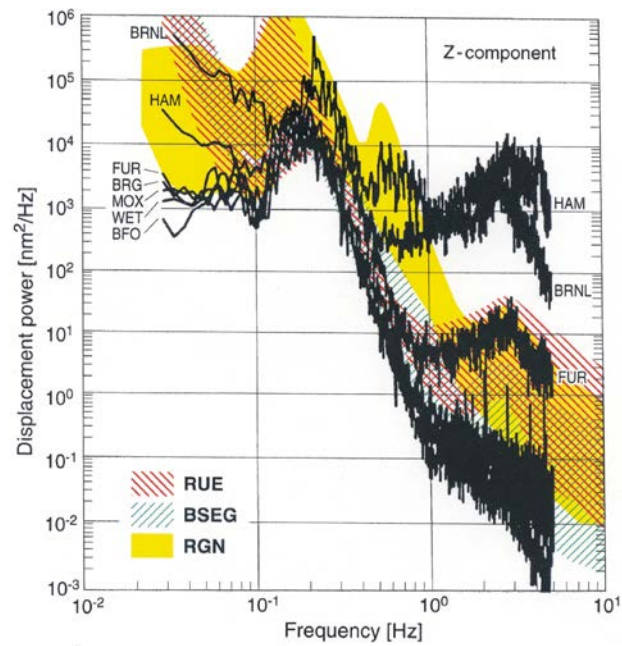
On very calm days at stations with very good environmental shielding (e.g., BFO, GRFO, TNS in Fig. 7.23), horizontal long-period noise might be equal to or only somewhat stronger than in vertical components. On stormy days with high wind pressure fluctuations and related tilts, however, the noise power in near-surface horizontal recordings might be 20 to 30 dB higher than in vertical ones. When the sensors are installed sufficiently deep in boreholes (as GRFO; 116 m below surface) or in mines (as BFO; 162 m below surface) this difference will be much less, even during stormy days.



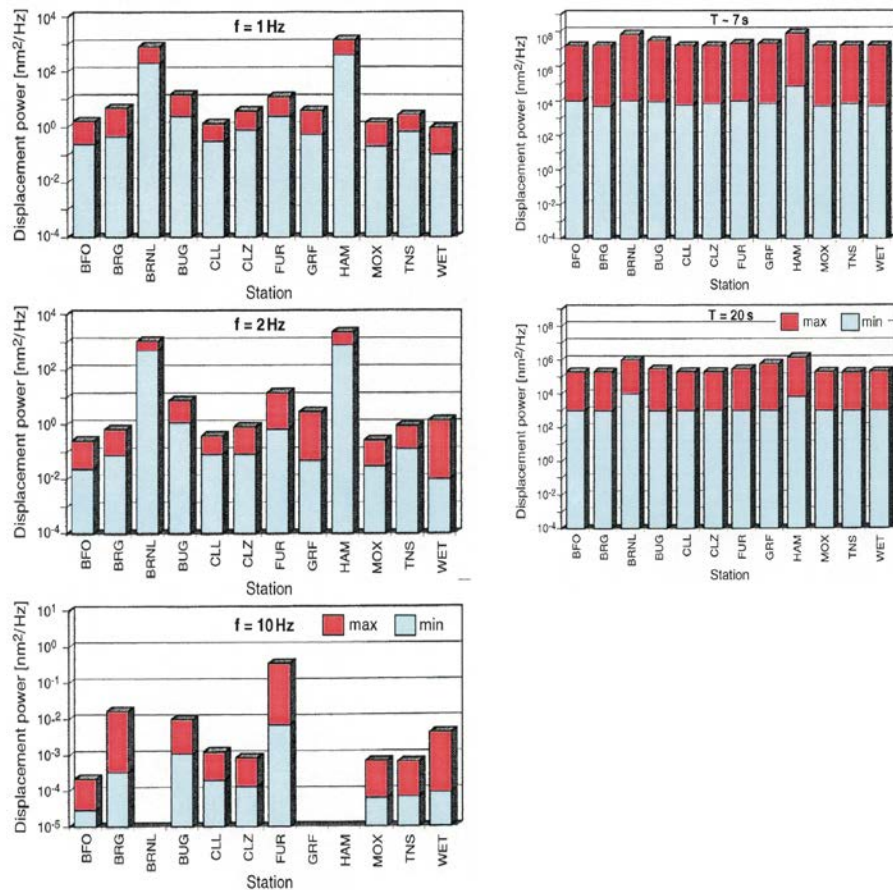
**Fig. 7.23** Three-component recordings at five GRSN stations after applying a long-period SRO filter characteristic (from Bormann et al., 1997).

Differences in the displacement PSD at the GRSN stations are most obvious for frequencies above 0.5 Hz. They may reach about 60 dB (Fig. 7.24) and are due to the varying proximity to man-made noise sources and differences in underground conditions. The stations BRNL (Berlin Lankwitz) and HAM (Hamburg) proved to be the worst sites. For longer periods ( $T > 2$  s) the differences in noise level between the GRSN stations are much less pronounced; less than 10 dB in most cases. However, over a long period of time (Fig. 7.25) the noise power variability at individual stations of the GRSN proved to be smallest (and seasonally independent) around  $f = 1$  Hz (about 5 to 10 dB variation only). It is larger between 2 to 10 Hz (up to about 20 dB) and largest for the secondary ocean-storm microseism peak around 7 s period (30 to 40 dB). Microseisms only occur episodically and with seasonally varying intensity (strongest at the time of winter storms). At periods around 20 s, the range of noise power variations still reaches 20 to

30 dB. This is equivalent to variations in the magnitude threshold for  $M_s$  determinations of up to 1.5 magnitude units.

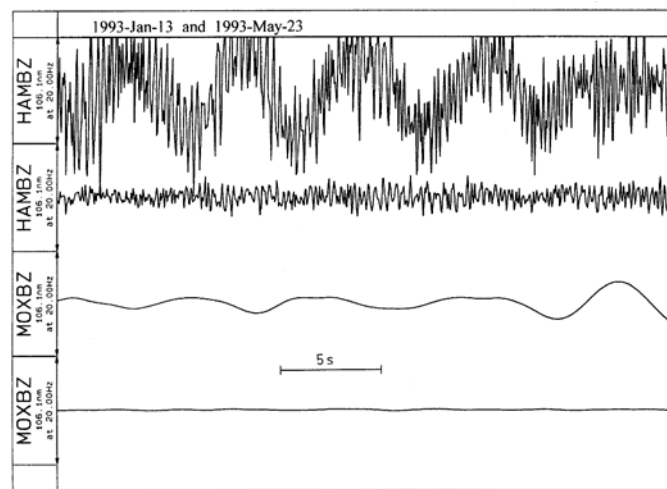


**Fig. 7.24** Displacement power density spectra at selected GRSN stations determined from noise records on the morning of April 13, 1993. For comparison the ranges of noise power observed at the new sites BSEG, RGN and RUE are given as shaded areas (from Bormann et al., 1997).

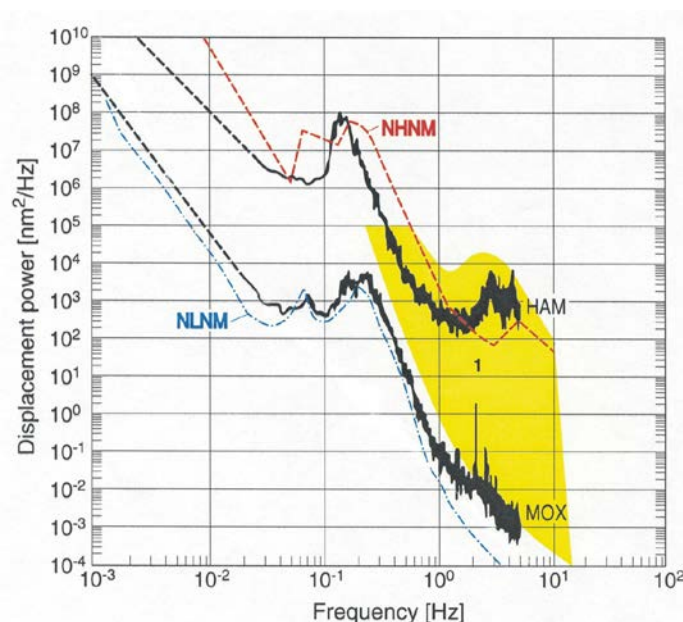


**Fig. 7.25** Comparison of the minimum and maximum levels of short-period and long-period seismic noise power observed at GRSN stations (modified after Friedrich, 1996; from Bormann et al., 1997).

Fig. 7.26 shows record sections of only 1 minute duration and with identical gain for one of the quietest and one of the noisiest days observed during a year at each of the stations MOX and HAM. The amplitudes of secondary ocean-storm microseisms with periods of about 6 to 7 s, on the noisy day, are at HAM only about twice as large as at MOX despite HAM being much closer to their origin along the European North Atlantic coastline. On the other hand, the high-frequency noise at HAM is always much larger than at MOX. The corresponding displacement power spectra for the quietest day at MOX (May 23) and the noisiest day at HAM (January 13) during 1993 are compared in Fig. 7.27 with the global New Low Noise Model (NLNM) and New High-noise Model (NHNM) according to Peterson (1993).



**Fig. 7.26** Comparison of record segments with largest (13 January) and lowest seismic background noise (23 May) observed in 1993 at stations HAM (upper two traces) and MOX (lower two traces) (from Bormann et al., 1997).



**Fig. 7.27** Spectra for the noisiest day observed at HAM (January 13) and the quietest day at MOX (May 23) during 1993. The NHNM and NLNM according to Peterson (1993) are shown for comparison. The shaded area (1) covers the range of short-period noise power calculated by Henger (1995) for all GRSN stations on March 1, 1994 (modified from Bormann et al., 1997).

The diurnal variations of man-made noise have also been investigated at all stations of the GRSN. The variations are very distinct (20 to 30 dB) at the stations BRNL, BUG and FUR, i.e. at sites in densely populated areas and with thick unconsolidated subsoil. They are much less (< 5 to 10 dB) at stations on hard rock in smaller and less busy towns (such as BRG and CLZ) or even at several km distance to the nearest villages (CLL, MOX and TNS).

Due to the large differences in noise conditions at the GRSN stations, the capability to detect and locate events with at least 3 stations was rather inhomogeneous over German territory. The detection thresholds ranged between  $M_l = 1.5$  and 3. Since the network was supposed to detect and localize all local events with  $M_l \geq 2$ , more suitable sites had to be found for some stations. This was particularly true for BRNL and HAM. The search for more appropriate alternative sites focused on areas not too far away from these stations in order to preserve the general configuration of the GRSN.

## **7.2.4 Searching for alternative sites in a given network**

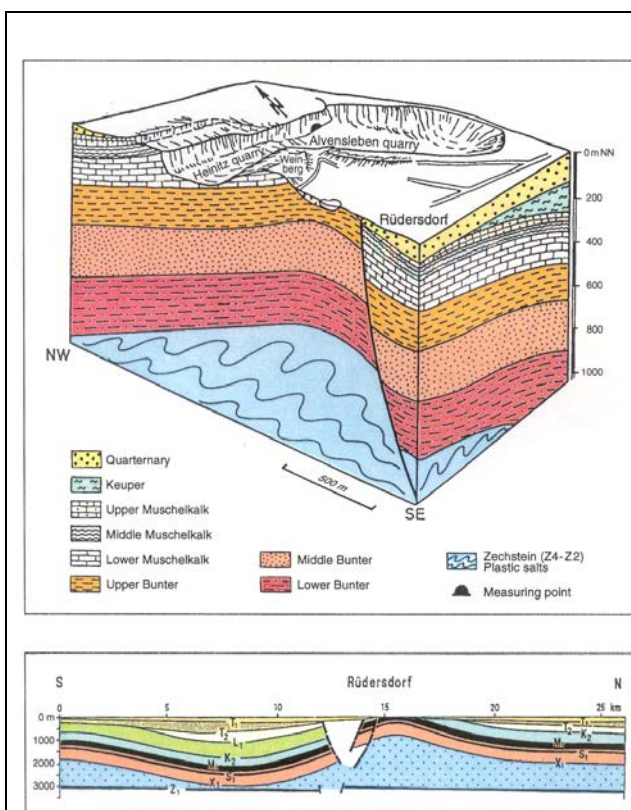
### **7.2.4.1 Geological and infrastructure considerations**

We consider here two case studies for replacing the seismic stations BRNL and HAM.

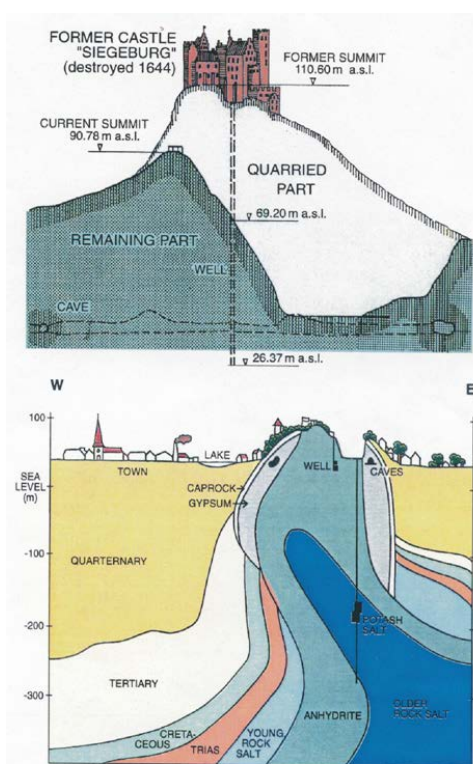
BRNL was located on the courtyard of the Geophysical Institute of the Free University of Berlin, about 12 km from the city center. The station underground consists of about 290 m unconsolidated Cenozoic sediments overlaying a thick sequence of Mesozoic sedimentary rocks. These unfavorable underground conditions, together with high population and nearby traffic density, made this station one of the noisiest in Germany. An alternative site had to be found in the wider surroundings of Berlin.

In the area of Berlin, the base of the Permian Zechstein subdivision is between about 2600 m and 4000 m below sea level. The pre-Permian basement is block-faulted with different vertical movements between adjacent blocks during post-Permian times. This mobilized the overlying plastic salt deposits of the Zechstein subdivision and resulted in the formation of dozens of salt-pillow structures, up-doming the Mesozoic sequences above. In a few cases, salt diapirs pierced through the post-Permian deposits to the present surface. The largest of these halokinetic structures exists beneath the small town of Rüdersdorf (Fig. 7.28) about 25 km east of the city center of Berlin. It was exposed by Pleistocene glacial erosion, thus forming the northernmost natural outcrop of Middle Tertiary limestones in Germany which has been mined for hundreds of years. Logistically, Rüdersdorf is easy to reach and has all the power and telecommunication connections needed for a GRSN station. The open-cast development stretches E-W and is about 0.5 to 4 km away from the eastern segment of the busy "Berliner Ring Autobahn" (motor highway). Despite the proximity to town and highway and the continuing surface mining in the quarries of Rüdersdorf, this area was considered to be the most promising alternative for the station BRNL both from a seismo-geological and logistical point of view. This was subsequently confirmed by measurements (see 7.2.4.3).

Hamburg is situated in the NW of the North German-Polish Depression. The regional geological conditions are similar to those around Berlin although the depression is much deeper here. The unconsolidated sediments above the basis of Tertiary are about 1.5 km thick beneath the station. HAM was situated about 12 km away from the city center but rather near ( $< 1$  km) to different segments of the dense highway network. Accordingly, the noise conditions were the worst of all the seismic stations in Germany. The most promising alternative site was on an outcropping, partially mined, salt diapir in the town of Bad Segeberg, about 50 km NNE from the center of Hamburg, not too close to either the North Sea or Baltic Sea, easily accessible and with suitable infrastructure and communications facilities. There are Quarternary unconsolidated sediments, about 100 to 400 m thick, and Cretaceous and Triassic sedimentary rocks at a few hundred meters depth, adjacent to the diapir. Fig. 7.29 shows a schematic cross section through the former castle hill and the upper few hundred meters of the diapir of Bad Segeberg.



**Fig. 7.28** Cross sections through the salt-tectonic up-doming at Rüdersdorf at local (above) and regional scale (below) (from Bormann et al., 1997).



**Fig. 7.29** Cross-section through the former castle hill of Bad Segeberg (above) and the related geological profile of the Permian salt diapir (below) (from Bormann et al., 1997).

#### 7.2.4.2 Recording conditions and data analysis of temporary noise measurements for alternative permanent broadband stations

Identical very broadband STS2 seismometers were used with PDAS digital data loggers for comparative measurements of seismic background noise at BRNL and with their potential alternative station sites RUE and BSEG. The data were sampled at 100 Hz. The seismometer at RUE was placed in a small tunnel in the quarry in order to reduce the influence of temperature



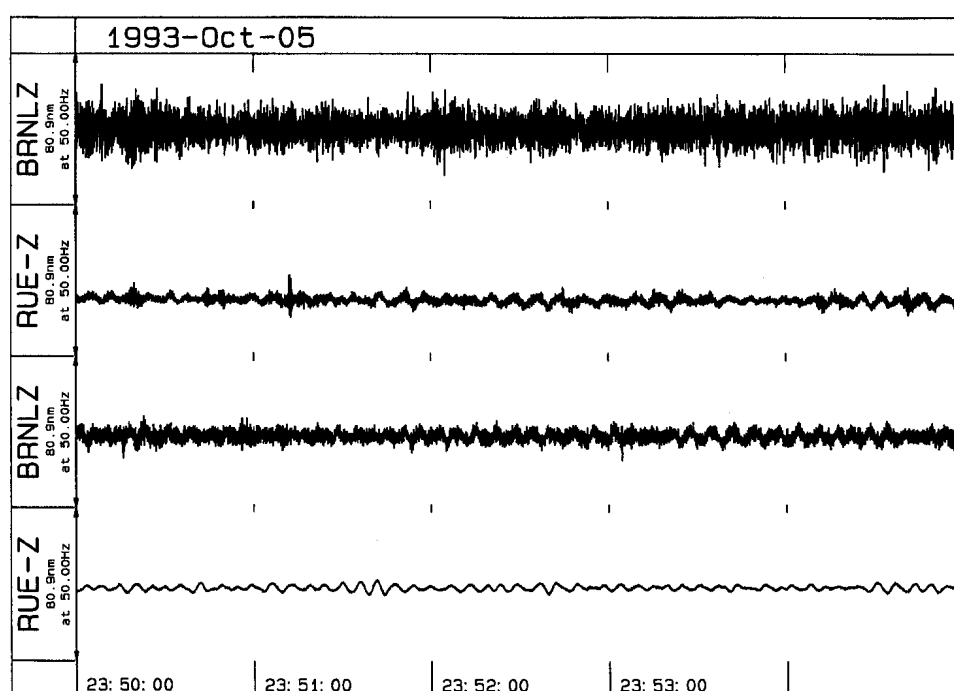
variations and to enable stable broadband recordings. The tunnel was about 10 m long, with 55 m of limestone overburden, and the site was 2 to 3 km away from the highway and the village of Rüdersdorf. At BSEG, the STS2 was installed in a gypsum cave within the diapir caprock of Bad Segeberg, about 20 to 30 m below the surface. The cave is only a few hundred meters away from the town center of Bad Segeberg.

In both cases the instruments were placed directly on a levelled hard rock surface. No additional thermal or pressure shielding was provided during the temporary measurements apart from the manufacturer's standard metallic sensor platform with cover hood. Therefore, in the data shown below, the long-period noise at RUE and BSEG is higher than it would be in a good permanent installation. Note that in contrast to temporary noise measurements with short-period seismometers, broadband sensors require about one day to adapt to the environmental conditions and find a stable zero position. Meaningful data can only be acquired after this.

For several days, continuous noise and signal measurements were carried out at BSEG and RUE parallel to HAM and BRNL, respectively. Data sampled at 100 Hz. were used for the determination of displacement noise power between 0.1 and 50 Hz and re-sampled 20 Hz data were used for the range 0.03 to 5 Hz. The PSD subroutine described in 7.2.3.2 was used, with a basic record length of 4096 samples. The average power spectrum was determined using 25 consecutive segments with 50% overlap. Thus the spectra are representative for noise records of 8.87 min and 44.37 min length depending on whether they are based on data sampled at 100 or 20 Hz.

#### 7.2.4.3 Results of noise and signal measurements at BRNL and RUE

Fig. 7.30 shows unfiltered 5-minute broadband segments of noise recordings at BRNL and RUE taken around noon and around midnight. Fig.7.31 shows the noise power at both sites in the frequency range 0.03 to 50 Hz.

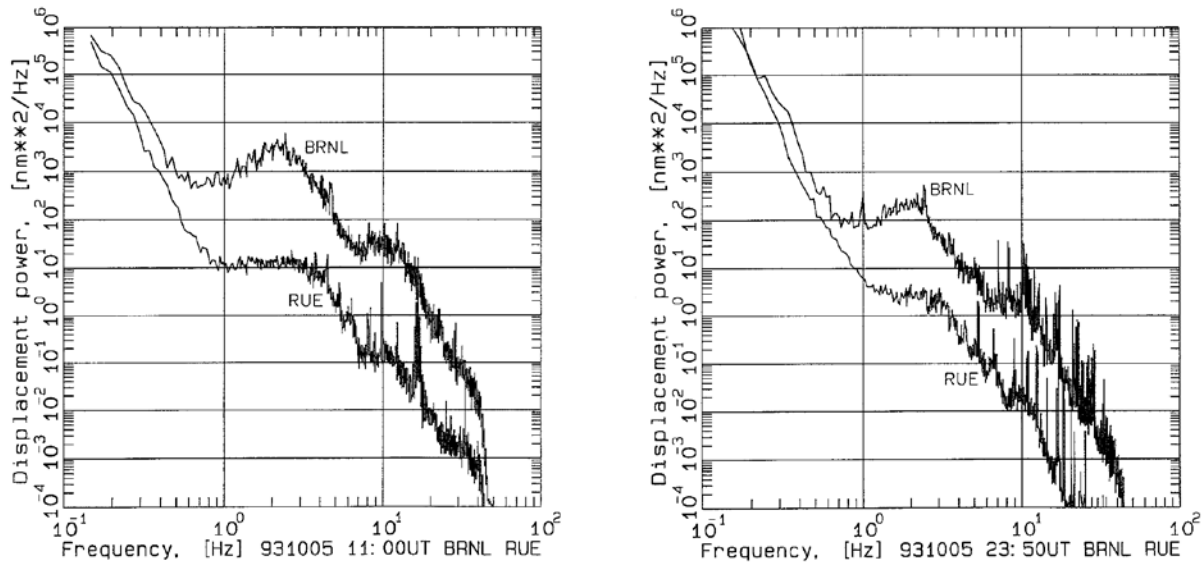


**Fig. 7.30** Unfiltered Z-component broadband records of seismic noise with identical resolution at BRNL and RUE. Upper traces: 11:50 - 11:55 UT; lower traces: 23:50 to 23:55 UT (from Bormann et al., 1997).

The comparison reveals that:

- the noise above 1 Hz at BRNL is some 15 to 25 dB higher than at RUE, both at day- and night-time;
- between 1 and 5 Hz the night-time noise is less than the day-time noise by about 10 dB at BRNL and by about 6 dB at RUE;
- below 0.5 Hz, BRNL has about the same noise power level as RUE with negligible diurnal variation at both sites;
- a range of different, spatially distributed random noise sources such as nearby traffic seem to dominate the short-period noise during day-time at both sites. This results in a rather high and "smooth" noise spectrum without any dominating spectral lines at BRNL and only a few sharp spectral lines at RUE (e.g. at  $f = 8, 10, 16$  and  $32$  Hz);
- during night-time, when the traffic noise is reduced, several sharp spectral lines become dominant for  $f > 5$  Hz at both BRNL and RUE. These are probably due to specific noise sources such as machinery rotating with constant frequency (and their lower and higher modes).

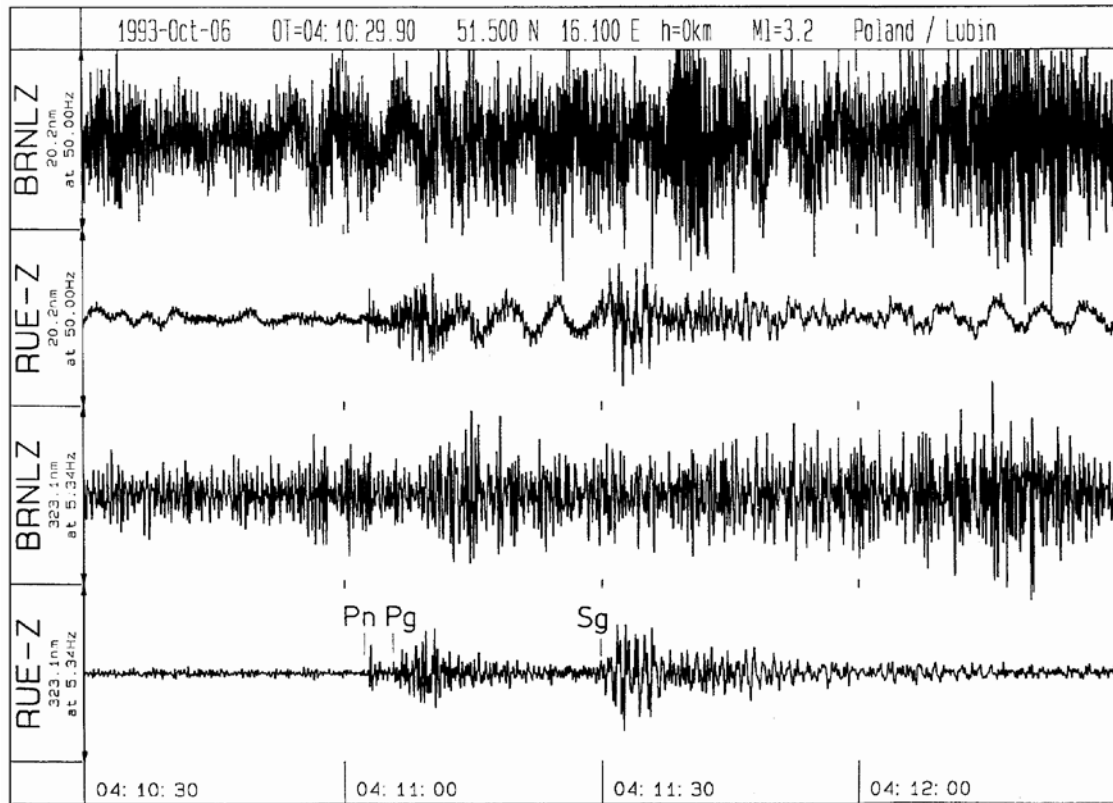
The last of these observations is clearly related to activity in the Rüdersdorfer quarry. Mining and stone crushing machinery are operating there throughout the day. Despite the generally lower noise level at RUE compared to BRNL, it is meaningful only to record at RUE low-pass filtered data ( $f_c = 5$  Hz) sampled at 20 Hz. According to Fig. 7.24 the noise power at RUE is comparable with that at station Fürstentfeldbruck (FUR), a site of intermediate quality. A better result is not achievable with a near-surface installation in the surroundings of Berlin.



**Fig. 7.31** Power spectra of seismic noise in Z-component broadband records at BRNL and RUE around noon (left) and midnight (right) (from Bormann et al., 1997).

Fig. 7.32 presents the broadband (top) and band-pass filtered (from 0.5 - 5 Hz, bottom) Z-component records at BRNL and RUE of a nearby event at approximately the same distance. In

both cases the event is not visible at BRNL but is clearly recorded at RUE with several distinct wave groups. The spectral signal-to-noise ratio (SNR) of this event is  $\leq 1$  at BRNL and varies between 3 and 30 at RUE for  $0.5 \text{ Hz} < f < 7 \text{ Hz}$ . This is a significant improvement of recording conditions. As a consequence, station BRNL was closed and its equipment permanently moved to RUE.



**Fig. 7.32** Unfiltered broadband (upper two traces) and band-pass filtered ( $f = 0.5 - 5 \text{ Hz}$ ; lower two traces) Z-component records of a near seismic event in Poland at BRNL ( $D = 214 \text{ km}$ ) and RUE ( $D = 191 \text{ km}$ ) (from Bormann et al., 1997).

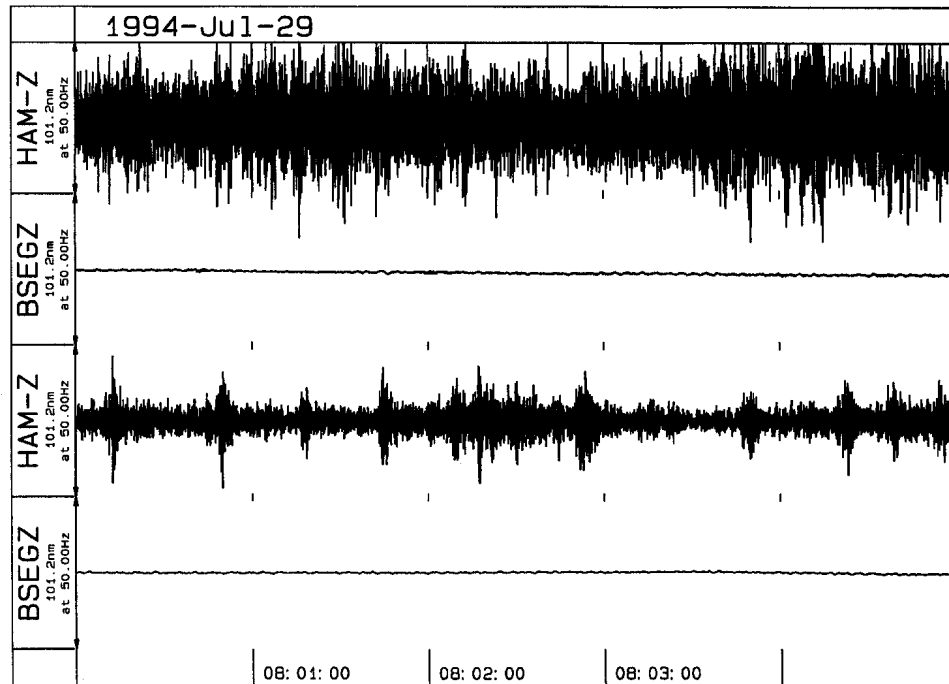
#### 7.2.4.4 Results of noise and signal measurements at HAM and BSEG

Fig. 7.33 shows an example of day-time and night-time noise records at HAM and BSEG with identical resolution and Fig. 7.34 shows the related power spectra. The comparison, including that with spectra from other days and with Fig. 7.24, shows that:

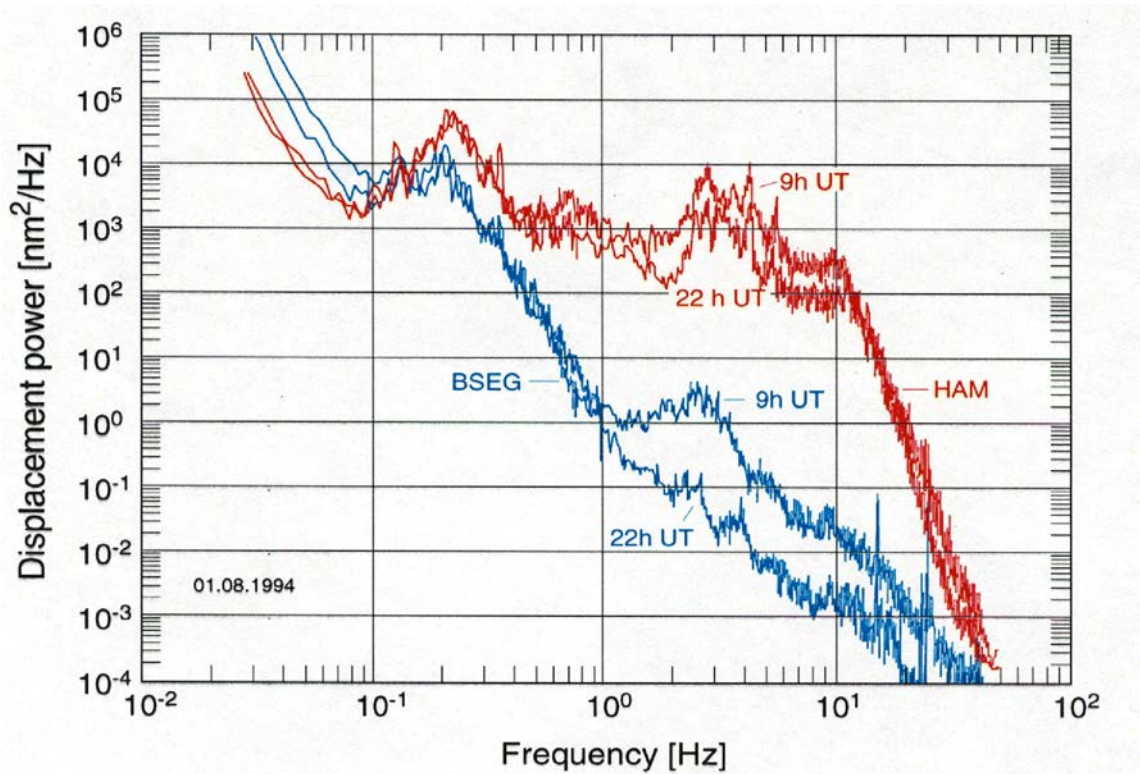
- diurnal variations in seismic noise are remarkably small ( $\leq 10\text{dB}$ ) at HAM. The cause is very intense traffic and industrial activity in this busy large harbor town that does not vary much between day and night time.
- diurnal variations are significant (about 10 to 20 dB) at BSEG above 1.5 Hz but negligible below 1 Hz;
- between 0.5 and 40 Hz the noise power at BSEG is about 20 to 50 dB smaller than at HAM;
- for medium-period ocean storm microseisms (around 3 to 5 s period) the noise power is reduced by about 10 dB at BSEG;



- there is sometimes larger long-period noise at BSEG compared to HAM. This mainly non-seismic noise was significantly reduced after final installation and the level is now comparable with other good GRSN sites;
- noise conditions at BSEG above 1 Hz are only slightly inferior ( $\leq 10$  dB) to good hard-rock sites of the GRSN.

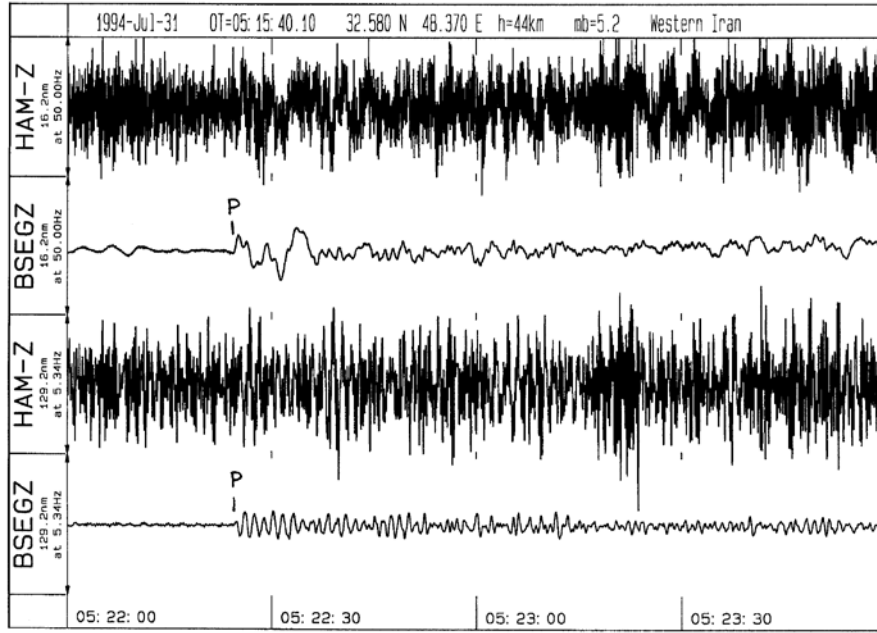


**Fig. 7.33** Five minutes of unfiltered Z-component broadband records at HAM and BSEG on July 29, 1994 at 8:00 UT in the morning (upper two traces) and after midnight (lower two traces) (from Bormann et al., 1997).

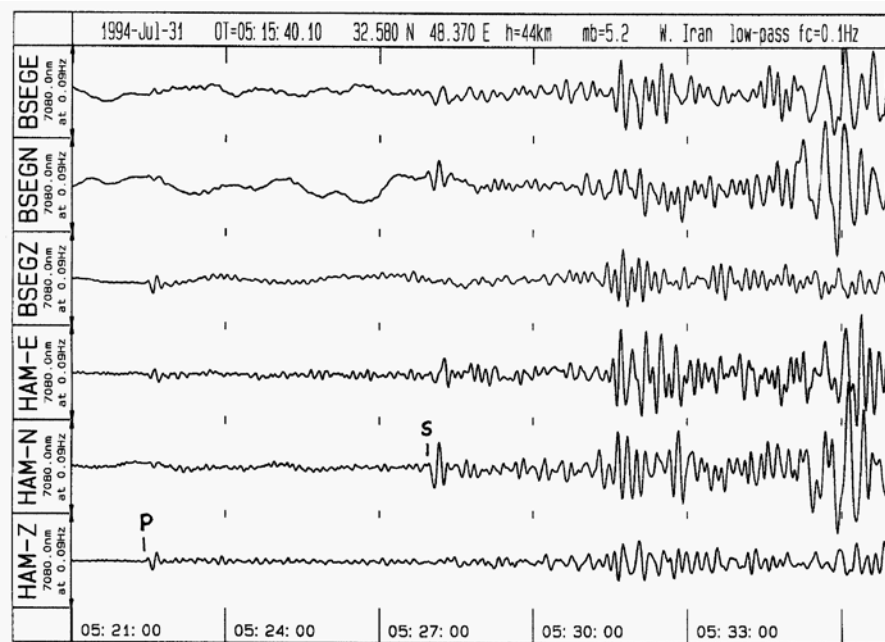


**Fig. 7.34** Noise power spectra at HAM (upper two curves) and BSEG (lower two curves) determined from Z-component records on August 1, 1994, around 9 h UT and 22 h UT, respectively (from Bormann et al., 1997).

Fig. 7.35 shows the Z-component broadband and short-period records at BSEG and HAM of a teleseismic event in Iran. The event was not recognizable at HAM but was recorded very well at BSEG. In contrast, the SNR for the P-wave onsets in long-period filtered records (Fig. 7.36) was comparable at HAM and BSEG since the P-wave wavelengths are  $> 50$  km and therefore much larger than the size of the noise-reducing velocity anomaly of the diapir structure at Bad Segeberg.



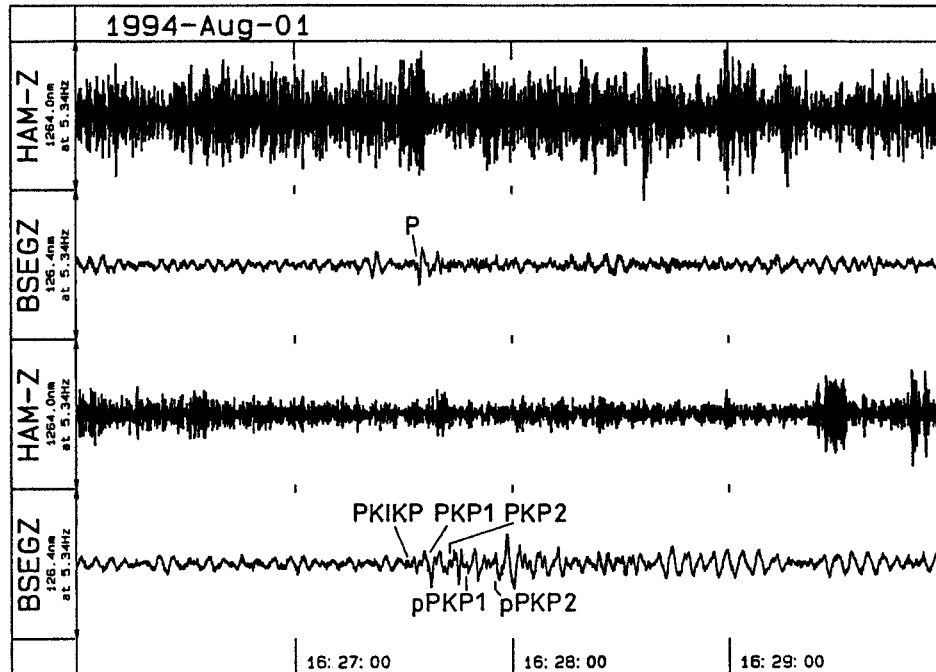
**Fig. 7.35** Z-component records of an earthquake in Iran (distance about 3800 km) at HAM and BSEG. Upper two traces: unfiltered broadband records; lower two traces: band-pass filtered with  $f = 0.5\text{--}5$  Hz (from Bormann et al., 1997).



**Fig. 7.36** Low-pass filtered ( $f_c = 0.1$  Hz) long-period 3-component records at BSEG and HAM of the Iran earthquake (from Bormann et al., 1997).

Two more examples of relatively weak ( $mb = 5$ ) earthquakes recorded at about  $76^\circ$  and  $150^\circ$  distance are shown in Fig. 7.37. Although the record traces for HAM have been reproduced at a resolution 10 times lower than the BSEG records the noise amplitudes are still much larger. The P and multiple PKP onsets (including depth phases) can be picked easily in the short-period filtered records of BSEG but not at HAM.

BSEG has now replaced HAM as a permanent GRSN station. Together with RUE, this has significantly improved the GRSN network detection and location performance for events in the northern part of Germany.



**Fig. 7.37** Short-period band-pass filtered Z-component recordings ( $f = 0.5 - 5\text{Hz}$ ) at HAM and BSEG. Upper two traces: P-wave onset of a Kurile Islands earthquake on 01.08.94 ( $D = 76.2^\circ$  to HAM,  $mb = 5.0$ ); lower two traces: PKP-wave group from an earthquake in the Tonga Islands on 30.07.94 ( $D = 150.1^\circ$  to HAM,  $mb = 5.0$ ) (from Bormann et al., 1997).

#### 7.2.4.5 Causes of spectral noise reduction at RUE and BSEG and conclusions

Bormann et al. (1997) estimated quantitatively the reduction of noise amplitudes when traveling from a medium with a low acoustic impedance to a medium with higher acoustic impedance through a sharp impedance discontinuity. Taking into account the best available values for P- and S-wave velocities as well as the densities of the various rock and sedimentary formations in the area of BSEG and RUE, it was estimated that a noise power reduction of about 18.5 dB for BSEG and of 15.6 dB for RUE would be due to the lateral impedance contrast of the anomalous geological bodies at these two sites with respect to the surrounding unconsolidated Quaternary sediments. This would explain about half of the noise power reduction observed at BSEG with respect to HAM (some 30 to 40 dB between 1 and 15 Hz). The remaining reduction of about 10 to 20 dB at BSEG can be accounted for by the distance of BSEG ( $\approx 40\text{ km}$ ) from the seismically noisy city of Hamburg.

For the noise power reduction observed at RUE with respect to BRNL (about 15 to 25 dB in the same frequency range), about 15 dB can be explained by the impedance contrast of the Rüdersdorf anticline. The change in distance to Berlin is less effective (RUE is about 20 km from the city center) because of the noise generated at a busy highway near RUE and ongoing production activity in the quarry.

Below 0.5 Hz, the effect of noise reduction due to these anomalous geological bodies is negligible because their near-surface diameter is then of the order of or smaller than the wavelength of the long-period noise. Large halokinetic, diapir or anticline structures do exist in many other parts of the world with dominating young soft sediment cover (e.g., around the Caspian Sea; west of the Zagros Mountains in Iran; in the USA). A systematic search and use of such structures (or of other anomalous local hardrock outcrops) as sites for permanent seismic recordings is recommended as a way to achieve significant short-period noise reduction. Otherwise, one has either to settle for rather bad noise conditions for near-surface installations or go for expensive borehole installations (see 7.4.5).

### **7.3 Data transmission by radio-link and RF survey**

(A. Trnkoczy, 2002; for an update on modern communication systems used in seismology see the new IS 8.2 of 2013)

#### **7.3.1 Introduction**

Radio links are often used for data transmission in a seismic network. Radio links offer seismic data transmission in real time, are continuous, independent, often robust to damaging earthquakes, and usually involve a reasonable cost.

However, experience shows that the most frequent technical problems with radio frequency (RF) telemetry networks originate in the RF links themselves. This is often the result of a non-optimally designed RF system. Many seismic networks in the world experience unreliable and noisy data transmission. There are even reports of some complete failures. This Chapter gives some general advice on how to design a seismic telemetry system, covering VHF (usually 160 - 200 MHz for seismology) and UHF (usually around 450 MHz for seismology) frequency band FM modulated links, as well as spread spectrum (SS; around 900 MHz or 2.4 GHz) RF data transmission and satellite . The need for a professional RF survey will be explained.

The UHF and VHF frequency bands are still the most frequently used. Spread spectrum and satellite links are becoming more popular in seismology.

#### **7.3.2 Types of RF data transmission used in seismology**

Most of today's RF telemetry seismic networks use the VHF or UHF frequency band. Both bands can be used for frequency modulated (FM) analog signal transmission or digital data transmission with a variety of modulation schemes. Both usually use standard 3.5 kHz bandwidth "voice" channels. It is much easier to obtain permission for these than for special channels with a higher bandwidth. Direct connection distances of up to 150 km (100 miles) are possible with less than one Watt RF power transmitters, if topography permits.

Unfortunately, the VHF band is almost completely occupied in most countries. It is therefore very difficult or even impossible to get permission to use this band. The band is also more susceptible to interference from other RF users and therefore is rarely used for new seismic networks.

Until very recently, the UHF band has been the most popular. But it is now becoming difficult to obtain permission for new frequencies within this band in many countries.

Spread spectrum RF telemetry is a new and increasingly popular alternative in seismology. These links operate at frequencies around 900 MHz or 2,4 GHz. Spread spectrum RF links do not use a single carrier frequency but instead use the entire frequency band dedicated for such links. Many users use the same frequency band so the corresponding transmitter and receiver must identify each other to discriminate from other users using special codes.

The practical advantages of spread spectrum links are that often no permission is needed for their operation and that they are very robust against RF interference (the technology was first developed for defense purposes for just this reason). There are limitations, however, because the maximum RF power of transmitters is defined by national regulations, varies greatly, and dictates the maximum practical connection distance between a transmitter and a receiver. This may impose severe limitations on the wider use of spread spectrum links for seismology. In Western European countries where the limit is 100 mW, connections are only possible up to 20 to 30 km. Direct connection distances around 100 km can be achieved using stronger transmitters (up to 4W) only in the countries that allow them.

Satellite links are becoming more popular in seismometry and undoubtedly represent the future for seismic data transmission. Costs are still a hindrance to the widespread implementation of this technology but these will surely come down.

Most of the commercially available satellite links are of the high throughput type. Usually they are purchased as 110 kHz bands in the GHz frequency range (e.g., Ku-band: 11 to 14 GHz). Frequently, the smallest available bandwidth (and consequently the baud rate) is much higher than usually required for a seismic station or even for a small seismic network. This makes satellite links relatively expensive for small networks. Prices for one 110 kHz band are currently around several hundred dollars per month (1998).

If the size of the network and the total bandwidth required is equal to or slightly smaller than any multiple of the available bandwidth increments, the cost of satellite data transmission may be more acceptable. This is easier to achieve in large national or regional seismic networks. The number of seismic data channels that can be transmitted in a 110 kHz frequency band depends on several parameters: the sampling rate; the number of bits per data sample (dynamic range); whether single direction (simplex) or bi-directional (duplex) links are used; the overhead bits required for error detection, forward error correction (FEC), and link management.

One of the important issues which varies from country to country relates to the central satellite recording site (the hub). In some countries, where the communication market is open, a seismic network owner may have its own 'private' hub directly at the central recording site. The cost of equipment for such a local hub varies from \$80,000 to about \$200,000 (in the year 2001). In countries with a more restricted communications market only a shared hub owned by a communications company may be available. In this case, not only is the cost of satellite communications higher but there will be additional costs for the communication links from the shared hub to the seismological central recording site. These usually use leased lines and the costs can be significant, particularly if the distance involved is large. Cost analysis of different satellite systems is complex and the prices vary significantly from country to country. A very careful cost analysis is recommended before making any final decision about satellite links.

A practical problem with satellite links is the relatively high power consumption of the equipment installed at a seismic station. In most cases, we must consider at least 50W power consumption for the data transmission equipment at each site. This significantly exceeds the power consumption of RF equipment traditionally used in seismology, including spread spectrum transmitters. It creates the need for large arrays of solar panels at stations without mains power and for bigger back-up batteries for a given station autonomy.

Nonetheless, the costs of satellite communications are constantly decreasing thanks to increasing liberalization in the communications market which will encourage the use of satellite links. No other communication system has the potential of satellite links for high reliability at the most remote and distant seismic stations.

### **7.3.3 The need for a professional radio frequency (RF) survey**

The design of VHF, UHF or spread spectrum RF telemetry links in a seismic network is a specialized professional technical matter. Practice shows that guesswork and an approach based on "common sense" usually lead to problems or even complete failure of a project. The following misunderstandings and oversimplifications are commonly encountered:

- the amount of data that must be transmitted in seismology is often underestimated. Seismology requires a much larger data flow (baud rate) than most other geophysical disciplines, for example several orders of magnitude more than meteorology;
- the required reliability for successful data transmission in seismology is also frequently underestimated. Missing data due to interruptions on the links, excessive noise, spikes, and data errors are particularly destructive for networks operating in triggered mode and/or having any kind of automatic processing. With old paper seismograms and analog technology, spikes, glitches, interruptions and other 'imperfections' are relatively easily "filtered out" by the seismologist's pattern recognition ability during the analysis. However, the same errors, if too frequent, can make the results of an automatic computer triggering and/or analysis totally unacceptable;
- a false comparison with voice RF channels is made frequently. People try to verify a seismological RF link between two points using walkie-talkies. If they can communicate, they expect that transmission of seismic data will also be successful. Note that voice channels allow a much lower signal-to-noise ratio while still being fully functional because human speech is highly redundant. Also, the RF equipment parameters in walkie-talkies and in seismic telemetry are very different, making such "testing" of RF links meaningless.
- another wide-spread belief is that the "line of sight" between transmitter and receiver is a sufficient guarantee for a reliable RF link. This may or may not be true. It is only certain for very short links up to about 5 km length with absolutely no obstructions between the transmitter and the receiver (such links may occur in some small local seismic networks). Fading, i.e., the variation of the intensity or phase of an RF signal due to changes in the characteristics of the RF signal propagation path with time, becomes a major consideration on longer links. The real issues in link reliability

calculations are the equipment's gains and losses, RF signal attenuation based on Fresnel ellipsoid obstruction, and the required fading margin. The resultant reliability of the link can then be expressed as a time availability (or probability of failure or time unavailability) as a percentage of time in the worst month of the year (or per year). During 'time unavailability', the signal-to-noise ratio at the output of the receiver is lower than required, or the bit error rate (BER) of digital data transmission link is higher than required. Many parameters are involved in the RF path analysis including transmitter power, frequency of operation, the various losses and gains from the transmitter outward through the medium, receiver antenna system to the input of the far end receiver and its characteristics. In link attenuation calculation, the curvature of the Earth, the regional gradient of air refractivity, the type of the link regarding topography, potential-wave diffraction and/or reflections, time dispersions of the RF carrier with digital links, processing gain and background noise level with spread spectrum links, etc. all play an important role.

We strongly recommend having a professional RF survey during the seismic network planning procedure. IS 7.1 provides the information on what preparation is needed if an RF survey is purchased as a service along with the seismic equipment.

### **7.3.4 Benefits of a professional RF survey**

The benefits of a professional RF survey are:

- it ensures that the links will actually provide the desired reliability, which has to be decided beforehand. During the RF survey, the design parameters of the links in a network are varied until the probability of an outage in the worst month of a year drops below the desired value. This may require additional investment in equipment, but it will prevent unreliable operation or may save some money by loosening the requirements where appropriate;
- it guarantees the minimum number of RF repeaters in a network. This results in a direct benefit to the user in having less equipment, fewer spare parts, and in cheaper and easier maintenance. There will also be lower instrumental noise in the recorded signals for FM analog networks and a better BER performance for digital networks. Note that in most designs for analog FM telemetry, every additional repeater degrades data quality to some extent and always decreases the network reliability;
- It will determine the minimum number of licensed frequencies required in a network without sacrificing data transmission reliability. Note that the required number of different carrier frequencies in VHF and UHF telemetry can be significantly smaller than the total number of the links in the network. This prevents unnecessary pollution of RF space in the country. Use of fewer frequencies also benefits the user since they are easier to obtain and fewer different RF spare parts are required;
- the robustness of the entire seismic network to lightning threat is significantly increased by a proper RF layout, for example, one should always avoid repeaters which relay data from many seismic stations because any technical failure of the repeater will result in severe data loss;
- reduced power consumption can be achieved by calculating the minimum sufficient RF output of the transmitters. This results in less pollution of RF space in the country. The user also benefits from lower power consumption at remote stations;



- minimizing the heights of antenna masts and the minimum gains of the antennae has potential for cost saving.

### 7.3.5 Radio-frequency (RF) survey procedure

An RF survey usually considers the RF equipment to be used, a topographical profile from each transmitter site (remote seismic station) to each receiver site (central recording site or repeater), local RF path conditions, and the desired reliability of the link. It is based on decades of experience of transmission statistics from all over the world and computer modeling using specialized software. Field RF measurements are rarely performed because they are expensive and time-consuming and they are often less reliable than calculations. RF transmission conditions vary with time (diurnal, seasonal, weather dependent), vary unpredictably and within climatic zones. Theoretical calculations include the full statistics of these variations whereas practical one-time measurements suffer from unpredictable variations in fading. However, even if no measurements are planned, a communications expert still has to visit all potential seismic sites during the site selection procedures to assess local topography and to check for the existence of potential RF obstacles which may not be evident from topographic maps.

If the RF link calculation based on a given set of input parameters does not give the desired reliability, some of the input parameters must be changed. We can change topographical profile by either repositioning stations or by introducing a new RF repeater. We can change the antenna type and/or increase their gain. We can increase antenna mast height or increase transmitter output power (seldom effective) or we can change the RF equipment completely (significantly more powerful transmitters and/or more sensitive receivers).

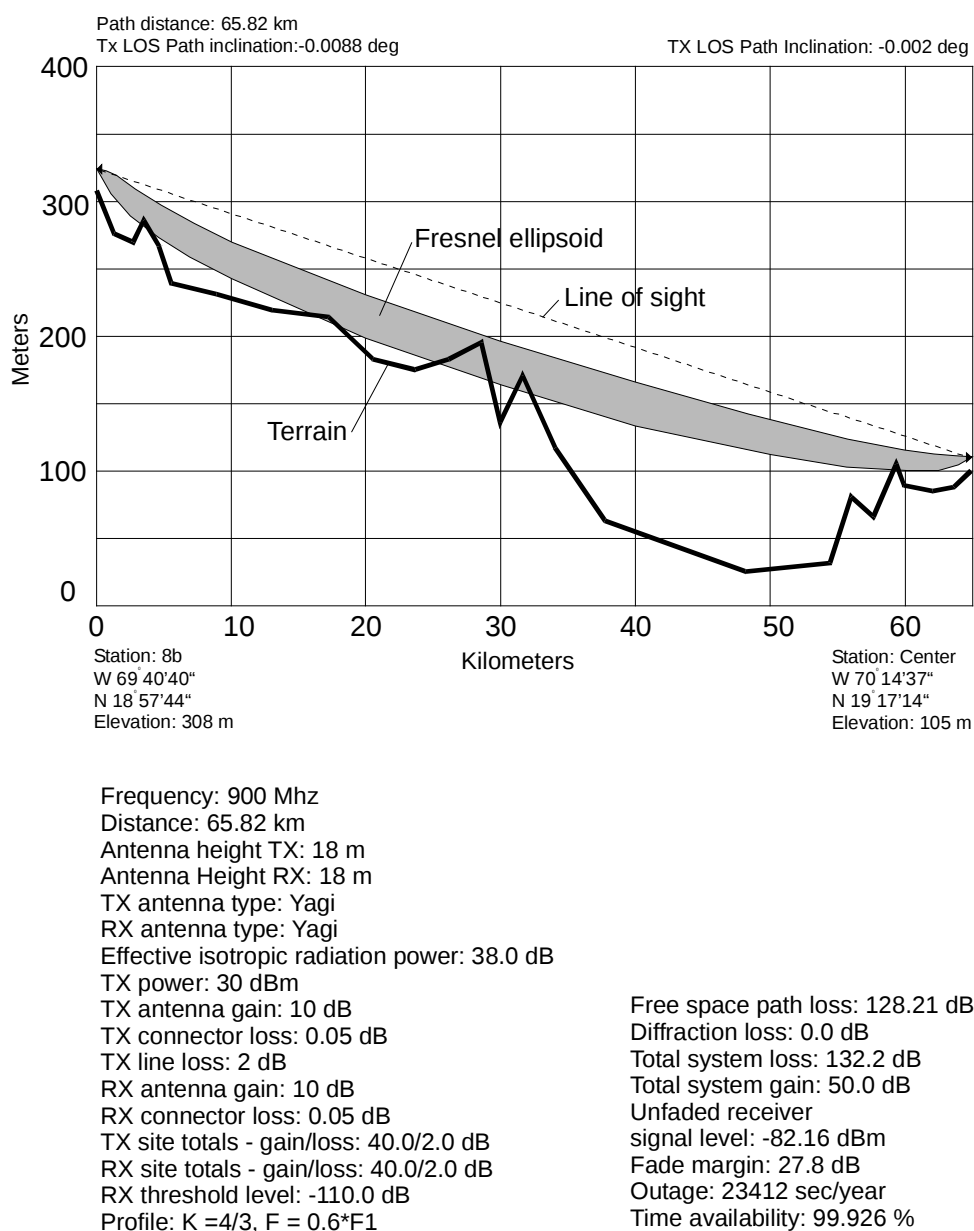
Topographic profiles are usually taken from 1:50.000 scale topographical maps. In most cases, many more profiles than stations available in the network are taken and links calculated before we determine the final RF layout of a network. A great deal of this work can be done before fieldwork starts, but profiling is always needed during the fieldwork.

The result of an RF link calculation is shown in Fig. 7.38 with input parameters on the left and output parameters on the right. The figure intentionally shows an example where there is a "direct line of sight", but the profile doesn't guarantee acceptable link operation. Note the curved path of the first Fresnel ellipsoid where the RF energy actually travels from the transmitter to the receiver. This curvature is mostly due to the regional gradient of air refractivity. In the example, this ellipsoid hits the mountain ridge and causes a significant loss of energy or possibly link failure.

For analog VHF or UHF telemetry it is usual to regard a time availability of about 99.95% (equivalent to about 15 minutes of outage of each link per month) in the worst month as being marginally acceptable and 99.99% as good. If we use an RF repeater between the seismic station and central recording site, we have to increase the required reliability of individual sections to give the required reliability for the entire link.

In digital data transmission, the bit error rate (BER) is used as a measure of data link reliability. BER strongly depends not only on physical reliability of the RF link but also on error detection and error correction methods used in the RF equipment (modems). For example, one-directional (simplex) links are generally far less reliable than bi-directional

(duplex) links, even if the RF links themselves are of the same quality in terms of RF signal to noise. This is because duplex links allow repeated transmission of corrupted data blocks until they are received without error whereas simplex links result in corrupted data, unless forward error correction (FEC) methods are used. Due to the complexity of the problem, a precise targeting of desired BER is usually beyond the scope of seismic network projects.



**Fig. 7.38** Result of an RF link calculation with input parameters on the left and output parameters on the right.

Something similar is the case for spread spectrum links where another factor complicates the situation. Spread spectrum receivers incorporate so-called "processing gain". These receivers are capable of resolving very weak RF signals, which may even be a few dB below the RF

noise at the receiver site. However, the problem is that the amplitude of the RF noise at the receiver site is generally unknown. Note that every new spread spectrum transmitter increases the background noise in the band of operation of the spread spectrum system and since this band is open to the public, it is difficult to predict its actual noise. Consequently we will not know exactly the sensitivity of a receiver, resulting in a less reliable estimate of the link availability.

Specialized spread spectrum measuring equipment is extremely expensive. The algorithms which are used to resolve the sub-noise level RF signals in the receivers also present a problem. They are mostly proprietary and therefore not generally accessible. Both facts significantly reduce the practicality of measurements of the reliability of spread spectrum links for seismological purposes.

Fortunately, some spread spectrum equipment manufacturers provide special software which allows easy but approximate link reliability measurements for the transmitters and receivers to be used in the seismic system. Taking into account a safety margin due to temporal variation of RF transmission conditions, one can successfully use these measurements for an approximate estimate of link quality. However, it is difficult to relate these proprietary 'reliability scales' to standard parameters like probability of link outage or BER. Nevertheless, classical RF signal attenuation calculations still give valuable information about RF energy propagation over a given topographic profile. These results, combined with measurements using manufacturer's proprietary 'reliability scales' and practical experience, suffice in almost all seismometric projects.

The cost of a professional RF survey is generally around a few percent of the total investment in a new seismological network. Practice shows that its benefits are well worth the investment. An RF survey is a major step toward the reliable operation of any future telemetry seismic network.

### **7.3.6 The problem of radio-frequency interference**

While spread spectrum links are fairly robust, radio-frequency interference between a VHF or UHF seismological system and other RF users is quite a common and difficult problem in many developing countries. In some countries, the lack of discipline in RF space causes unforeseen interference. In others, insufficient maintenance of high-power communication equipment results in strong radiation from the side-lobes of powerful transmitters that may also interfere with seismological links. Army facilities, particularly if they operate outside civil law, especially some types of radars, frequently interfere with seismological links. The risk of interference is very high if seismic stations are installed at sites which are also used for other high power RF communication equipment (see IS 7.2). Extensive use of walkie-talkies can also cause problems.

In some developing countries, the use of RF spectrum analyzers, which can frequently reveal the origin of interfering signals, is prohibited for security reasons, particularly for foreigners. In any case, interfering RF sources may appear only very intermittently and so are difficult to detect.

Note that RF interference problems due to indiscipline in RF space are generally beyond the control of a seismic equipment manufacturer and/or foreign RF survey provider. They can

only be solved, or at least mitigated, by involving local RF communication experts during the very early phases of network planning. These people are familiar with the real RF conditions in the country and can provide better advice than any foreign expert. If a new seismic network experiences interference problems, only very tedious and time consuming trial-and-error procedures (swapping frequencies of the links or even VHF/UHF bands, changing antenna orientation and polarization, or even re-positioning of stations or repeaters) may help. However, the results are unpredictable. One should also be aware that the allocation of frequencies may change in future and disturbances remedied today may reoccur later.

## **7.4 Seismic station site preparation, instrument installation and shielding**

### **7.4.1 Introduction and general requirements (A. Trnkoczy; version 2002)**

When installing a seismometer inside a building, vault, or cave, the first task is to mark the orientation of the sensor on the floor. This is best done with a geodetic gyroscope although a magnetic compass will often suffice. The magnetic declination must be taken into account. A compass may be deflected, showing a false reading, when inside a building so the direction should be taken outside and transferred to the site of installation. A laser pointer may be useful for this purpose. When the magnetic declination is unknown or unpredictable (such as in high latitudes or volcanic areas), the orientation can be determined with a sun compass. Special requirements and tools for sensor orientation in boreholes are dealt with in 7.4.6.2.

To isolate the seismometer from stray electric currents, small glass or perspex plates should be cemented to the ground under its feet. The seismometer can then be installed and tested. Broadband seismometers should be wrapped with a thick layer of thermally insulating material. The exact type of material does not seem to matter; alternate layers of fibrous material and heat-reflecting blankets are probably the most effective. The edges of the blankets should be taped to the floor around the seismometer. Further information on suitable and proven thermal insulation for broadband seismometers, including illustrations, can be found in 7.4.2.1, 7.4.4.2 and 5.5.3. One has to be aware that electronic seismometers generate heat and so may induce convection in any open space inside the insulation. It is therefore important that the insulation fits the seismometer tightly.

For the permanent installation of broadband seismometers under unfavorable environmental conditions, they should be enclosed in a hermetic container. A problem with such containers (as with all seismometer housings) is that they cause tilt noise when they are deformed by barometric pressure. Essentially three precautions are possible: either the base-plate is carefully cemented to the floor, or it is made so massive that its deformation is negligible, or a "warp-free" design is used, as described by Holcomb and Hutt (1992) for the STS1 seismometer (see DS 5.1).

To prevent or reduce corrosion in humid climates, desiccant (silica gel) should be placed inside the container, including inside the vacuum bell, of an STS1 seismometer. Broadband seismometers may also require some magnetic shielding (see 5.5.4).

Civil engineering work at remote seismic stations should ensure that modern seismic instruments can be used to their fullest potential by sheltering them in an optimal working

environment. Today's high dynamic range, high linearity seismic equipment is of such quality and sensitivity that seismic noise conditions at the site and the environment of the sensors have become much more important than in the past. Apart from site selection itself, the design of seismic shelters is the determining factor in the quality of seismic data acquisition.

Seismic vaults are currently the most common for new seismic stations (see 4.2). They are the least expensive but suffer more from seismic noise because of their near-surface installation. Alternatives include seismic installations in abandoned mines, in specially constructed tunnels (see 7.4.3) and in boreholes (see 7.4.5 and 7.4.6). These have the advantage of high temperature stability and significantly reduced surface and tilt noise because of the significant overburden. The low tilt noise is of particular importance for long-period and broadband seismometers because of their high tilt and temperature sensitivity (see 7.4.4, 5.3.3 and 5.3.5). A variety of factors must be considered before the optimal technical and financial solution for a seismic installation is found. These include the type of monitoring or research to be carried out, the kind of equipment to be installed, existing geological and climatic conditions, already existing potentially suitable structures and sites, available construction materials or alternative technical solutions, accessibility of and available infrastructure/power supply at the station.

Various solutions can be employed with equal success. Much depends on potential future upgrades of the instrumentation and site, what working conditions are desired for maintenance and service personnel, and, of course, on the funds available. Because of these diverse considerations, no firm design and civil engineering drawings are provided in this document. Instead, the general requirements that must be satisfied are described in detail so that, e.g., in the case of seismic vaults, any qualified civil engineer can design the shelter for optimal performance, taking into consideration local conditions in a given country and at a specific site.

## **7.4.2 Vault-type seismic stations (A. Trnkoczy; version 2002)**

This section describes the general conditions to be considered when constructing seismic vaults. A vault for seismic data acquisition and transmission equipment should satisfy the following general requirements:

- provide adequate environmental conditions for the equipment;
- ensure the proper mechanical contact of seismic sensors with bedrock;
- prevent seismic interaction between the seismic shelter and the surrounding ground;
- mitigate seismic noise generated by wind, people, animals, and by potential noise sources within the vault;
- ensure a suitable electric ground for sensitive electronic equipment;
- provide sufficient space for easy access and maintenance of the instruments.

These requirements will be discussed in detail below. A design example of a seismic vault for a three-component short-period (SP) station together with its upgrade for broadband (BB) and potentially very broadband (VBB) seismic sensors will be given, complemented by some technical hints at the end of this section. Other examples of vault-type seismic shelters are given in 7.4.4.3. Detailed installation guidelines for BB and VBB stations can be found in Uhrhammer et al. (1998). Alternative vault designs of typical 'classic' seismometer vaults are given in Figures 4.5b-e of the old MSOP (Willmore, 1979; also accessible on the IASPEI website <http://www.iaspei.org/projects/NMSOP.html> via the link Manual of Seismological

Observatory Practice (1979 Edition)). Some recent complementary comments have been added by the Editor in the inserted box below.

### **Complementary comments on seismometer installations in vaults and on piers (2011)**

Several NMSOP-1 users asked to specify somewhat more in this section how to “decouple” best the place or pier on which a seismometer is installed in a vault or in a building from ambient and human disturbances. The editor asked two experts. Their answers see below:

#### **Amadej Trnkoczy on 15.06.2011**

Regarding the vault design my opinion has always been the same: dig the vault until you reach a good bedrock, then put as little as possible fine concrete on it to make a flat surface big enough to put a seismometer or seismometers on it. Build around the vault as small and as light as possible 'housing', the best is none. Just a firm (strong wind!), water tight cover over the vault and that is it. In this case there is no structure/soil or structure/seismic pier interaction possible since there is practically no 'structure'. The problem is solved.

Generally modern seismic instruments are so sensitive that there should be no housing for people to live or work close to them. To my opinion there is no way to decouple a seismic pier from human activity so close to it. Also such housings may cause problem during windy periods.

In old times, when sensors were much less sensitive, this was allegedly possible using several different methods (like "mechanical dampers", or separate foundation for the housing and for the pier, etc). I would not recommend such approaches for regular seismic stations of a seismic network today.

#### **Thomas Forbriger on 27.06.2011**

“Decoupling” is probably meant to avoid noise on the pier caused by forces exerted on the external surface of the vault. Most effective decoupling should always be achieved by underground installations in solid rock and good coupling of the pier to bedrock. Better decoupling as provided by overburden competent rock probably cannot be achieved by any other measures.

Explicit decoupling from the housing in above-surface installations may be an issue in sedimentary settings (loose soil). In such cases you should avoid mechanical coupling of the pier to the walls of the above-surface building. There should always be a gap in between with the pier being only coupled to the ground (if possible on bedrock). This will however not avoid noise coupled to the ground surface outside the building which can be significant for long-period observations in above-surface installations.

Better focus on good coupling of the pier to bedrock.

### **7.4.2.1 Controlling environmental conditions**

Adequate shelter for seismic equipment should:

- prevent large temperature fluctuations in the equipment due to day/night temperature differences or because of weather changes;
- prevent large temperature fluctuations in the construction elements of the vault, resulting in seismometer tilt;
- ensure adequate lightning protection;
- mitigate electromagnetic interference (EMI);
- prevent water, dust and dirt from entering the shelter;
- prevent small animals from entering the shelter.

At very low seismic frequencies and in VBB seismometers, air pressure changes also influence seismometer output. Special installation measures and processing methods can be used to minimize the effect of air pressure. However this issue will not be treated here. For more information see Beauduin et al. (1996).

### ***Mitigating temperature changes***

In general, seismic equipment can operate in quite a broad temperature range. Most of the equipment on the market today is specified to function properly between  $-20$  and  $+50$  degrees Celcius. However, this is the operating temperature range – that is, guaranteeing only that the equipment functions at a given constant temperature within these limits.

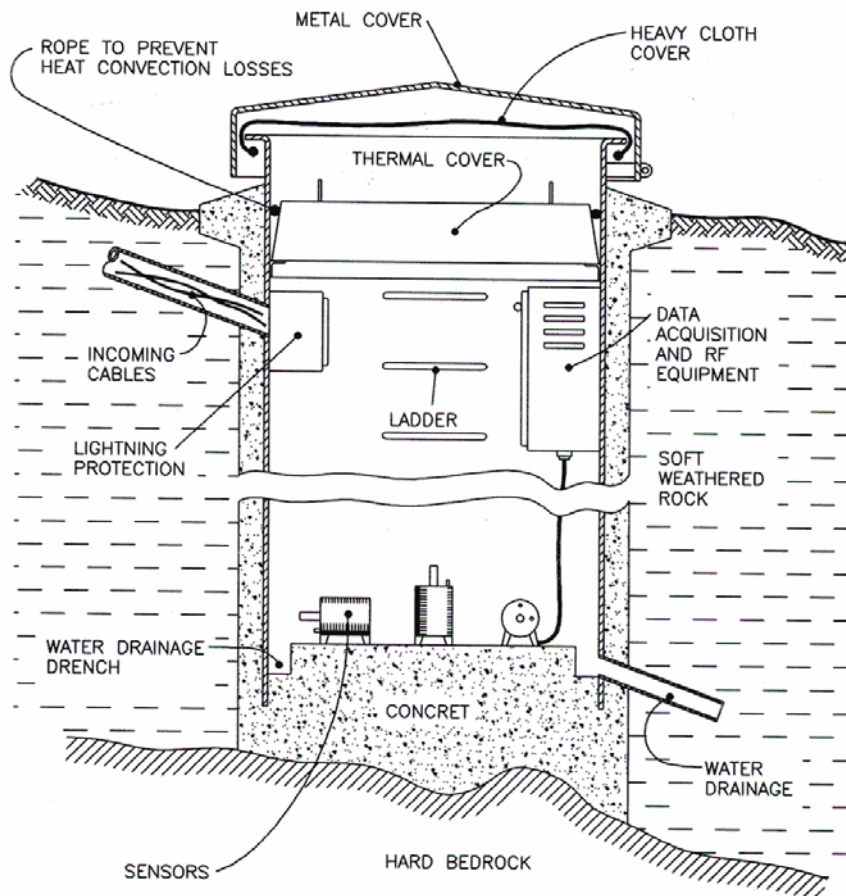
Temperature changes with time, particularly diurnal changes, are far more important than the high or low average temperature itself. Many broadband seismometers require mass centering if the temperature "slips" more than a few degrees Celcius, although their operating range is much wider. Even small temperature changes can cause problems with mechanical and electronic drifts which may seriously deteriorate the quality of seismic data at very low frequencies. Unfortunately, the practical sensitivity of the equipment to temperature gradients is rarely provided by manufacturers. Very broadband (VBB) seismometers require extremely stable temperature conditions which are sometimes very difficult or impossible to assure in a vault-type shelter. VBB sensors usually require special installations (see Uhrhammer et al., 1998). Short-period (SP) seismometers, particularly passive ones, and accelerometers are much less sensitive to temperature changes.

In general, thermal drifts should be kept acceptably small by thermal insulation of the vault. However, the requirements differ significantly. Maximum  $\pm 5$  deg C short-term temperature changes can be considered a target for passive SP seismometers and force-feedback active accelerometers. To fully exploit the low-frequency characteristics of a typical 30-sec period BB seismometer, the temperature must be kept constant within less than one degree C. To fully exploit a several-hundred-seconds period VBB sensors only a few tens of millidegrees C per month are recommended.

Data loggers and digitizers can tolerate less stable temperatures, i.e., on average, the temperature change would be ten times greater than on a BB seismometer for the same change in output voltage. The best digitisers, for example, change their output voltage less than  $\pm 3$  counts in room temperature conditions. If daily temperature changes are less than 1 deg C, their output voltage changes less than  $\pm 1$  count (Quanterra, 1994).

Some elements such as some computer disk drives, diskette drives, and certain time-keeping equipment, may require narrower operating temperature tolerances. The most effective way to assure stable temperature conditions is an underground vault that is well insulated (see Fig. 7.39). Underground installations are also the best for a number of other reasons.

Thermal insulation of active seismic sensors is done in two places. First, the interior of the vault is insulated from external temperatures, and second, the sensors themselves are insulated from residual temperature changes in the vault. In the most critical installations, the seismic pier itself is insulated along with the sensors.



**Fig. 7.39** Example of a vault for a short-period three-component seismic station made of a large-diameter metal pipe with thin concrete walls.

Underground vaults are usually insulated with a tight thermal cover made of styrofoam, foam rubber, polyisocyanuratic foam, or other similar, non-hygroscopic insulation material (Fig. 7.39, Figs. 7.41 and 7.42). Such materials are usually used in civil engineering for the thermal insulation of buildings. They come in various thicknesses, often with aluminum foil on one or both sides. This aluminum layer prevents heat exchange by blocking heat transfer through radiation. Thinner sheets can be glued together to make thicker ones. Casein-based glues are appropriate for styrofoam and expanding polyurethane resin is used to glue polyisocyanuratic foam sheets.



In continental climates, a 20 cm (8") layer is considered adequate but in extreme desert climates, up to 30 cm (12") of styrofoam is recommended. In equatorial climates a 10 cm (4") layer is considered sufficient.

There are two thermal cover design issues that are particularly important. Special care must be taken to assure a tight contact between the vault's walls and the thermal cover. If it is not tight, heat transfer due to convection through the gaps can easily be larger than the heat transfer through the thermal cover by conduction. This can undo the insulating effects of the cover. One way to achieve a tight thermal cover is shown in Fig. 7.43. A "rope" is tightly pressed into the gaps between the vault's walls and the thermal cover as well into the wedge-like gap between the cover halves seen in Fig. 7.41. This "rope" can be made of insulating fibers and is usually used for industrial hot water pipe insulation. It is available in different sizes and is inexpensive.

The cover should be placed at or below the depth at which the ground heats up during the day – not on the top of the vault. In desert areas, surface ground temperatures can exceed 80 deg C. At 30 cm (12") depth, temperatures of 50 deg C are not unusual. In such conditions, the thermal cover must be placed 40 - 50 cm (16" - 20") below ground level. A thermal cover of any thickness at the top of the vault, particularly if the vault's rim stands significantly above the surface, has almost no effect.



**Fig. 7.40** Interior of a seismic vault made of welded metal sheets. The vault is big enough to accept weak- and strong-motion instrumentation together with data acquisition and transmission equipment.

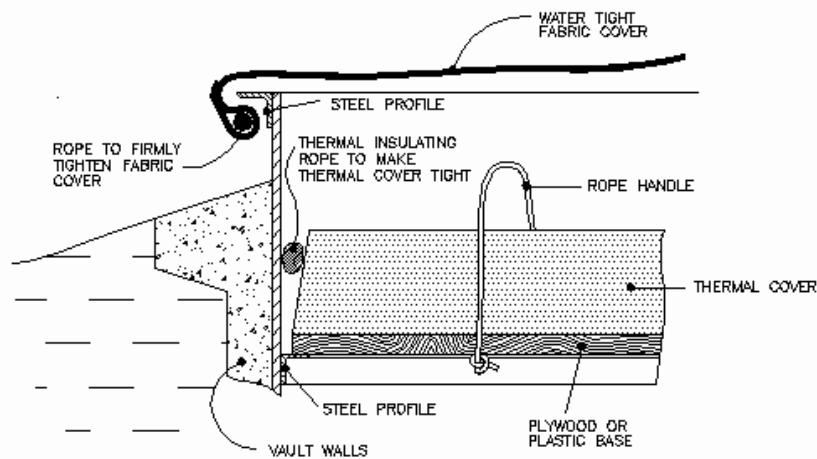


**Fig. 7.41** Thermal cover of a seismic vault in two pieces made of thick styrofoam. The gaps between the cover and the vault walls and between both pieces must be tightly sealed.

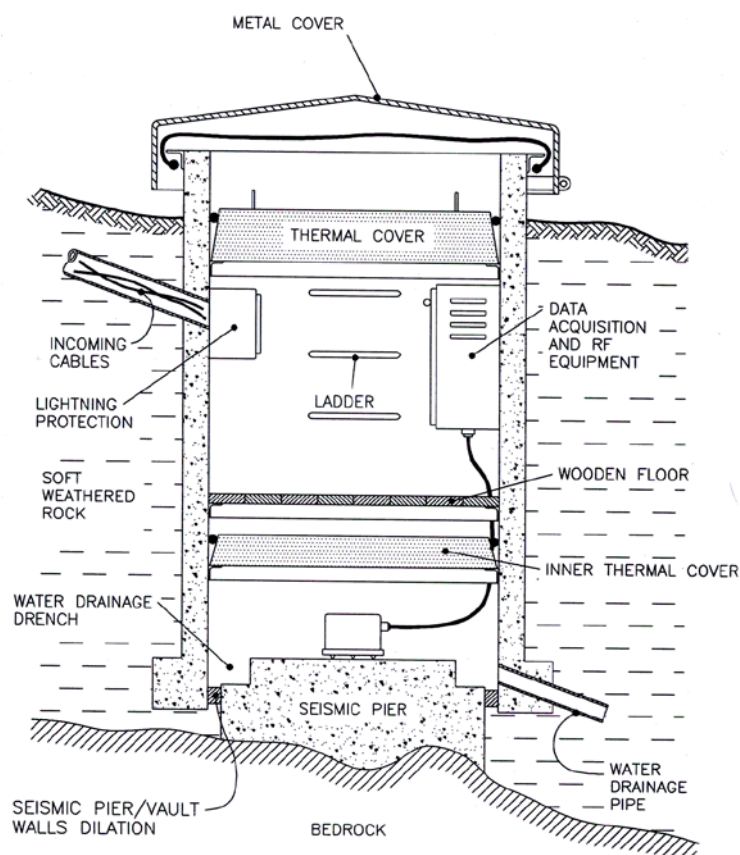


**Fig. 7.42** Installing thermal cover in a seismic vault. In climates with large diurnal temperature changes the cover should be positioned lower in the vault where external ground temperature does not change significantly.

If vaults are used for BB or even VBB stations (see Wielandt, 1990), it is advisable to make a second inner thermal cover just above the sensor but below the floor where all other equipment is installed (see Fig. 7.44). Since most maintenance work relates to batteries, data recording, and data transmitting equipment, the thermal- and mechanical-sensitive BB/VBB sensors are not disturbed at all during service visits.

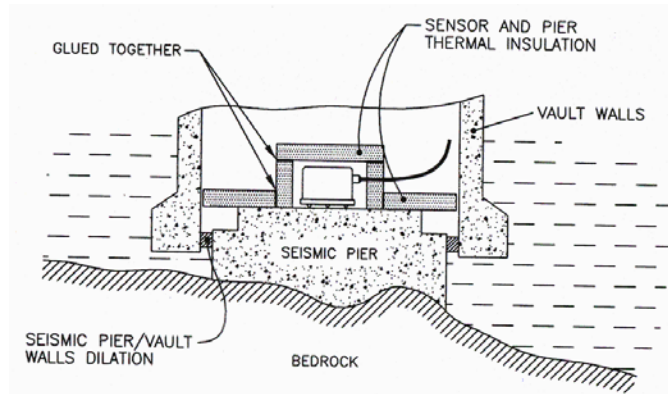


**Fig. 7.43** Detail of making a thermal cover effective by filling up the gaps between the cover and vault walls with insulation material and making the vault tight against dust, dirt, and rain during windy periods with a fabric cover.



**Fig. 7.44** Example of a BB or VBB seismic vault with a separate compartment for sensors and double thermal cover. Usually, the sensor itself is additionally isolated (see. Fig. 7.50). A thermal isolation box is usually put around the sensors to additionally insulate them.

Thermal insulation of the seismic pier itself, together with the seismometer, is the best method of insulation (Fig. 7.45). This method keeps the heat transfer between seismometer and vault interior as low as possible, while at the same time assuring good thermal contact with the thermally very-stable ground. Thus, the thermal inertia of the system is very large, limiting the rate of temperature changes to a minimum.



**Fig. 7.45** Thermal isolation of a VBB sensor and surrounding seismic pier and mechanical separation of the pier from the vault walls for the most demanding applications.

### ***Thermal tilt mitigation***

Special measures are required to prevent thermal deformation and tilt of the seismic pier in a vault to allow the study of extremely low frequency signals with VBB seismometers. Modern VBB sensors, the horizontal components in particular, can detect tilts of a few nanoradians. A human hair placed under the corner of a level football field or an air pressure difference of only 0.1 mbar over a distance of several km would cause such a tilt. According to Wielandt (see section 5.3.3) a tilt of about  $10^{-9}$  rad would result in a noise ground acceleration amplitude of  $10^{-9}$  g in the horizontal components but only of  $10^{-11}$  g in the vertical one.

Homogeneity of the seismic pier and surrounding soil, as well as civil engineering details of vault design are very important. Uhrhammer et al. (1998) recommend the physical separation of the seismometer pier and the vault walls (see Figs. 7.44 and 7.45). This separation assures that minute changes in the dimensions of the vault walls due to temperature change do not tilt the seismic pier. However, since such seismic vaults are not constructed "in one piece," one has to be particularly careful that the contact between the pier and vault walls is still watertight.

The seismic pier should be made of homogenous material and neither it nor the walls of the vault should use any steel reinforcement. Steel and concrete have different temperature expansion coefficients which cause stress and unwanted minute deformation of the structure of the vault if the temperature changes. Steel is unnecessary anyhow because structural strength is practically never an issue except in the very deepest of vaults. Sand aggregates used for concrete should be homogenous, fine-grain, and of uniform size rather than of varying size as in the usual concrete mixture. Uhrhammer et al. (1998) recommend sieved sand with 50% Portland cement. After the pier is poured, the concrete must be vibrated to remove any trapped air.

### ***Lightning protection***

Lightning causes most of the damage to seismic equipment around the world and lightning protection is probably the most important factor in preventing station failure. We know of several seismic networks that lost half or more of their equipment less than two years after installation because of inadequate lightning protection. Of course a direct hit by lightning will cause equipment damage despite the best protection. Fortunately, this rarely happens. Most lightning-related damage is caused by induction surges in cables, even when the source is some distance from the station.

Climatic and topographic conditions at a site vary greatly and determine the degree to which one should protect the system from lightning. Tropical countries and stations on top of mountains are the most vulnerable and therefore require the most lightning protection measures.

Lightning protection includes the following measures:

- proper cabling that minimizes voltage induction during lightning;
- proper use of special electronic devices to protect all cables entering the seismic vault from voltage surges;
- a good grounding system since no practical lightning protection measure works without grounding;
- enclose the equipment in a "Faraday cage" either by making a metal shielded seismic vault or a loose mesh of ground metal strips around the vault. This creates an equipotential electric field around the equipment, thus decreasing voltage drops on equipment and cables during lightning strikes.

If any one of these measures is not undertaken, the others become largely ineffective.

The best lightning protection is a metal seismic vault. The exterior of the vault should not be painted so that good electrical contact can be made with the surrounding soil, thereby lowering impedance. If the main cover or any other part of the vault is metal, it should be connected to the vault's walls using a thick flexible strained wire.

In any event it is necessary to protect all cables entering the seismic vault. Many high quality seismic instruments already have internal lightning protection circuitry, but these measures are sometimes not enough for high lightning threat regions. Lightning protection may include gas-discharge elements, transient voltage suppressors (transorbs), voltage dependant resistors, and similar protection components.

The lightning protection equipment of the cables must be installed at the point where they enter the vault. It must be grounded at the same point with a thick copper wire or strip that is as short as possible. The unprotected length of any cable within the vault must be kept to an absolute minimum.

All cables entering the vault must be protected. Voltage surges usually occur in all cables, and leaving a single long cable unprotected is virtually the same as leaving all the cables unprotected.

All metal equipment boxes should be grounded with a thick copper grounding wire or strip (> 25 mm<sup>2</sup> cross-section) to the same point where the lightning protection equipment of incoming cables is grounded. Use a tree-shaped scheme for grounding wires. All these wires

should be as short as possible and without sharp turns. All the cables in a vault should be kept to a minimum length. No superfluous cables or even coiled lengths of excess cable are acceptable. These are true lightning catchers.

Telephone and power companies usually install lightning protection equipment for their lines. This should be required of them when arranging these services. Manufacturers of seismic equipment can also provide and install such equipment if asked.

Note that there is never a 100% safe lightning protection system. However, for high lightning risk regions and for expensive and delicate seismic equipment, long years of practice show that investing in an effective lightning-protection system pays off in the long run.

### ***Electro-Magnetic Interference protection***

The problem of electro-magnetic interference (EMI) is not normally a very important issue because seismic stations are generally situated in remote rural locations. However, in such regions the main power lines can frequently be of low quality. We recommend using mains power voltage stabilizing equipment in such cases. This equipment usually incorporates EMI filters and voltage surge protection, which further protects seismic equipment from failures and EMI-generated noise. In general, metal seismic vaults protect equipment from EMI very effectively.

Some passive seismometers with moving magnets and separate components generate EMI during mass motion. Since this may influence surrounding sensors, you should not install such seismometers too close together. A minimum distance of 0.5 m (1.5 feet) is recommended. A simple test can assure you that cross talk is insignificant. Disconnect and un-damp one component, move the seismometer mass by shaking it slightly and measure the output of both the other components. There should be no cross-talk.

In addition, seismometers should not be placed too close to the metal walls of a vault. This minimises potential changes in the static magnetic field, which may slightly influence the generator constant of some seismometers.

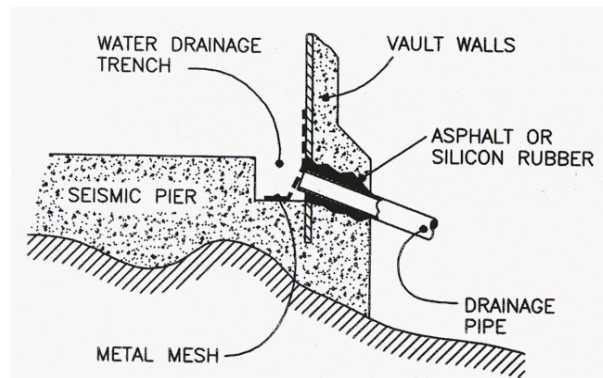
Data recording equipment with mains transformers should not be installed next to, or on the same pier as sensors. The transformer may cause noise in the seismometer signals either through its magnetic field or due to direct mechanical vibrations at 50 or 60 Hz. The same is true for magnetic voltage stabilizers, if used at the site. Place such equipment in a metal housing for additional magnetic shielding and install it on the wall of the vault.

### ***Water protection***

Water entering seismic vaults is probably the second most common cause of station failure. The most effective way to prevent water damage is vault drainage (Fig. 7.46). Use a hard plastic tube of about 3 cm (1") diameter, such as used for water pipelines. The drainage pipe must be continuous and have at least a 3% gradient, particularly in regions where the ground freezes during the winter. If drainage is impossible, as is often the case for deep vaults, water tightness of the vault is of the utmost importance. Note that a high ground water level and porous concrete vault walls more or less guarantee water intrusion.

Water tightness is easy to achieve if the walls of the vault are made of metal welded from plain or corrugated iron sheets or from large-diameter metal tubes, providing the welds are of

good quality. If the vault is made of concrete and has no water drainage, the concrete should be of a very good, uniform quality. Water-resistant chemicals should be added to the mix to help keep it water-tight. The concrete must be vibrated during construction to assure homogeneity of the walls.



**Fig. 7.46** Water drainage pipe and vault trench around the seismic pier.

The bottom of the seismic vault - the seismic pier - is always made of concrete. Once again, use good quality, uniform-aggregate concrete with water-resistant additives. The bottom should have a water drainage ditch (see Figs. 7.39 and 7.46) around the flat central pier on which the sensors are installed. For vaults with external water drainage, the ditch should be at least 5 cm (2") deep and 10 cm (4") wide. For the vaults without drainage, this ditch should be larger (at least 15-cm by 15-cm or 6"x 6") so it can collect more water.

Making the joint between the vault walls and floor requires special care. Use asphalt to seal any cracks by heating the concrete with a hot-air fan and then pouring hot asphalt into them. The cables entering the vault also require special care. They are normally installed in a plastic or metal tube that should fit snugly into the appropriate hole in the vault wall. Use silicon rubber or asphalt to seal any gaps.

In vaults designed for VBB seismometers whose seismic pier is mechanically separated from the walls, water tightness represents a special challenge. Once again use soft asphalt to make the gap between the walls and the pier watertight.

The upper rim of the vault must be at least 30 cm (1 foot) above the ground. At sites where a lot of snow is expected, this should be higher, up to 60 cm (2 feet). Slush is particularly troublesome with regard to keeping vaults watertight. Where possible, the surrounding terrain should descend radially from the top of the vault.

One practical measure is to create a small "overhang" at the top edge of the vault (see Fig. 7.43). This ledge should be about 5 cm (2") out from the vault wall. A thick, watertight fabric cover can be hooked over this metal edging. The cover is pulled tight to the vault by rope and prevents water from entering the vault during windy, rainy periods. It also protects against dust and dirt and provides some additional thermal insulation.

To minimize the danger of equipment flooding, install all equipment, apart from the sensors, on the wall of the vault or on a raised platform.

### ***Protection from small animals***

At first glance the issue of small animals may seem amusing. However, animals frequently use seismic vaults as dwellings. We have seen some very strange "seismic" records caused by ants, grasshoppers, lizards, and mice. Worse, such animals can cause severe damage to cables and other plastic parts of the equipment.

Tight metal (particularly effective), fabric or thermal vault covers usually prevent animals from entering the vault from above. Plastic tubes for cables and drainage should be protected by metal mesh. Placing metal, wool or glass shards in the free space in these tubes also helps. Insecticides can be used to drive away ants and other insects.

In extreme circumstances, animals may be deterred from chewing cables and other equipment by applying paints developed to prevent animal damage to trees.

### **7.4.2.2 Contact with bedrock**

Good contact between seismic sensors and bedrock is a basic requirement. Soil and/or weathered rock layers between the sensor and the bedrock will modify seismic amplitudes and waveforms.

The depth of bedrock and the degree of weathering of layers beneath the surface can be determined by shallow seismic profiling of the site, by drilling (most often too expensive), or by actually digging the vault. Only rarely will a surface geological survey provide enough information about the required depth of the seismic vault (except where the bedrock is clearly outcropping).

If you choose not to carry out a shallow seismic profile, then expect surprises. You will need to dig until you reach bedrock, and that can sometimes be very deep; a vault may have to be repositioned and re-dug if weathered bedrock is extremely deep. These risks make the cost of shallow profiling a wise investment.

A definition of "good" bedrock is necessary when digging vaults without a seismic profile. Unfortunately, the definition is fairly vague, especially because some recent studies show that even a site with apparently hard, but cracked, rock may still have significant amplification compared to true solid bedrock. As a rule of thumb, "good" bedrock is rock hard enough to prevent any manual digging. If profiles are available, P-wave velocities should be higher than 2 km/s.

Seismic vaults are on average 2 to 6 m (7 to 20 feet) deep. At sites where the solid, non-weathered bedrock is outcropping, the required depth is defined solely by the space required for the equipment. One meter (3 feet) or even less may be adequate if the requirements regarding temperature changes associated with the local climate allow. On some highly weathered rock sites, the required vault depth may exceed 10 m (30 feet). In some places a reasonably deep seismic vault can not reach bedrock at all and a borehole installation would ideally be required. Vaults are sometimes still used in such cases for financial reasons. More details on borehole installations are given in 7.4.5.



### 7.4.2.3 Seismic soil-structure interaction and wind-generated noise

The ideas behind the design and construction of seismic stations have greatly evolved in the last few decades. The increased sensitivity of seismometers and the complexity of seismic research, based more and more on waveforms, require very quiet sites and distortion free records. Sixty years ago, seismic stations were usually situated in houses and observatories. Sensors were installed on large, heavy concrete piers, mechanically isolated from structural elements of the buildings, sometimes well above the ground (see Figure 4.2 in the old MSOP; Willmore, 1979; or <http://www.seismo.com/msop/msop79/sta/sta.html> via link “Examples of stations”). Scientists increasingly observed that the interaction between surrounding soil and civil engineering structures in such installations substantially modified seismic signals during seismic events, particularly if the site was on relatively soft ground. Structures swinging in the wind also caused undesired seismic noise, and strong unilateral wind load or insolation on a building’s walls or the rock face of seismometer tunnel entries caused intolerable drifts in long-period or VBB records.

Further evidence arose (Bycroft, 1978; Luco et al., 1990) that every structure at a site modifies seismic waves to some extent. Therefore, today’s seismic stations are mostly ground vaults jutting only a few decimeters (about a foot) above ground level. All buildings, antennae and other masts are positioned well away from the vault to minimize the interaction.

In theory, there is no modification of the seismic signal by the soil-vault structure interaction if the vault's average density (taking into account the empty space in the vault) equals the density of the surrounding soil. However, seismic station design is never based on calculated average densities. The most important factors are that:

- the design is not too heavy, particularly if the surrounding soil is soft;
- all potential buildings and masts are placed away from the seismic vault;
- the vault rises above ground level as little as possible to minimize wind-generated seismic noise.

### 7.4.2.4 Other noise sources

We recommend that seismic stations are fenced, despite the fact that fences usually represent a significant expense. There are a few exceptions, such as stations in extremely remote desert or mountain sites. The fence minimizes seismic noise caused by human activities or by animals that graze too close to the vault. It also contributes to the security of the station.

The optimal size of the fence depends on several factors:

- density of population around the site and human activity close to the station;
- potential agricultural and other activities in the near vicinity;
- the probability of animal interference;
- general seismic noise amplitudes at the site (quiet stations require a bigger fenced area);
- seismic coupling between ground surface and bedrock. Non-consolidated surface ground and seismometers installed on good bedrock allow a smaller fence. A very deep vault has a similar effect.

The smallest recommended fenced area is 10 x 10 m (30 x 30 feet). In the worst case, a fence could be 100 x 100 m (300 x 300 feet). A height of about 2 m (6 - 7 feet) should be sufficient. Light construction with little wind resistance is preferable so that wind-generated seismic noise is minimized.

The equipment and the vault itself can also generate seismic noise. Equipment that includes mains transformers or rotating electromechanical elements like disk drives, diskette drives, cooling fans, etc. should be installed on the vault wall rather than on the seismic pier.

If the vault cover is not firmly fixed to the vault, it can swing and vibrate in strong winds, which can totally ruin seismic records. Be sure that the cover is very firmly fixed to the top of the vault, as its own weight may not be sufficient to prevent vibration in strong wind. When closed and strongly shaken by hand, there should be no play whatsoever between the vault and the cover. If there is, it will cause seismic noise during strong winds.

If a seismic station uses an antenna mast, place it well away from the vault to prevent seismic noise being generated by the antenna swinging in the wind. The required distance is usually between 5 and 50 m, depending on a number of factors such as:

- the maximum expected wind speed and the probability of windy weather at the site (the higher the speeds and the more often they appear, the greater the required distance);
- the antenna's height (the higher the antenna mast, the greater the required distance);
- the vault's depth (the deeper the vault, the smaller the distance);
- the degree of seismic coupling between sensors and antenna base (strong coupling requires larger distances); and
- general seismic noise at the site (very quiet sites require larger distances).

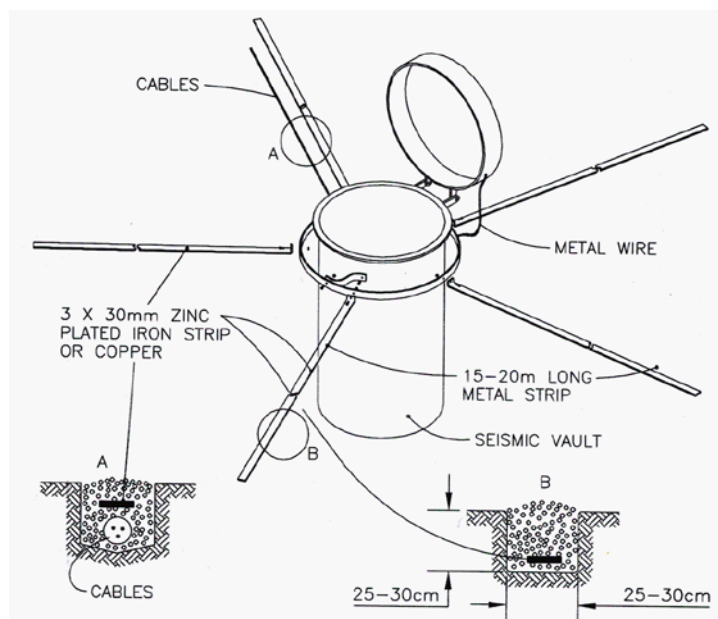
#### **7.4.2.5 Electrical grounding**

A grounding system is required for the proper functioning of electronic equipment. Grounding of equipment and cables keeps the instrument noise low. It is also a prerequisite for lightning-protection equipment and for interference-free RF telemetry. The grounding system design is usually a part of the RF link design in telemetry seismic systems.

A ground impedance below 1 ohm is usually desired. Generally, a radial star configured system, of five to six "legs" with 15 to 20 m (45 - 60 feet) length each, is required for a grounding system (see Fig. 7.47). The total length of the required grounding metal strips depends strongly on climate and local soil type and its humidity. The strips, made of zinc plated iron or copper, 3 x 30 mm (1/8" x 1.5") in cross-section, should be buried from 25 to 35 cm (~1 foot) deep in the soil. In dry regions they should be deeper. The strips should be straight. No sharp turns (around rocks, for example) are allowed because this decreases *lightning protection* efficiency as a result of increased inductivity of the grounding system.

In arid regions, high deserts, or completely stony areas, longer and thicker strips are required. In these cases, a different approach to grounding and lightning protection is sometimes taken by trying to obtain an electric equipotential plane all around the station during lightning strikes. Grounding impedance is no longer the most important issue. High lightning threat regions and very dry or rocky ground usually require a specially-designed grounding system.

In seismic vaults without metal walls, bury a loose mesh made of grounding strips around the vault and connect them to the rest of the grounding system. The grid dimension of this mesh should be around 60 to 100 cm square (~2 to 3 feet square).



**Fig. 7.47** An example of a seismic station grounding system. Note that its dimension depends on local soil humidity conditions.

At seismic stations with RF data transmission and antenna masts, the star-configured grounding system should be centered on the antenna mast, not on the seismic vault. The seismic vault should be included in one of the legs of the grounding system. One of the grounding strips must be laid exactly above the cables connecting the antenna mast and seismic vault (see Fig. 7.47, detail A). This ensures a minimum voltage drop along the cables during lightning strikes and therefore a minimum induced voltage surge in the cables.

The antenna mast itself should be grounded and equipped with a lightning protection rod. Its highest point should be at least 1 m (3 feet) above the highest antenna or solar panel installed on the mast.

Note that any grounding system requires periodic service checks because contacts between the metal parts may slowly corrode. It is recommended that the grounding impedance of the system be checked once every two years. Regular maintenance visits should always include a check of the lightning protection system and equipment and replacement of any burnt-out elements. For more on lightning protection see <http://www.protectiongroup.com/Home>.

#### 7.4.2.6 Vault construction

Seismic vaults can be made with metal walls. Plain iron sheets or corrugated iron can be welded together, or pieces of large-diameter metal pipes can be used. We recommend zinc-plated metal for durability. It is not necessary to make metal vaults very strong and heavy. Water tightness is relatively easy with this design.

If the vaults are made from thin sheet metal (a few mm), then pour relatively thin, 15 - 20 cm (6 - 8") concrete walls around the metal to add strength. The quality and homogeneity of this concrete does not need to be high because water tightness is not a problem. Locally-available sand aggregates can be used in most cases. Such vaults, however, may cause problems if deformation and tilts of the vault due to external temperature changes are important.

The walls can also be made of only concrete – in which case it is easiest to make the vault rectangular. Note that the quality of the concrete must be good to make the vault watertight, as explained earlier. Apart from very deep vaults, strength is not a problem and therefore no steel reinforcement is needed.

At sites where accessibility allows, vaults can be made of the prefabricated concrete pipe sections used in sewerage systems. They are cheap and can be obtained in different diameters and lengths. In deeper vaults you can simply stack them to the required depth of the vault. Care must be taken to ensure that the joints between sections are watertight.

The bottom of the seismic vault – the seismic pier - is always made of high-quality, watertight concrete. Special requirements must be fulfilled for VBB sensors. More details are given in 7.4.2.1 above.

The depth of seismic vaults is determined by seismo-geological parameters. Apart from providing adequate space to put all the equipment, the diameter is primarily a matter of the desired ease of installation, maintenance and service.

For three-component stations with single component sensors, between 1 and 1.5 m<sup>2</sup> (10 to 15 square feet) of space on the seismic pier is required. Less space is needed for three-component seismometers, three-component accelerometers, or a single component sensor. If the vault contains (or will contain in future) three-component weak-motion and strong-motion sensors, about 1.5 - 2 m<sup>2</sup> (15 - 20 square feet) is required.

We have found that a minimum vault diameter for installation and maintenance is 1.4 m (4.5 feet). If the vault is deeper, a 1.5 to 1.6 m (5 to 5.5 feet) diameter is recommended. Deep vaults (> 4 m (13 feet)) require a diameter of at least 1.6 to 1.7 m (5.5 to 6 feet). Vaults deeper than 1.2 m (4 feet) require a ladder.

#### **7.4.2.7 Miscellaneous hints**

##### ***Vault cover design***

A seismic vault cover should have the following:

- at least 5% slope so that water drains quickly;
- vertical siding all around that extends at least 15 cm (6") below the upper rim of the vault to prevent rain from entering in windy conditions;
- a mechanism which firmly fixes the cover to the ground and a lock to mitigate vandalism;
- handles for easy opening and closing;

- be painted a light color, preferably white, that will reflect as much sun as possible, particularly in hot and dry desert regions.

The metal cover and thermal insulation cover of the vault should not be too heavy. They should be designed in such a way that a single person can open and close the vault smoothly and easily. Otherwise, maintenance visits will require two people in the field. For large vaults, the cover can be designed in two parts, or a simple pulley system may help.

### ***Alternative materials***

As material for a vault cover, metal is less appropriate in very hot and very cold climates as it becomes difficult to handle under extreme temperature conditions. UV light-resistant plastic or water-resistant plywood is a better alternative in dry regions. Plywood also has lower thermal conductivity, which improves thermal insulation, and less weight, making handling the cover easier.

### ***Mitigating vandalism***

Experience shows that, apart from political instability in a country, most vandalism of seismic stations is driven by people's curiosity. Therefore we believe that a large sign with a short and easy-to-understand explanation of the purpose of the station and posted at the entrance to the fenced area, may significantly mitigate vandalism.

### ***Fixing seismometers to the ground***

In regions where earthquakes with peak accelerations of 0.5 g or more can occur, seismometers must be firmly fixed to the seismic pier, a common practice with strong-motion sensors. Obviously, sensitive seismometers are clipped during very strong earthquakes. However, they should not shift or move during such events otherwise, the sensors will not be properly orientated for the recording of aftershocks .

## **7.4.3 Seismic installations in tunnels and mines (L. G. Holcomb; 2002)**

Abandoned mines have been used for many years as ready-made quiet sites for installing seismic instrumentation. In some cases, active mine tunnels have proven to be successful, even though they may be somewhat noisy as a result of mining activity during the workday.

Existing tunnels in solid rock provide a low-cost, ready-made and accessible facility that often provides nearly ideal conditions for the installation and operation of high sensitivity seismic sensors. The bedrock in a mine tunnel is usually already exposed, providing an excellent firm foundation on which to install standard surface instruments. If unventilated, as is usually the case for abandoned mines, a mine tunnel provides an essentially constant-temperature environment that is ideal for seismic sensors. Depending on its thickness, the overburden above the mine tunnels provides isolation of the seismic sensors from the seismic noise that is always present at the surface of the Earth.

Obtaining permission to use an abandoned mine property may be difficult, even for non-working mines, because the operational organization or the property owners may quite understandably be reluctant to allow access because of legal liability. Access to working

mines is usually even more difficult because the additional equipment and personnel involved in station activities tend to interfere with mining activities.

Mines are usually concentrated in mineralized zones. It is therefore unlikely that an existing mine will be found near the location of a proposed seismic installation. Tunnels are sometimes constructed solely for the purpose of the installation of seismic sensors. Digging tunnels in hard rock is a very expensive endeavor because tunneling on a small scale is highly labor-intensive.

In many respects, a tunnel installation is very similar to a surface vault installation. A poured concrete floor or pier is usually constructed on the rough bedrock floor of the tunnel to provide a flat and level surface on which to install the sensors. Despite the improved temperature stability found in a tunnel, it is still necessary to provide adequate thermal insulation around the sensors themselves in order to reduce thermally generated noise. Some type of air pressure variation reduction system is also necessary for long period sensors because the air pressure varies in underground tunnels. Usually, this is accomplished in the same manner as it is in a surface installation although sometimes an effort is made to seal off all or parts of the tunnel itself. Sealing a volume enclosed by natural rock walls is difficult because most tunnel walls are riddled with fractures.

However, there are significant differences between surface vault installations and tunnel sites. Rockfall is a real hazard in a tunnel installation. Both personnel and instrumentation must be protected at the actual location of the instruments and along access routes. Another hazard is the build up of harmful gasses (bad air) underground if the tunnel is not adequately ventilated.

The presence of water and high humidity levels in most underground passages is a common problem in tunnel installations. It is very difficult to keep instrumentation dry and the wet environment is frequently unpleasant to work in. The high humidity slowly corrodes the contacts in delicate electrical connectors, which frequently causes poor electrical contact and intermittent operation. The presence of moisture also slowly degrades the effectiveness of thermal insulation materials, and precautions must be taken to prevent moisture accumulation in the isolation system.

Access to power and communication lines is usually more difficult in tunnel installations, depending, of course, on how far the equipment is placed in the tunnel. Frequently, power and or communication lines must be installed throughout the entire length of the tunnel. In the case of power, this can be quite expensive if long distances are involved; either large diameter cables or a high voltage line coupled with a step-down transformer must be installed to ensure that sufficient voltage is available at the site.

Determining the orientation of an underground sensor is considerably more difficult than in a surface installation. Usually, one must transfer an already known azimuth from outside the tunnel to the installation site using standard surveying techniques. Specially designed gyroscopic systems can be used to determine the orientation underground but they are relatively expensive.

It is more difficult to provide timing to a tunnel site than to a vault. This is particularly true for modern GPS based timing systems because the distance between the antenna (outside the tunnel) and the timing receiver is usually limited. Inline radio frequency amplifiers can be used for long antenna runs. It is preferable, however, to place the GPS receiver near the antenna, e.g., at the tunnel entrance. A serial connection can then be used between the

receiver and the recorder either using RS422 (up to 1 km distance) or fiber optic cable. This approach has been used successfully in the Swiss digital seismic network.

#### **7.4.4 Parameters which influence the very long-period performance of a seismological station: examples from the GEOFON Network (W. Hanka, 2002)**

##### **7.4.4.1 Introduction**

The goal for a very broadband (VBB) station for the GEOFON network is to resolve the full seismic spectrum from high frequency (regional events) to very long period (VLP) (Earth's tides) with sufficient dynamic range. The overall instrument noise should remain below the New Low Noise Model (NLNM, Peterson 1993) throughout this frequency range. The GEOFON project (Hanka and Kind, 1994) aims to achieve this goal at minimum cost. This sets strict limits on costs for instrumentation, vault construction and remoteness of the sites.

It is relatively straightforward to get good station performance in the high frequency and medium-period band since the "only" measures to be taken are to get away from man-made noise sources and the sea shore and find a station site on as hard rock as possible. Good VLP performance is usually much more costly to achieve since adequate instrumentation and vaults with sufficient overburden or borehole installations are necessary. However, there are certain measures which can be taken to optimize the VLP station performance in shallow vaults. The parameters to be taken into account for good VLP performance are:

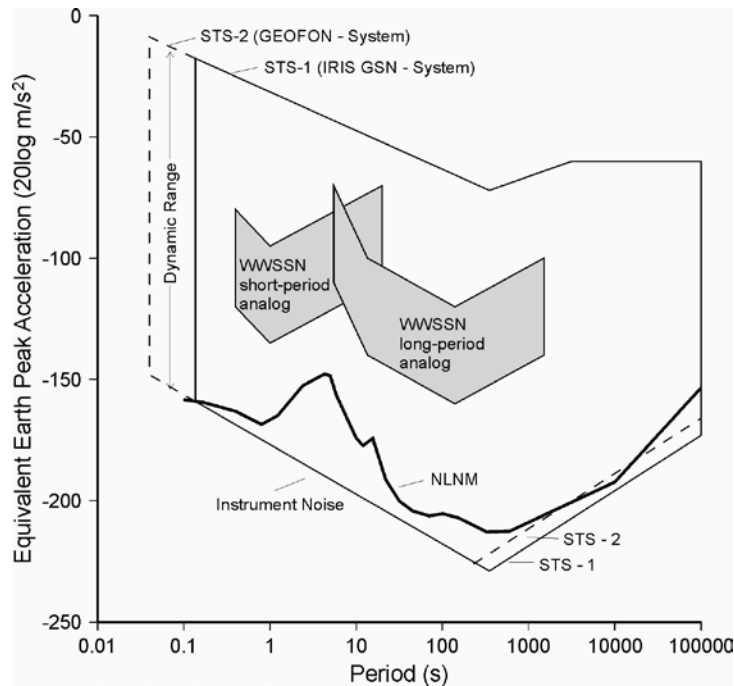
- Instrumentation
- Installation of instruments
- Vault construction
- Geology
- Depth of burial
- General climate

The influence of these different parameters will be demonstrated in the following case studies from the GEOFON network.

##### **7.4.4.2 Comparison of instrumentation and installation**

###### ***Which seismometer to choose?***

The longer the period of ground motion to be recorded, the larger the potential influence of environmental disturbances, such as temperature and air pressure fluctuations and induced ground tilts on the seismic recording, and the larger the need for effective shielding against them. The instrument currently with the best VLP resolution is the Wielandt-Streckeisen STS1/VBB (Wielandt and Streckeisen, 1982; Wielandt and Steim, 1986). It is widely deployed in the IRIS GSN and GEOSCOPE global networks as well as in some regional networks (e.g., MedNet). The permanent GEOFON network uses mostly Wielandt-Streckeisen STS2 and a few STS1/VBB instruments (see DS 5.1) with comparably good results. Fig. 7.48 shows the resolution of the STS1/VBB and the STS2 in relation to the New Low Noise Model by Peterson (1993).



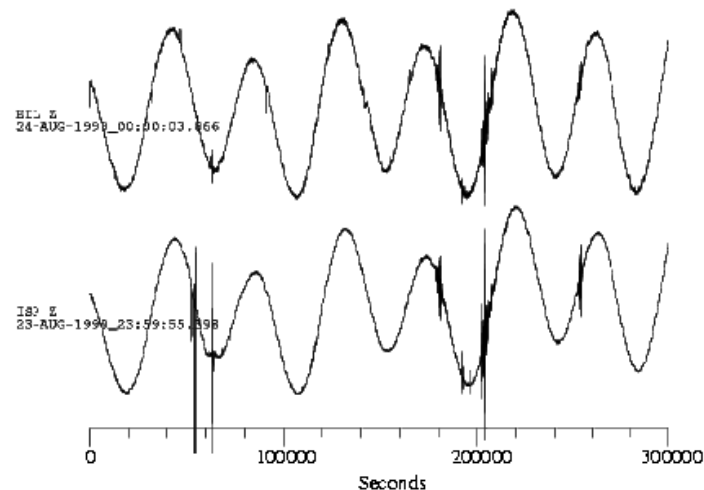
**Fig. 7.48** A representation of the bandwidth and dynamic range of a conventional analog (WWSSN short- and long-period) and digital broadband seismographs (STS1/VBB and STS2 with GEOFON shielding, respectively). The depicted lower bound is determined by the instrumental self-noise. The scale is in decibels (dB) relative to  $1 \text{ m/s}^2$ . Noise is measured in a constant relative bandwidth of  $1/3$  octave and represented by "average peak" amplitudes equal to 1.253 times the RMS amplitude. NLNM is the global New Low Noise Model according to Peterson (1993).

The more compact, lighter and cheaper triaxial STS2 has a pass band with a slightly higher low-frequency corner (0.00833 Hz instead of 0.00278) and a significantly higher high-frequency corner (dashed lines in Fig. 7.48). Depending on the properties of the recording system, 50 Hz can easily be reached compared to the 10 Hz of the STS1. For nearly all sites on Earth, a properly installed STS2 seismometer will give nearly the same performance as a set of STS1/VBB seismometers. The maximum long-period resolution can only be achieved when the seismometers are properly shielded.

The GEOFON project exclusively uses Wielandt-Streckeisen seismometers. The discussion above and below reflects this fact and is not an endorsement of one make of seismometer over another. Potential instrument purchasers need to establish for themselves what instruments are best suited for their own purposes.

The discussion of the shielding efficiency at GEOFON stations in surface or shallow depth vaults or tunnels in the next Chapter is only based on the VLP channel plots (sampling frequency 0.1 Hz) of STS1 records (original or simulated from STS2 records by recursive filtering). The low self-noise of the STS2 allows us to effectively simulate STS1/VBB records down to tidal periods. Fig. 7.49 illustrates this using the recordings of a tidal wave recorded by an STS1/VBB and an STS2. It is difficult to tell the difference between them.





**Fig. 7.49** Tidal recordings of STS1/VBB and STS2 do not differ very much when properly installed in a comparable environment. The two traces were recorded in the Eastern Mediterranean in buried vaults in limestone. At the station EIL (Eilat, Israel) an STS2 with additional GEOFON shielding and at ISP (Isparta, Turkey) a set of STS1/VBB are installed.

### ***Installation of an STS1/VBB***

Seismometers must be shielded against environmental influences, namely pressure and temperature variations as well as magnetic disturbances. The proper installation of an STS2 to achieve good VLP performance is discussed in detail in the next paragraph. Comments on installing the STS1/VBB are kept short here since this is a well known procedure and is described elsewhere (Wielandt and Streckeisen, 1982, Holcomb and Hutt, 1992).

The three separate STS1/VBB components are supplied with different shieldings: a permalloy helmet as magnetic shield (vertical only), an aluminum helmet and a glass bell jar for evacuation. The feedback electronics are placed in a separate container. There are two basic methods used for the installation of STS1/VBB seismometers. The "conventional" one, also suggested by the manufacturer, uses a plane glass plate which has to be cemented to a plane pier. The second method, introduced by Albuquerque Seismological Lab, uses a warp-free rigid stainless steel base plate (similar to the aluminum one used in the GEOFON STS2 shielding) on which the vacuum glass bells and the metal helmets are installed above the actual seismometer. The second method is faster and easier in practice and gives additional flexibility (see Holcomb and Hutt, 1992).

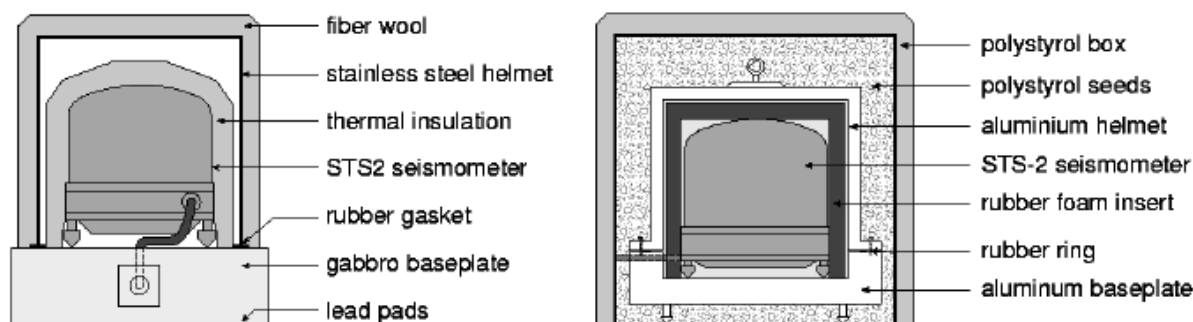
### ***Installation of an STS2***

The STS2 is not supplied with any shielding. All three components and the electronics are contained in a single casing. This casing provides magnetic and pressure shielding to some extent. Nevertheless, temperature shielding is still important in order to obtain longer period signals with a good signal/noise ratio. This is especially important because of thermal convection generated by heat from the electronics. A rather sophisticated shielding (see Fig. 7.50 a) was introduced by Wielandt (1990) for the first STS2 based network, the German Regional Seismograph Network (GRSN). The STS2 was installed on a 10-cm thick gabbro plate covered by an airtight aluminum helmet. Before being covered, the STS2 is insulated with a thermal blanket.

A simpler and more practical approach is used for GEOFON stations (see Fig. 7.50 b). This uses an aluminum casing consisting of a rigid thick base plate (3 cm) and a thinner aluminum helmet with a cylindrical foam rubber insert. As with the gabbro plate, the base plate can not be easily distorted by pressure variations and gives, together with the foam rubber insert, extra thermal stability. In addition, this shielding helps prevent corrosion and is separated from the pier or ground surface by adjustable tripod screws.

The GRSN shielding has extra internal cabling and a socket, whereas the GEOFON casing does not, and the casing is penetrated by the original STS2 cable through a special hole which is made tight with silicon. The GEOFON shielding potentially gives better electrical performance but has worse pressure integrity. The GEOFON shielding is portable and readily available which are problems with the GRSN shielding.

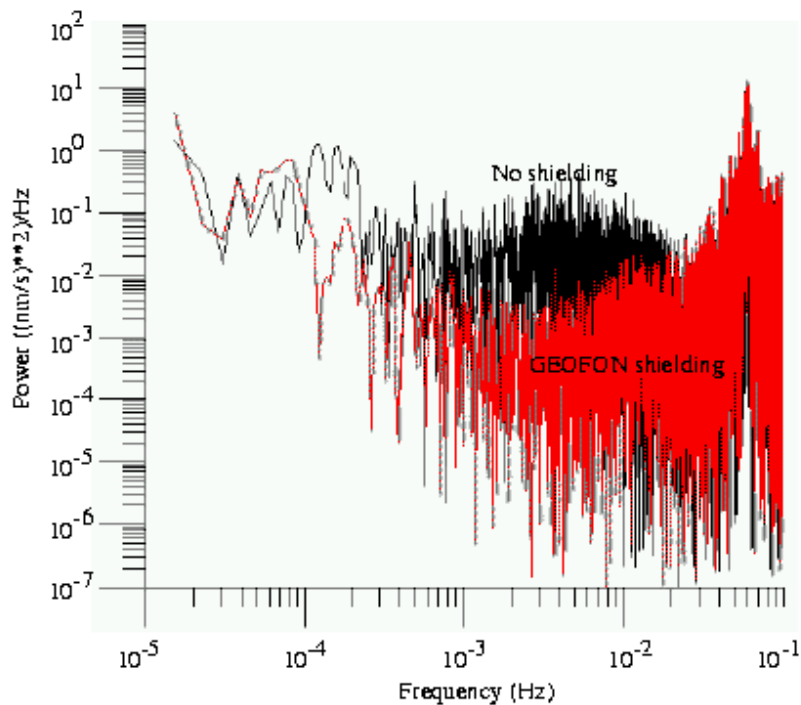
Even better thermal insulation than the one discussed above can be achieved for both installation methods when an additional styrofoam box, completely filled with styrofoam pieces, is used (as shown in Fig. 7.50 b). The box should be tightly glued to the pier or ground surface and the box lid glued to the box walls after filling with the styrofoam beads. Depending on the site conditions, this can give an additional order of magnitude in VLP noise reduction.



**Fig. 7.50 Left:** GRSN shielding after Wielandt (see Fig. 5.16 in Chapter 5) and **right:** GEOFON type shielding for the STS2.

Fig. 7.51 shows the substantial LP and VLP noise reduction which can be achieved even by an incomplete GEOFON type shielding (aluminum casing only, no polystyrol box) in the period range from 30 to more than 10,000 seconds. A reduction of about two orders of magnitude in terms of spectral power (one order of magnitude in terms of amplitude) can clearly be seen between 100 and several thousands of seconds and again around 10,000 sec.

The GRSN shielding gives exactly the same result in most cases. It is only in very rare situations - probably in connection with large air pressure variations – that the performance of Wielandt's approach is slightly better at periods of several hundreds of seconds.



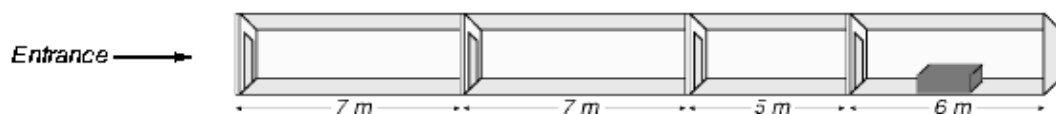
**Fig. 7.51** VLP noise reduction achieved by using the simpler version of the GEOFON shielding method (no additional polystyrol box and beads ). The relative noise power spectra of the vertical component of two STS2s positioned side-by-side are shown. No instrument correction has been applied. The black spectrum is from the unshielded STS2, the red spectrum from the shielded one.

#### 7.4.4.3 Comparison of vault constructions, depth of burial, geology and climate

The harder the rock and the deeper the vault and the more stable the temperature and air pressure remain in the vault, the better is the VLP performance of a VBB station. In contrast, the shallower a vault is, the greater the influence of the general climate.

Fig. 7.52 shows the scheme of an artificial tunnel vault which is used at several IRIS/USGS installations in cases where no other existing underground facility can be used. The tunnel is about 25 m long and segmented using four doors (air locks). The last chamber contains the large seismometer pier. Since the tunnels are drilled into mountain slopes, depth of overburden is of the order of the tunnel length.

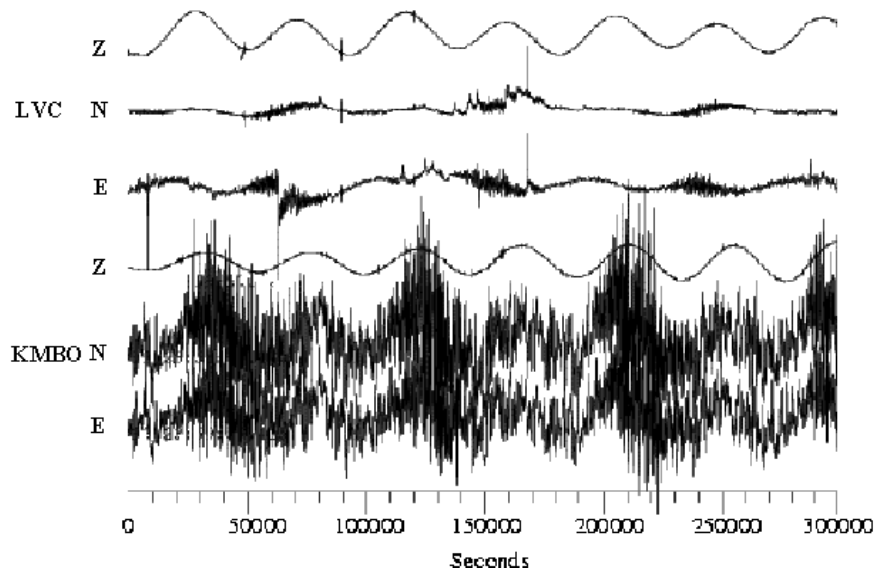
#### *Tunnel vaults*



**Fig. 7.52** Sketch of an artificial horizontal tunnel construction with different chambers to host a VBB seismological station. This type of construction is widely used within the IRIS/USGS

part of the IRIS GSN network. The total length of the tunnel is approximately 25 m. The construction cost of such a vault can reach up to US\$ 100,000 depending on local conditions and infrastructure.

Although the vault construction is identical, the VLP performance at different sites is not. This is shown by the recordings Earth's tides in Fig. 7.53. The tunnel of the IRIS/GEOFON station LVC (Limon Verde, Chile) is built in hard basaltic rock and the traces show remarkably low VLP noise, while at KMBO (Kilima Mbogo, Kenya, also an IRIS/GEOFON site) a soft volcanic conglomerate drastically increases the noise, especially on the horizontal components. Another effect which can clearly be seen on the horizontal components is the large day-night noise variation. The general temperature increase and perhaps also the deformation of surface rocks caused by direct sunshine during the day, as well as stronger winds cause substantially larger VLP noise levels on the horizontals at both sites. This shows that even this kind of sophisticated and expensive tunnel vault construction gives no guarantee of seismic recordings free of environmental influences.

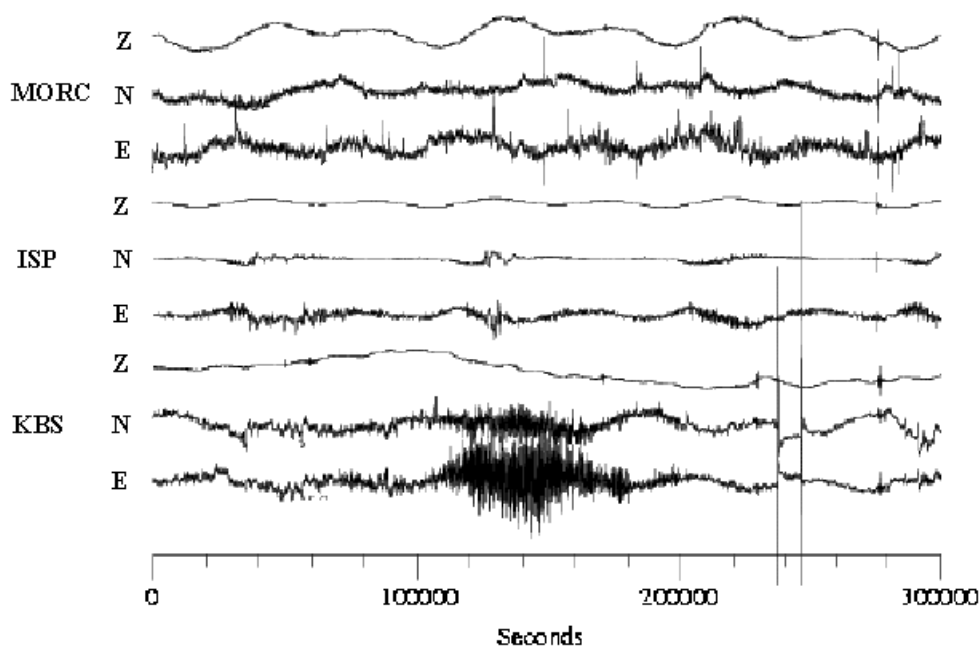


**Fig. 7.53** Comparison of two 3-component STS1 VLP traces recorded in identical tunnel constructions but in different geological and climatological environments. LVC (Limon Verde, Chile) is built in hard basaltic rock in a full desert environment, KMBO (Kilima Mbogo, Kenya) is placed in rather soft volcanic conglomerate influenced mostly by a humid tropical environment. Day-to-night temperature gradients are high in both cases.

#### ***Shallow vaults***

If tunnel vaults are not affordable, other less expensive methods of getting the seismometers sufficiently buried have to be used. Several cases are discussed below.

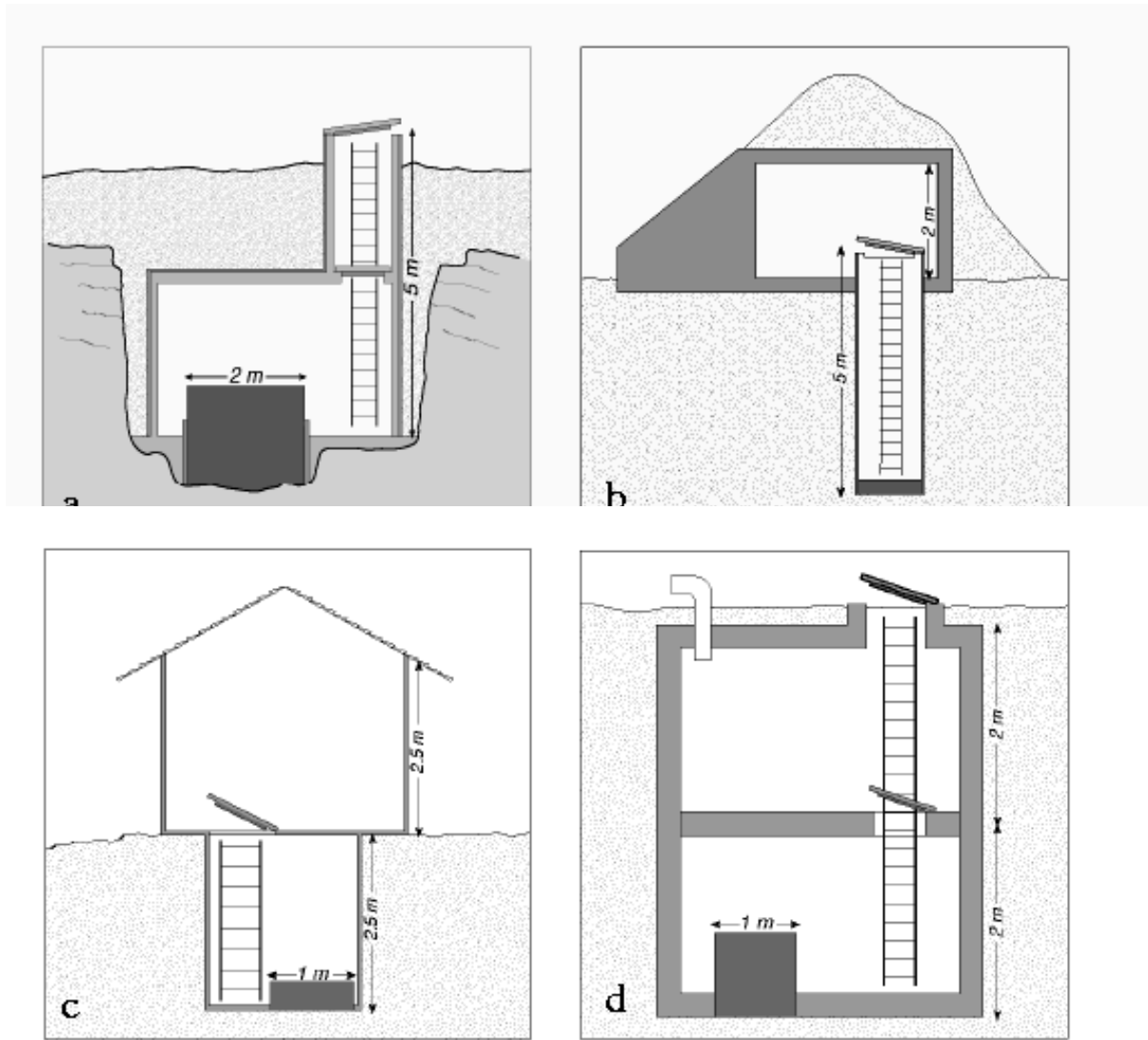
Fig. 7.54 compares the recordings made at three different stations. The depth of burial is only about 4-5 m in all cases, which is very poor compared to tunnels. Nevertheless, the moderate climate at MORC (Moravsky Beroun, Czech Republic) and at ISP (Isparta, Turkey) gives a relatively good VLP performance. These vaults are built in hard rock and limestone, respectively. The spikes which can be seen mainly on the horizontal traces are due to human activity close to the site. In the arctic climate of KBS (Ny Alesund, Spitzbergen = Svalbard), the more drastic temperature changes cause increased VLP noise level on the horizontals. Here, there are also some spikes caused by local man-made disturbances.



**Fig. 7.54** Comparison of three 3-component VLP records in shallow vaults (4 - 5 m). At MORC (Moravsky Beroun, Czech Republic) an STS2 is installed in a 1 m wide borehole in hard rock; in ISP (Isparta, Turkey) and KBS (Ny Alesund, Spitsbergen) sets of STS1/VBB are installed in underground bunker vaults in limestone and weathered rock (permafrost), respectively.

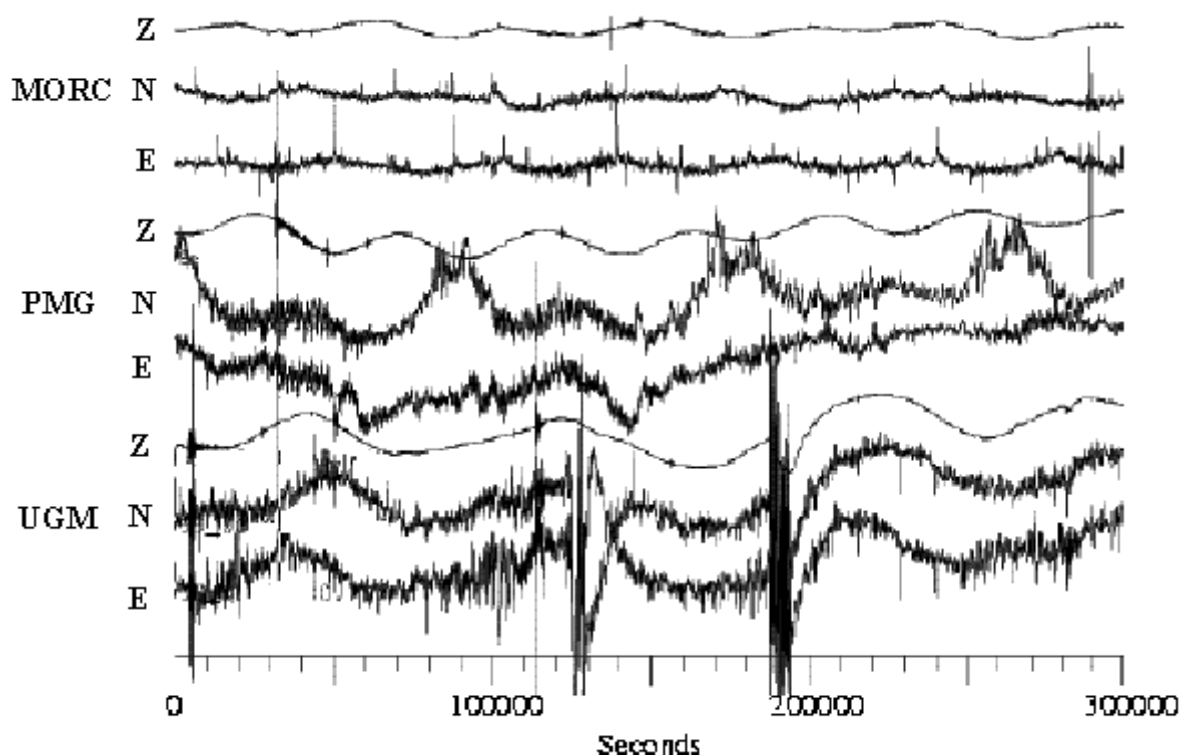
The vaults at KBS and ISP are very similar: about 5 m deep large underground concrete bunkers with large concrete piers for the installation of the STS1/VBB seismometers (see Fig. 7.55 a). The geologies are different: weathered rock in permafrost (KBS) and limestone (ISP). The recording system at KBS is located elsewhere, while at ISP, recording is local in a house built above the vault. A very different construction is used at MORC: a very wide shallow vertical borehole has been drilled into hard rock and a one-meter wide steel tube placed into

it, with a concrete floor on the bottom. The STS2 in GEOFON shielding has been installed on this at about 5 m depth (see Fig. 7.55 b). Here and in the two other examples of construction schemes for STS2 stations (see Figs. 7.55 c and d), a recording room hosting all computer and communication equipment is located above the seismometer vault.



**Fig. 7.55** a) Underground bunker vault construction for the installation of a set of STS1/VBB (remote recording); b) "wide & shallow borehole" vault construction for the installation of an STS2; c) and d) simple bunker vault construction schemes for an STS2. The vault constructions b - d allow onsite data recording thanks to the existence of a separate recording room.

Fig. 7.56 shows, again in comparison to MORC, the recordings at shallow vaults in locations near the equator. At PMG (Port Moresby, Papua New Guinea) a two-room underground vault hosts a set of STS1/VBB seismometers. The two-room construction is situated in a sedimentary layer above rock and is comparable in size with the one at KBS and ISP, but shallow (3 m) and with a horizontal entrance into the first (recording) room. UGM (University Gadjadara, Yogyakarta, Indonesia) uses a very simple, 2.5 meter deep bunker in limestone (construction after Fig. 7.55 c) with an STS2 and a small open recording hut above. Both show rather similar results to MORC, especially on the horizontals. The extreme large amplitudes at UGM during daytime are caused by human activity close to the station.

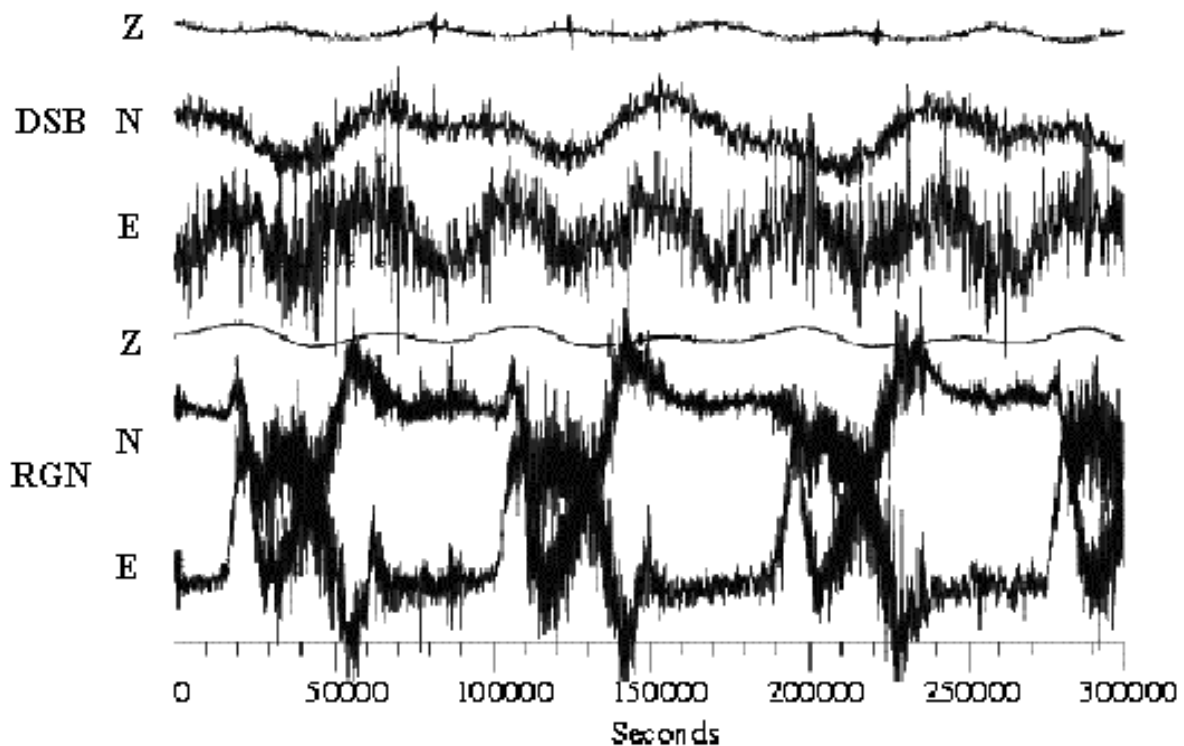


**Fig. 7.56** Comparison of three 3-component VLP records from shallow vaults in rock in different climates. Data from two sites close to the equator (Port Moresby, Papua New Guinea, PMG, and Wanagama, Indonesia, UGM) are shown together with data from MORC (same station as in Fig. 7.54).

In principle, the VLP station performance is not so different at both equatorial sites, particularly the horizontal components which are not as good as at a site in a more moderate climate. The instrumentation and construction details do not play any significant role in determining the VLP noise performance.

#### ***Surface vaults in moderate climate***

The STS2 records in Fig. 7.57 were obtained in a simple above-surface vault on rock (DSB) and a very shallow vault in soft sediments (RGN). Both sites are located in an area with a very moderate climate and close to the sea. Temperature shielding is a little better at RGN due to complete soil coverage on three sides and up to one meter on top. Therefore the general VLP performance - as seen on the vertical components - is better at RGN, but the horizontals show large additional distortions during daytime. These are most likely caused by temperature-induced swelling and related up-bending of the sand hill which is a very typical behavior for sediments (it can also be seen to some extent on the PMG records in Fig. 7.56). This is not the case with rock at DSB. It is remarkable that there is almost no day-night variation on the DSB records although the vault is completely above the surface. This is due to the maritime climate with very small day-night temperature changes.



**Fig. 7.57** Comparison of two surface vaults in moderate climate on rock and in sediments. At DSB (Dublin, Ireland), a small surface bunker was built in an old granite quarry. At RGN (Rügen Island, Germany), an old one-room military bunker with a thin ( $< 1\text{m}$ ) soil cover on top is used. Both sites host STS2 seismometers with GEOFON shielding.

#### ***Surface vaults in arctic climate***

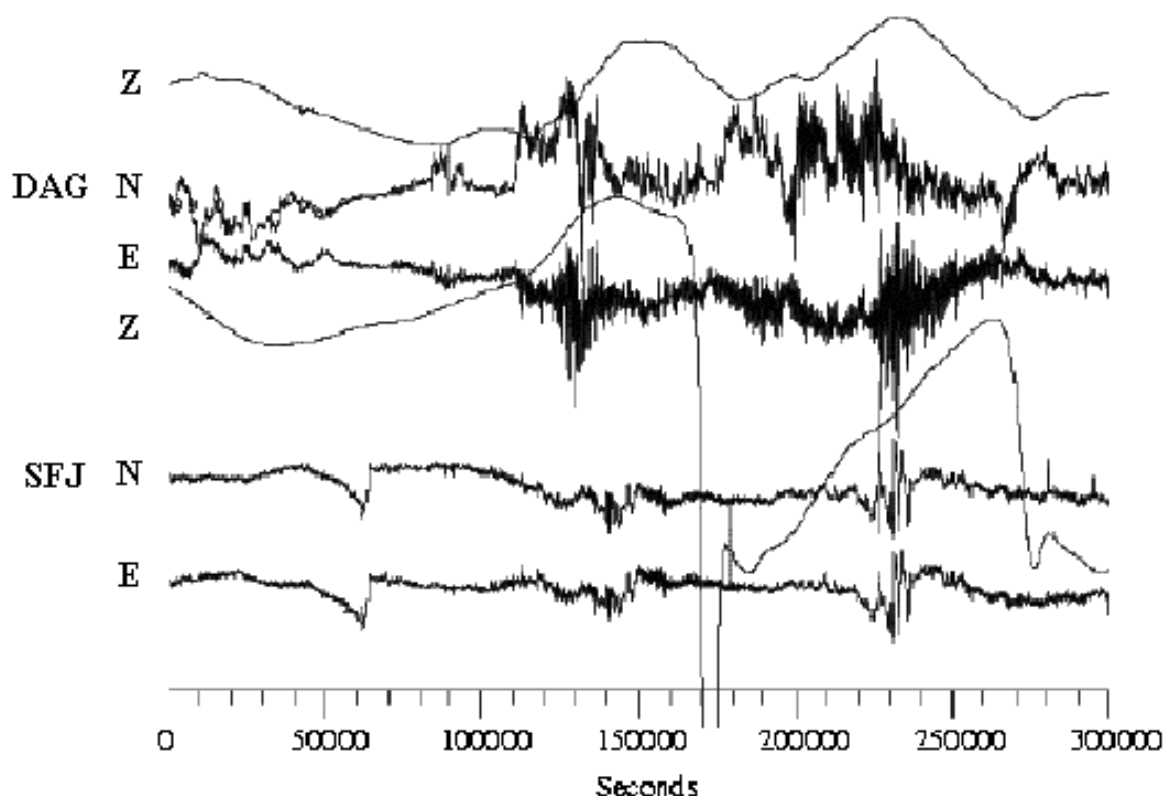
The very poor VLP noise performance of surface stations in arctic climates can be seen in Fig. 7.58, where data from two stations in Greenland are shown. Both vaults are located in surface wooden huts built on weathered rock, more or less open to all kinds of atmospheric turbulence in terms of air pressure and temperature changes. The vaults are heated in winter. This results in about the worst conditions one can imagine for VLP noise performance. Earth's tides are no longer seen very clearly and the daily noise variations are large. However, there is no other choice in these regions. DAG is in one of the most remote places on Earth where it is almost impossible to build an underground vault.

#### **7.4.4.4 Conclusions**

The VLP performance of a VBB seismological station is directly dependent on several instrumental and environmental parameters. High quality VBB seismometers, a true 24-bit A/D converter and a continuous multi-stream data recording are essential. In the GEOFON network, only STS1/VBB and STS2 seismometers and Quanterra data loggers are used for this reason. With the appropriate shielding, the VLP performance of the STS2 is not much different from the STS1/VBB. Only in very rare cases at extremely quiet sites can the extra infrastructure, installation, maintenance and financial efforts related to the usage of STS1/VBB sensors be justified. The same is true for vault construction. The construction



scheme itself has not much influence on the station performance as long as the depth of burial is deep enough and the environmental disturbances can be reduced to a minimum. With an adequate casing, a seismometer pier is not required to install an STS2 sensor properly underground. The geology plays a very important role. The harder the rock, the lower is the VLP noise at a certain depth since surface tilts caused by atmospheric influences do not penetrate as deep. Sediments show special tilting effects, which reduce drastically the daytime VLP performance of the horizontal components. The shallower a vault is, the more the influence of the general climate. In very moderate climates, e.g., close to the sea, even surface vaults can have a reasonable VLP noise level. In summary: Although the task of establishing a VBB station that is capable of recording with sufficient dynamic range the full seismic spectrum from high-frequency regional events up to the very long-period (VLP) Earth's tides seems to be a very difficult and costly effort, it can be achieved with rather simple means.



**Fig. 7.58** VLP records obtained at two surface vaults on Greenland. At DAG (Danmarkshavn, NE Greenland) a STS2 in GEOFON shielding is installed in a wooden hut on weathered rock close to the sea shore. At SFJ (Sondre Stromfjord, SW Greenland) a set of STS1/VBB is located in a container-like building on top of a mountain. In both cases the geology is weathered rock in a permafrost environment.

More details on the installations made at various VBB stations and the comparison of noise data can be found on the web page <http://www.gfz-potsdam.de/geofon/>.

## **7.4.5 Broadband seismic installations in boreholes (L. G. Holcomb, 2002)**

### **7.4.5.1 Introduction**

Borehole seismology is a relatively new technology that has developed over the last 30 years or so. In the early years of seismology, installing a seismometer in a borehole was virtually impossible because of the relatively large physical size of instruments. As seismological technology matured, the instruments became smaller and it became more practical to consider borehole installations as alternatives to surface vaults or tunnels. There are several practical reasons for placing seismic instrumentation in boreholes; these include reduced noise levels, temperature stability and reduced pressure variability.

Experience gained over many years of installing both short- and long-period instruments has shown that sensor systems which are installed at depth are usually quieter than those installed at or near the surface of the Earth (see 4.4.). This is why abandoned underground mines are frequently used as sites for low-noise seismological stations. However, abandoned mines are not always found at the desired location of a seismic station. A borehole provides a practical solution to the need to install seismic sensors at depth almost anywhere.

A borehole is also a very stable operating environment in which to operate sensitive instruments because the temperature at depth is very stable and the pressure in a cased sealed borehole is very constant. Temperature changes and pressure variations at frequencies within the pass band of the sensor system are common sources of seismic noise (see 7.4.2.1). Systems installed on the surface or in shallow vaults require extensive thermal insulation systems in order to reduce the influences of temperature to acceptable levels. Similarly, elaborately designed pressure containers are required to eliminate pressure-induced noise particularly at long periods in vertical instruments. Both temperature and pressure considerations have become more important with the advent of broadband instruments because these instruments are sensitive to outside influences over a broader frequency range thereby making it more difficult to sufficiently isolate broadband instruments from extraneous influences. A sealed borehole of only moderate depth provides excellent temperature stability because of the tremendous thermal mass and inertia of the surrounding Earth. Furthermore, most seismic boreholes are cased with steel casing whose cylindrical walls are quite thick; this casing constitutes a quite rigid container, which greatly reduces atmospheric pressure variations within the borehole (assuming that both the top and bottom are sealed).

Boreholes are frequently considered to be expensive, but they sometimes represent the only practical alternative if an abandoned mine is not available. Excavating tunnels purely for seismological purposes into competent rock deep enough to provide sufficiently quiet seismic data is also a very expensive solution (see below). In many cases, a borehole may actually be the cheapest method for achieving an installation at depth unless the local manual labor costs are very low.

One advantage of a borehole installation over a vault is that there can be much less surface equipment on site, especially if no recording equipment is deployed on the site, say in a seismic array or small network. This can significantly save on costs and improve security. These advantages have led, in some cases, to the use of very shallow boreholes, or postholes, which are drilled to depths similar to vaults.

It is impossible to state exactly how much it would cost to construct either a borehole or a tunnel type vault because too many factors are involved. Precise costs will depend on the type

of material in which the facility is constructed, raw material costs, local labor costs, etc.. However, here are some examples of approximate costs that have been encountered in constructing facilities for the IRIS program over the past 5 to 10 years. In Africa, IRIS has excavated three tunnel type seismic vaults that extended 25 to 40 meters horizontally into hillsides. The costs of these three projects ranged from approximately US\$ 150,000 to US\$ 250,000. For a typical borehole (100 meter deep), project costs range from approximately US\$ 25,000 to US\$ 200,000 at large landmass sites with boreholes in hard rock being significantly more costly than in soft soil. On the other hand, at small isolated Pacific Ocean island sites, borehole costs are in the US\$ 150,000 to US\$ 250,000 range.

#### **7.4.5.2 Noise attenuation with depth**

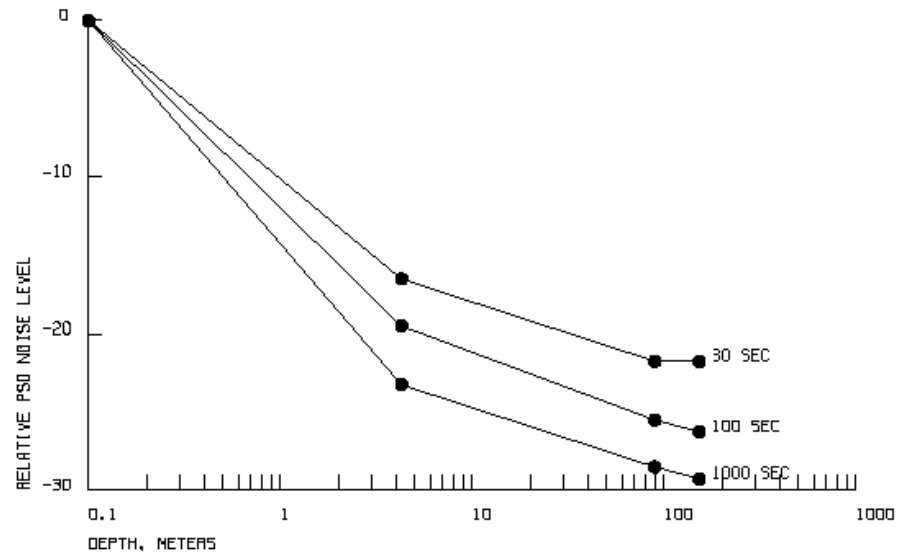
The main reason for installing broadband sensors in boreholes is to reduce the long-period tilt noise which plagues horizontal sensors installed on the surface. The question commonly asked by seismologists who are contemplating a borehole installation is how rapidly does the tilt noise decrease with depth and so how deep does the borehole need to be. There is no easy answer to this question because a borehole never eliminates all of the long-period tilt noise however deep it is. In general, the noise attenuation rate (db per unit depth) decreases as the depth increases; most of the noise reduction occurs in the upper parts of the borehole.

Fig. 7.59 illustrates the attenuation of long-period horizontal noise with depth. It shows the relative power spectral density (PSD) noise levels obtained from the simultaneous deployment of four broadband sensors located close to one another at the same site and installed at various depths. The first sensor was installed in a small vault on or near the surface. Three other three sensors were installed in boreholes at depths of 4.3, 89 and 152 m below the surface. The site consists of about 18 m of unconsolidated (soft/weathered) overburden overlying fractured Precambrian granite bedrock. In Fig. 7.59, noise attenuation data points in db relative to the noise level in the surface sensor are plotted for periods of 30, 100, and 1000 seconds. Note the very rapid decay in the noise level over the first few tens of meters followed by a much slower rate of decrease in noise levels at greater depths. Note that, in general, a depth of 100 m is sufficient to achieve most of the practicable reduction of long period noise.

The data in Fig. 7.59 should only be regarded as an example of noise attenuation with depth. Apparent surface noise levels at a particular site are frequently highly dependent on the methods used to install the instrumentation. This is particularly true of noise levels at many surface installations where faulty installation of broadband horizontal sensors causes excessive tilt noise at long periods.

Choosing the optimum depth for a borehole for a particular site involves comparing the cost of drilling the borehole to a given depth against the desired data quality, the anticipated surface noise levels (they are frequently determined by the anticipated wind speeds and wind persistence at the site), and the depth of the overburden at the site. Unfortunately, studies detailed enough to yield the precise relationships between the various factors have never been conducted. Therefore, choosing the depth of a borehole for a particular site usually involves non-quantitative consideration of the various factors involved. Many years of experience has demonstrated that 100 meter deep boreholes drilled at sites with a few tens of meters of overburden overlying relatively competent bedrock will provide a sufficiently quiet

environment for installing a high quality borehole instrument. Most broadband IRIS borehole instruments are installed at or near 100 meters depth. Boreholes at sites with more overburden and/or softer lower quality bedrock are sometimes deeper depending on construction costs and anticipated surface noise levels.



**Fig. 7.59** Horizontal surface noise attenuation as a function of depth at three selected periods. The depths were 0, 4.3, 89, and 152 meters.

#### 7.4.5.3 Site selection criteria

There are several criteria for selecting the site for a borehole installation. Ideally, one should select a site at which the surface background seismic noise over the band of interest is as low as possible. However, there are other factors such as accessibility, availability of power, improved network configuration, the presence of wide-spread thick alluvial fill, and/or the presence of cultural activity within the monitored area, which may force the choice of a site with higher background noise levels.

A good borehole should penetrate well into bedrock (70 to 100 meters) (see 7.4.2.2), so the site should have bedrock at or near the surface to minimize the need to drill through excessive overburden. If possible, the bedrock should be a relatively hard rock (see 7.1.2.2) such as granite or quartzite. Harder, more competent rock increases the rate of attenuation of surface noise with depth and also decreases the chances of borehole collapse during drilling. Soft rocks such as shale, mudstone, or low grade limestone should be avoided if possible.

Good bedrock is highly desirable for providing the best results from a borehole installation, but benefits are still there for boreholes in poorer rock. Note that the first data point in Fig. 7.59 (only 4 meters down) was obtained in a very shallow borehole that was drilled entirely in loose alluvial fill. Therefore, the lack of shallow bedrock should not preclude the consideration of a borehole installation for a particular site.

As with vault and tunnel installations, a reliable source of electricity will be necessary to power the site, a shelter will be needed to house the recording equipment, and some form of communication capability (telephone line, internet connection, RF or satellite link) is

frequently desirable (see IS 8.2). Accessibility for both the drilling equipment and maintenance personnel (see also 7.1.2.4) should also be considered during site selection activities.

Unfortunately, the need to be able to provide adequate security is also becoming a major factor in selecting station sites in many parts of the modern world. There is little point to investing in a good site if it can not be protected from vandalism. Adequate security has many different meanings depending on the particular situation. It may be as simple as a passive protective fence or as elaborate as alarmed fences and entry ways or even an on-site caretaker depending on the anticipated level of potential damage.

It should be noted that stations on very small islands (such as most coral atolls) do not benefit from borehole installations because the ground motion generated by ocean-wave loading of the beach penetrates rather deeply into the subsurface environment. For this reason, all borehole sites should be at an adequate distance (at least several km) from any coastline.

#### **7.4.5.4 Contracting**

Seismic boreholes are usually drilled by a local contractor using specifications supplied by the organization building the station. Hiring a local driller helps reduce mobilization and setup charges, which are frequently a significant portion of the cost of a seismic borehole. Specifications should be rigid and specific enough to ensure that the finished borehole will be suitable for seismology but flexible enough to prevent excessive costs. Most drilling contractors will have little or no experience of seismic boreholes and it is recommended that the contracting agency use an independent expert with extensive drilling and casing experience whose duties include on-site observation and supervision of all drilling and casing operations. This precaution is advisable to ensure that the drilling contractor performs all operations according to the specifications because departures from specifications are hard to detect, document, and prove after the project is finished. The contract should be specific about who is responsible for unexpected difficulties which might arise during the drilling and casing operations; courses of action should be specified if operations are delayed for any reason whatsoever. These include but should not necessarily be restricted to on-site down time which might be due to bad weather, shortage of drilling materials, crew availability, drill rig breakdowns, loss of circulation, injuries on the job, delays in subcontractor availability, holidays and unexpected changes which might be encountered in the quality of the subsurface rock.

#### **7.4.5.5 Suggested borehole specifications**

The drilling specifications for a seismic borehole should be written in such a way as to ensure that the completed borehole will be suitable for acquiring high quality seismic data. Parameters such as borehole verticality, depth, diameter, and casing type must be clearly specified. It is also important to specify how these parameters will be measured during construction or in the finished borehole.

Borehole verticality is the specification which drillers have most trouble meeting. Borehole verticality must be specified because all borehole seismometers have only a limited range of tilt over which their mechanical internal leveling mechanisms operate. Therefore, the sensor

package must be aligned within a given tolerance from true vertical. This in turn requires that the borehole itself be aligned within a certain tolerance of true vertical. The required verticality specification will depend on which borehole seismometer is to be installed in the completed borehole because each seismometer has a unique mechanical leveling range (typical examples: CMG-3TB has a 3 degrees range, the KS-36000 has 3.5 degrees and the KS-54000 has 10 degrees). In general, the closer the verticality specification requirement is to vertical the higher the cost of the borehole.

The working depth of the borehole is usually specified as the depth of the open cylinder within the borehole confines after construction is complete. The driller is usually left with determining the depth of the hole to be drilled in the rock in order to achieve the desired working depth.

Most boreholes are cased with standard casing used in oil fields because it is readily available throughout the world. This casing is usually specified in terms of its outside diameter (OD) and its weight per unit length; the combination of these two parameters determines the wall thickness and in turn the inside diameter (ID) of the casing. The seismometer manufacturer usually recommends a range of casing in terms of the ID's of the casing in which his sensor will operate satisfactorily. These two methods for specifying borehole diameter must not be confused when writing specifications. As an example of typical hole diameters, a KS-54000 requires a casing with at least a 15.2 cm ID whereas, if equipped with proper hardware, a CMG-3TB (see DS 5.1) will fit into a slightly smaller casing. The specification usually permits the use of a range of OD's and weight specifications in order to facilitate acquiring the casing locally to decrease shipping costs. The individual threaded casing sections should be assembled together with a thread sealing compound and enough torque to ensure that each joint is properly sealed against leakage.

The bottom end of the casing is often equipped with a one way valve (called a float shoe) to seal the lower end against water entry and to facilitate cementing operations. This device allows the cementing mixture to be forced out of the bottom of the casing and prevents water from entering the borehole once the cementing operation is completed.

The casing must be firmly cemented to the surrounding rock walls of the borehole in order to ensure good mechanical coupling. The cementing operation usually consists of pumping a premixed cement mixture down the inside of the casing, out through the float shoe at the bottom, and forcing it back up to the surface between the casing and the bedrock. This operation ensures that all of the annular volume between the steel casing and the rock is filled with cement without voids containing air or liquid. When return cementing mix is observed in the annulus at the surface, a cleaning plunger is forced down the inside of the casing with water under high pressure. This expels the cement mix contained within the casing volume out of the bottom through the float shoe and finally sets (locks) the one way valve within the float shoe to prevent fluids from re-entering.

After the cement has set, it is advisable to require the driller to perform a leak test to ensure that the casing has been adequately sealed. Leak testing usually consists of first pressurizing the water-filled borehole to a specified pressure, sealing it off and leaving the pressurized borehole for a specified time period. The pressure within the borehole should not drop more than a pre-specified amount.

The upper end of the casing is normally terminated with a "packoff" device. This assembly is normally provided and installed by the contracting organization at the time of seismometer

installation. The packoff unit seals the top of the borehole and provides a means of passing instrumentation cables into the borehole.

#### **7.4.5.6 Instrument installation techniques**

It is a relatively simple operation to install a borehole sensor but certain precautions are required. The sensors are usually fitted with two cables. The first cable is intended to provide sufficient strength to lift the weight of the sensor and any extra pulling force required to removing the sensor from the borehole. This is usually a steel cable or "wire rope". The second cable contains the electrical connections for power, control of the various mechanical operations within the sensor, and to transmit the seismic signals back up the borehole. For holes of significant depth, a small lightweight electrically driven winch and mast assembly can be used to lower the sensor into the hole and to retrieve it if necessary. Lowering and raising the sensor should be done fairly slowly because the sensor package sometimes catches on the casing pipe joints as it moves up or down the borehole. On the way down, this problem is usually temporary but usually results in a short free fall of the sensor and a sudden stop when the load-bearing cable becomes taut. If severe enough, the sudden stop can damage a sensitive instrument. If the sensor catches on a pipe joint on the way up, tension in the load bearing cable rapidly increases to dangerous levels if the winch is not stopped in time. If the sensor disengages from the pipe joint while the lifting cable is under high tension, the sensor will undergo possibly damaging levels of acceleration. If the sensor does not disengage and if the winch is powerful enough, the lifting cable may break and endanger personnel.

It is advisable to carry out a dummy run in the completed borehole using a metal cylinder with similar dimensions and weight to the seismometer package. This will help minimize the risk of damage to or losing the equipment during installation. Such a dummy run could be part of the acceptance procedures for the drilling contract.

Traditionally, borehole seismometers are rigidly clamped to the inside of the cased borehole with manufacturer-supplied mechanical hardware to ensure adequate coupling between the sensor and ground motion. The hardware usually includes a mechanically driven locking mechanism for clamping the sensor to the walls of the borehole. This device sometimes consists of a motor driven or spring loaded pawl that is extended on command from the side of the sensor package to contact the borehole wall opposite the sensor (GS-21, CMG-3TB). Sometimes this function is performed by a separate piece of hardware known as a "holelock" that is clamped into the borehole and on which the sensor package is subsequently placed (KS-36000, KS-54000, and earlier Guralp sensors). In the second case, additional hardware is sometimes required to stabilize the upper end of the sensor package (the centralizer assembly in KS instruments). Mechanical clamping mechanisms have been used successfully for many years and have produced satisfactory data from many installations

However, many installations of this type produce more long-period noise in the horizontal components than in the vertical component. In some of these installations, the horizontals were orders of magnitude noisier at long periods than the vertical. The source of the excess noise in the horizontal components has been difficult to isolate and eliminate. For many years, it was suspected that some of this noise was somehow generated by air motion in the vicinity of the sensor package. Conventionally designed horizontal components of long-period seismometers (this includes all sensors with garden-gate type of suspension such as the STS1, CMG-3 series, KS-36000 and KS-54000) are extremely sensitive to tilt because of their inability to separate the influences of pure horizontal acceleration input to the sensor frame

(the desired input) from the signal that arises from the tilting of the sensor package (tilt noise). Therefore, fairly elaborate schemes for reducing the potential for air motion around the sensor within the borehole have been devised and utilized with varying success. Through trial and error, it has become customary to wrap the sensor package (KS-36000's and KS-54000's) with a thin layer of foam insulation in an attempt to somehow modify the flow of heat near the seismometer in the borehole. In addition, it has become common to place long plastic foam borehole plugs immediately above these sensor packages deep in the borehole and near the top of the borehole to block air motion in these sections of the borehole. Additional insulation, which is intended to further reduce air motion within the borehole, is sometimes utilized near the top of the sensor package.

Recently, a highly successful method for significantly reducing the long-period noise levels in borehole installed horizontal components has been developed at the Albuquerque Seismological Laboratory. It consists of simply filling the entire empty air space below and around the sensor package with sand. In this type of installation, none of the auxiliary installation hardware such as the borehole clamping mechanism or holelock, the azimuth ring, the pilot probe, the centralizer, the foam plugs and/or insulation are utilized to install the seismometer. The seismometer package is simply lowered onto a bed of sand at the bottom of the borehole - sometimes, a piece of hardware called a sand foot is installed on the bottom of the sensor. A volume of sand is then poured into the borehole to a depth extending to the top of the seismometer package. The volume required can be easily calculated from the dimensions of the package and the inner diameter of the borehole.

Experimental investigations have demonstrated that it is easy to remove the seismometer from the sand if necessary for maintenance purposes even when the sand is saturated. Normally, the sand left in the hole from a previous installation is not removed from the hole prior to the next installation. Only a fraction of a meter of borehole depth is lost per installation; if necessary, the sand can be removed from the borehole with a downhole vacuum cleaner that has been designed at ASL.

This method of installation is expected to reduce horizontal noise to levels approaching the noise level of the vertical component at any particular site. The horizontals should be expected to always be slightly noisier than the vertical component because remnant real ground tilt will always be present regardless of how deep the sensor is installed. To date, extensive testing at ASL utilizing both KS and CMG (see DS 5.1) instruments and several actual KS sand installations in the field have indicated that sand does indeed produce significantly reduced levels of horizontal noise. The sand installation method has been adopted for future installations by the IRIS GSN program.

One additional advantage of a sand installation is that the seismometer package costs considerably less than for a clamped installation.

One note of caution should be introduced at this point. Conventional hole-lock based installations produce very noisy horizontal data if the sensor package is immersed in water or another liquid such as motor oil. Therefore, every effort is normally made in the field to keep liquids out of the borehole. Although not thoroughly tested to date, sand installations are expected to provide quiet horizontal data even if the sensor is immersed in water as long as the water is not flowing.

Determining the orientation of the horizontal components of a seismometer installed in a deep borehole is not a simple matter because one can not physically get at the instrument once it is



installed. One must resort to indirect methods for determining how the instrument is oriented. For the past 25 years, the KS series instrument installations have relied on a gyroscopic procedure to determine the seismometer orientation as follows. First, the hole-lock is installed in the borehole at the intended operating depth; then, a gyroscopic probe is lowered into the hole and mated with an alignment slot in the hole-lock. The gyro system determines the orientation of this alignment slot with respect to a known azimuth (usually north) on the surface. An adjustable azimuth ring located on the base of the KS sensor is then set to compensate for the alignment of the hole-lock slot to north. This ensures that when the seismometer is lowered into the borehole and the key on the alignment ring is mated with the alignment slot in the hole-lock, the sensor is in a north-south, east-west orientation.

This method was considered adequate to determine the azimuth of borehole installations for many years, but it had some serious shortcomings. The method was subject to errors due to mechanical assembly tolerances and was frequently plagued by nonlinear gyro drift. The major problem with the system was the fragile nature of the gyro probes themselves; they proved to be very susceptible to shipping damage and extremely expensive to repair. In addition, the manufacturer was not willing to warrant his expensive repair work in any way. Therefore, a much cheaper alternate method of orienting borehole seismometers has been developed and is currently replacing the gyro probe approach in programs with limited budgets.

The new method involves the installation of a horizontal reference seismometer on the surface near the borehole at a known orientation. The digitally recorded output of this surface sensor is then compared using the coherence and correlation functions with the digitally recorded outputs of the horizontal components of the sensor installed in the borehole to determine the relative azimuthal orientation of the borehole components with respect to the surface horizontal.

With the advent of sand installations, the horizontal components of newly installed seismometers are no longer being oriented in the conventional north-south east-west configuration. Instead, many borehole sensors are being installed at arbitrary azimuths with respect to north; the alignment of the horizontals with respect to north then becomes part of the data set. This approach has become feasible because modern computing power and digital data trivializes the task of rotating the data to any azimuth desired by the data user.

#### **7.4.5.7 Typical borehole parameters**

As the result of the SRO and IRIS programs, there are now many broadband borehole installations in use around the world. Most of these boreholes are geometrically quite similar because they were designed to accommodate the same seismic instruments. All of these boreholes are approximately 16.5 cm in diameter and most of them are drilled to a maximum of 3.5 degrees departure from true vertical. They are all cased with standard oil field grade casing and most of them are watertight.

There is some variation in the depths of these boreholes. As mentioned above, the vast majority of seismic boreholes are approximately 100 meters deep. However, some of these boreholes are considerably deeper if they were drilled in areas with thick overburden or poor bedrock. For instance, the borehole sensor at DWPF (Florida) is installed at 162 meters depth because the overburden at DWPF is approximately 46 meters thick and the upper layers of bedrock consist of interleaved units of varying grades of soft limestone. The borehole at

ANTO, which is drilled in competent rock for most of its depth, is the deepest and oldest IRIS borehole at 195 meters. This was the first field borehole that was drilled for the SRO program: as more experience was gained, it became apparent that boreholes that deep were not cost-effective. A few of the boreholes are shallower primarily because severe difficulties were encountered during the drilling operations that necessitated finishing the borehole at a shallower depth than originally desired. For example, the sensor at JOHN (Johnson Island) is at a depth of 39 meters because severe borehole collapses were encountered while attempting to drill deeper. The site is on a coral atoll and the surface layers are very poorly consolidated; true bedrock probably lies at very great depths. Drilling in volcanic regions often proves to be very difficult. The borehole at POHA on the island of Hawaii was terminated at 88 meters because the drillers experienced severe "loss of circulation" conditions throughout the drilling operation. The surface layers at POHA consist of badly fractured weathered basalt layers and basalt rubble separated by scoria rubble, ash flows, sand and other assorted debris produced by an active volcano. Drilling conditions in the volcanic deposits on Macquire Island proved to be so difficult that it was impossible to complete a borehole.

#### **7.4.5.8 Commercial sources of borehole instruments**

Currently, there are only two known commercial sources of high sensitivity broadband borehole seismometers. For many years, Teledyne Geotech in Dallas Texas, USA (now Geotech Instruments LLC; [www.geoinstr.com](http://www.geoinstr.com)) was the only source of high sensitivity instruments (KS-36000, KS-54000, GS-21, and 20171) designed specifically for borehole installation. Both the KS-36000 and the KS-54000 are three-component broadband, closed loop force feedback sensors that are designed for deep (up to 300 meters) borehole installation. The GS-21 is a conventionally designed short-period vertical deep borehole instrument intended for superior high frequency performance. The 20171 is a slightly noisier and slightly cheaper version of the GS-21. The KS-36000 is no longer manufactured but there are many of these instruments still in operation in boreholes around the world. Recently, Geotech has introduced a new sensor, the KS-2000, which is available both as a surface package and a 4-inch borehole package.

For the past few years, Guralp Systems Ltd. ([www.guralp.com](http://www.guralp.com)), Reading, UK, has been producing a borehole version of the CMG-3T (see DS 5.1; referred to by some as a CMG-3TB). This instrument is much smaller and much lighter than is a KS sensor; it is also considerably less expensive. This is a three component, broadband, closed loop, force feedback instrument that is very easy to install. Guralp Systems has recently introduced a new borehole sensor that has both a velocity and an acceleration output and is integrated with its own digitizer. In addition, they are willing to work with the customer to meet any specific requirements.

A borehole version of the STS2 has been under development for several years. Currently, a basic prototype of the instrument exists but the instrument requires further development of the remote control functions and the final packaging design is yet to be determined. Streckeisen has not announced an availability date for the new instrument.

It is somewhat hazardous to quote sensor prices because they are continuously subject to change by the manufacturer and international currency exchange rates change daily, but here are some approximate current relative costs for borehole sensors in 1999 US dollars. These prices should be viewed as being approximate; potential buyers should consult the manufacturer for a current quote.

A basic Geotech KS-54000 was priced at nearly US\$ 65,000. Additional costs will be about US\$ 40,000 for a conventional installation or about US\$ 13,000 if installed in sand and if all the associated installation hardware has been purchased. However, this price may be reduced if the instruments are ordered in sufficient quantities (25 or more). A GS-21 was priced at about US\$ 8,000 and the 20171 was around US\$ 6,000 for the instruments themselves. The associated hardware (soft electrical cable, wire rope, winch etc.) is additional. Estimated delivery time for these instruments is 120 days or more after receipt of order depending on the availability of non-Geotech manufactured parts. The soon to be introduced KS-2000 sensor will be priced at below US\$ 10,000 for the surface system and the borehole version will probably be below US\$ 20,000.

A Guralp Systems CMG-3TB costs about \$28,000 if the instrument is to be installed in sand; and about \$39,000 for a hole-lock equipped version. Delivery is currently about 9 months but they are trying to decrease this to about 6 months.

#### **7.4.5.9 Instrument noise**

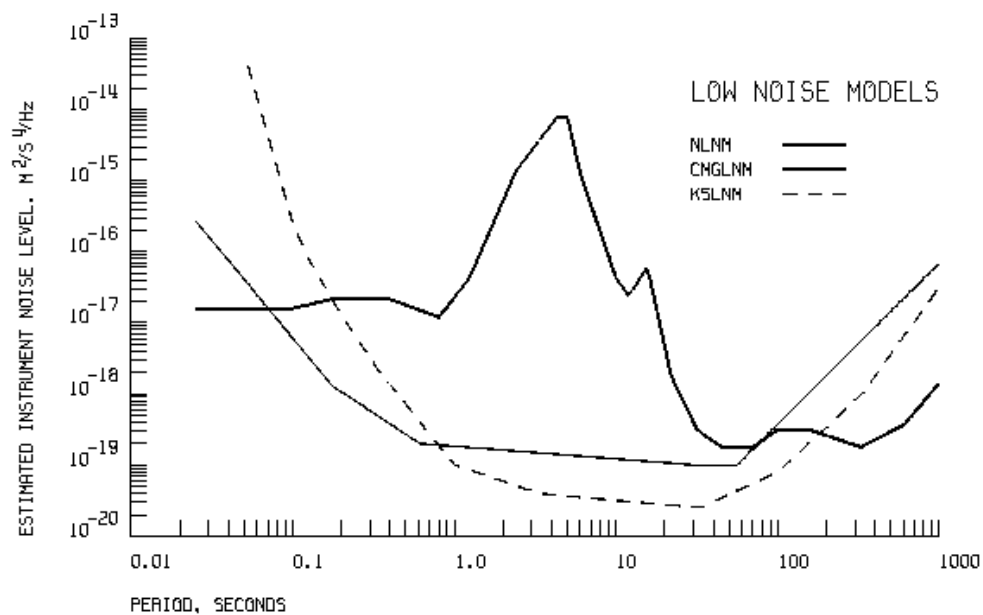
It is important to remember that the purpose of installing seismic instrumentation in boreholes is to obtain quiet seismic data. This will be foiled if the seismic sensor system itself is too noisy to resolve the lower levels of background noise of the Earth which are expected to be found at the bottom of the borehole. As delivered from the factory, sensor self-noise levels sometimes vary over a wide range and some instruments may be far too noisy to operate successfully in a quiet borehole. Therefore, it is recommended that the self-noise of all borehole instruments be measured before installation to ensure that they are quiet enough to be able to resolve the background noise levels anticipated at the bottom of the borehole. Self-noise measurements are usually made by installing two or more sensors physically close enough together to ensure that the ground motion input to all of the sensors is identical. The data produced by the sensors is then analyzed to determine the level of the incoherent power in each sensor's output; this incoherent power is usually interpreted as the sensor internal noise level (see 5.6). To achieve high fidelity recording of true ground motion, the seismometer system self-noise level should be well below the anticipated background Earth's motion levels across the band of interest at the site.

The low-noise models in Fig. 7.60 can serve as guidelines to the instrument noise levels that may be expected from the CMG-3TB and the KS-54000 sensor systems. In this figure, the CMG-3TB low-noise model (CMGLNM) is the thin solid line and the KS-54000 low noise model (KSLNM) is the thin dashed line. The solid heavy line in the figure is Peterson's (1993) new low-noise model (NLNM) for the background seismic noise at a quiet site. The reader should recognize that there is no single known site in the world whose background power spectral density levels reach NLNM levels across the entire band. Instead, the NLNM is a composite of the lowest Earth's noise levels obtained from many sites. Similarly, the low-noise models for the instruments should not be regarded as being typical of all instruments because each seismic sensor has a distinct personality of its own. Instead, the low-noise models for the instruments should be regarded as lower limits of instrument noise just as is the case for the NLNM of ambient Earth's noise. In all probability, individual instruments will be noisier than the low-noise model for that instrument over at least portions of the spectrum.

The CMGLNM plot in Fig. 7.60 is based on a composite of experimental test data obtained at the Albuquerque Seismological Laboratory over a period of several years. The central portion

(from about 0.6 to about 20 seconds) of the model was not actually measured because of numerical resolution limits of the data processing algorithms and this portion of the model is an estimate. As a general rule, many CMG-3TB instrument noise levels approach the CMGLNM at short periods (less than 0.6 seconds); fewer of these instruments achieve the indicated noise levels at long periods (greater than 20 seconds).

The KSLNM plot in Fig. 7.60 is a factory-derived theoretical instrument noise level. As such, it should be regarded as an optimistic estimate of the lower limits of the self-noise in the KS-54000. Most KS-54000 instruments are probably noisier than the levels indicated by the KSLNM curve.



**Fig. 7.60** Low-noise models for the KS-54000 (KSLNM) and the CMG-3TB (CMGLNM) sensor system self-noise relative to Peterson's (1993) new low-noise model (NLNM) for background Earth's motion.

#### 7.4.5.10 Organizations with known noteworthy borehole experience

As an organization, Teledyne Geotech (Geotech Instruments – [www.geoinstr.com](http://www.geoinstr.com)) certainly has the longest history in seismic borehole technology. However, personnel turnover in the past few years has significantly depleted Geotech's direct hands-on experience in boreholes. Another organization with a long history of borehole experience is the United States Air Force Technical Applications Center (AFTAC – [www.aftac.gov](http://www.aftac.gov)). Over the past 25 years they have deployed many KS instruments throughout the world. Most of these installations involve multiple sensor configurations deployed in arrays. Prior to the KS era, AFTAC used the older Geotech Triax instruments in boreholes at some of their arrays. As was the case with Teledyne Geotech, AFTAC does not have personnel with long-term borehole experience; US Air Force personnel tend to rotate in and out of their duty assignments every two years.

The Albuquerque Seismological Laboratory (<http://earthquake.usgs.gov/regional/asl/>) has been deploying KS sensors in boreholes since 1974 at sites located all over the world and recently has begun installing Guralp CMG-3TB sensors at some sites. The Laboratory has

borehole experience on all seven continents ranging from tropical jungle in Brazil to the permafrost of Antarctica. At ASL, the personnel situation has remained relatively stable and there are several personnel with many years of experience working with boreholes – some have been at it for over 25 years.

Southern Methodist University (Dr Eugene T. Herrin, e-mail: herrin@passion.isem.smu.edu) has been active in the borehole field off and on over the years. Recently they have been quite active in developing innovative economical methods for installing broadband borehole arrays. As an organization, Sandia National Laboratories ([www.sandia.gov](http://www.sandia.gov)) has considerable experience in borehole technology, most notably with their Remote Seismic Telemetered Network (RSTN) program. However, the lack of continuity in their seismic program has resulted in the loss of many of the personnel with real field experience in borehole technology.

During the past 10 years, the IDA group at the Scripps Institution of Oceanography at the University of California, San Diego ([www-ida.ucsd.edu/public/home.nof.html](http://www-ida.ucsd.edu/public/home.nof.html)) has become involved in land-based borehole seismology as a part of the IRIS GSN program. They now have experience in drilling boreholes and deploying instruments at several sites around the world.

In conjunction with personnel from the Woods Hole Oceanographic Institute, Scripps is also leading the US effort aimed at developing pioneering borehole seismology techniques for use on the ocean floor. Independent programs in ocean bottom borehole seismology are also currently conducted by groups in France and Japan. Installing seismic sensors in the deep ocean is developing rapidly and we will not attempt to summarize practices in this field.

#### **7.4.6 Borehole strong-motion array installation (R. L. Nigbor, 2002)**

##### **7.4.6.1 Introduction**

*"An important factor in understanding and estimating local soil effects on ground motions and soil-structure interaction effects on structural response is the three dimensional nature of earthquake waves. ...For these purposes it is necessary to have available records of the motion at various points on the ground surface, along two mutually orthogonal directions, as well as at different depths."*

These words, published in the proceedings of the 1981 U.S. National Workshop on Strong-Motion Earthquake Instrumentation in Santa Barbara, California, are echoed in every important meeting where policies and priorities have been set regarding strong-motion monitoring. Earthquake engineers and seismologists alike agree: borehole strong-motion data continue to be a priority for better understanding of site response and soil-structure interaction issues.

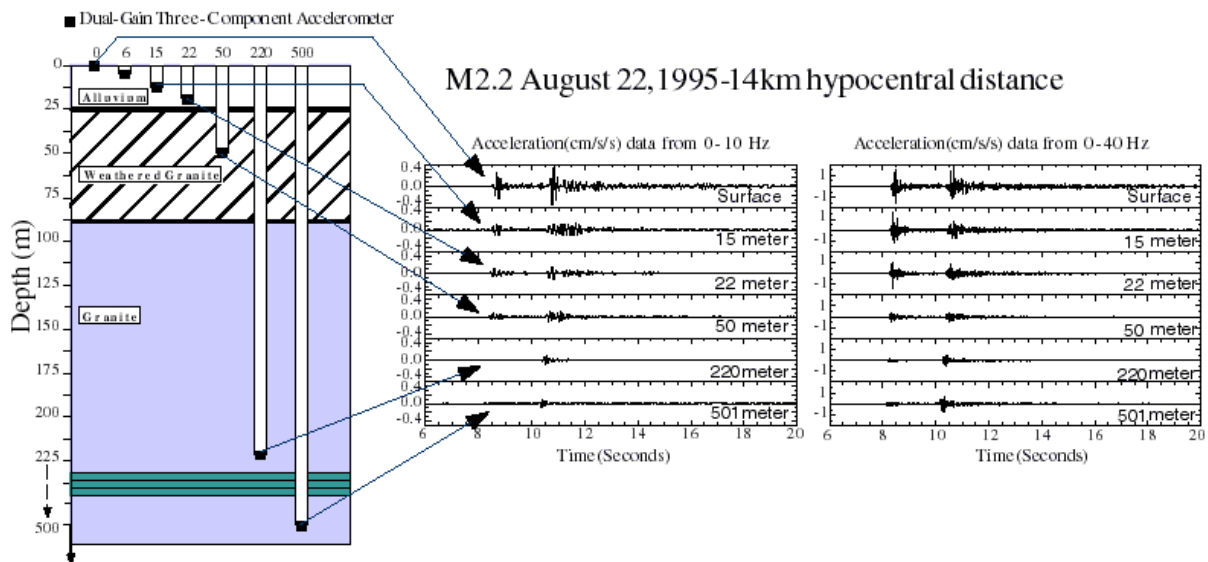
This section is somewhat of a departure from much of the New Manual of Seismological Observatory Practice, as borehole strong-motion observations are primarily focused on site response and not on the seismic source or wave propagation path. For engineering purposes, borehole data in shallow (< 100 m) soils are of primary importance; these data are used to study amplification of earthquake shaking in the soil layers. However, borehole data in rock, especially weathered rock in the upper 30 m, are also important for the understanding of

strong ground shaking in earthquakes. Rock sites often show larger variability in measured ground motions than do soil sites. Examples of well-documented strong-motion borehole arrays are the EuroSeisTest Project (Riepl et al., 1998, and Fig. 7.65) and the Garner Valley Downhole Array (see Fig. 7.61). The user of this Manual should consult these references for further information about the details of borehole arrays and the use of borehole strong-motion data.

As important as borehole data are, many practitioners experience difficulty designing and constructing such arrays. As with the more traditional seismological borehole systems (see 7.4.5), strong-motion borehole arrays present a variety of challenges. Fortunately, much has been learned about borehole strong-motion instrumentation and vertical strong-motion array construction. In the past, borehole systems rarely survived more than two years. However, today there are many successful, long-term three-dimensional strong-motion arrays throughout the world. This accomplishment can be traced to better design, to new instrumentation, to better understanding of the historical failures, and to improved installation procedures.

This section is intended to assist with planning and implementing a successful borehole strong-motion array. Details of the instrumentation are not directly discussed but are available from the manufacturers of borehole strong-motion systems such as Kinemetrics (<http://www.kmi.com>). The sections that follow discuss borehole array planning, borehole preparation, geotechnical/geophysical measurements, installation procedure, and costs.

Fig. 7.61 shows representative borehole array data from the Garner Valley Downhole Array, Fig. 7.62 is a sketch of a typical, simple borehole strong-motion installation and Fig. 7.63 shows an example of a borehole strong-motion array. These sketches are meant to show the various components and terminology that will be discussed in this section.



**Fig. 7.61** Sample borehole strong-motion array data from Garner Valley downhole array

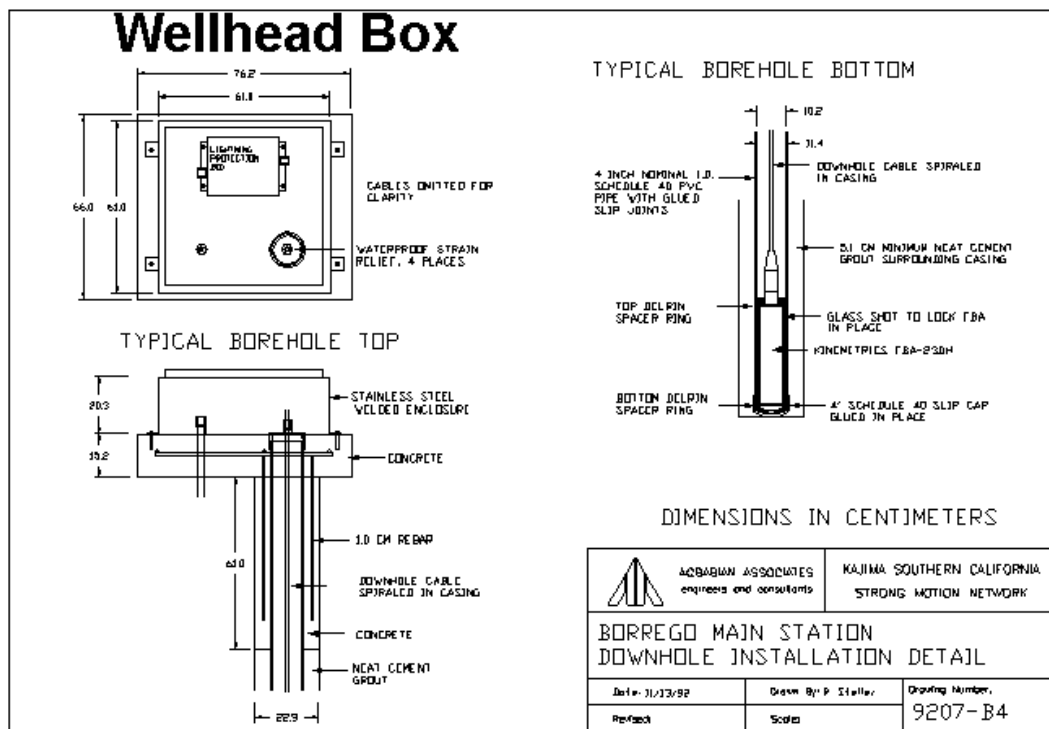


Fig.7.62 Sketch of borehole strong-motion accelerometer installation details.

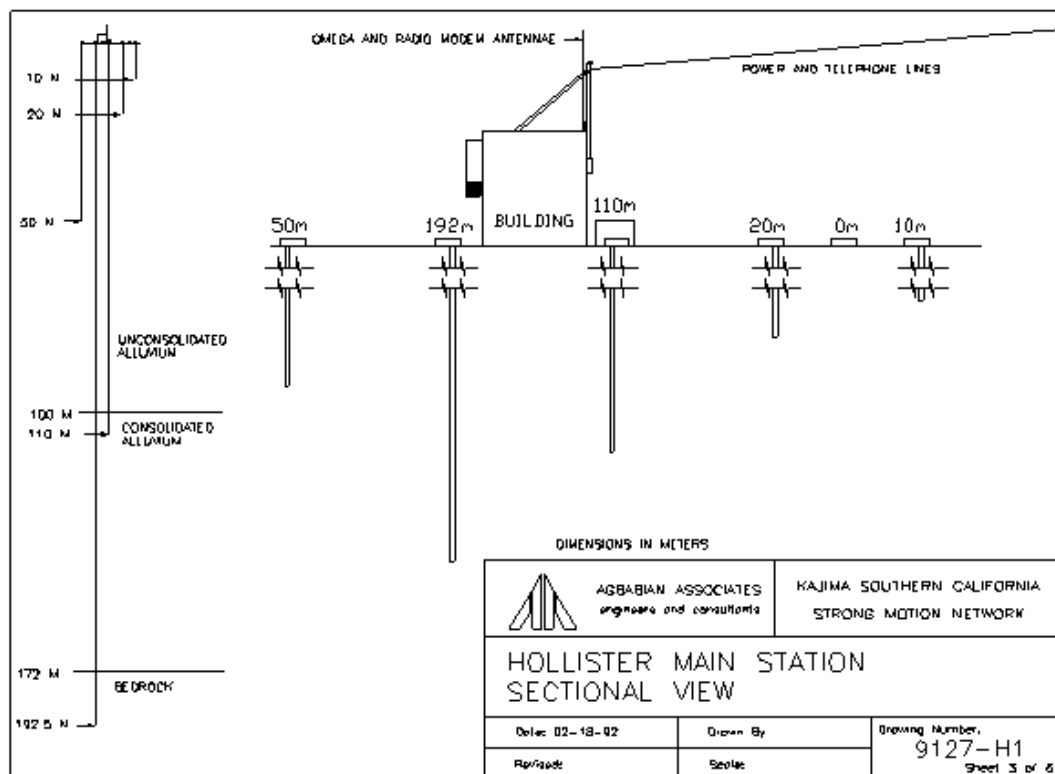


Fig. 7.63 Sketch of a borehole strong-motion array.

#### 7.4.6.2 Borehole array planning

This section focuses on the planning issues related to borehole strong-motion array installation. The most important step in implementing a successful borehole accelerometer system is good planning. Done properly, by the time the borehole accelerometer package is actually lowered into the hole (as in Fig. 7.64 below), 95% of the effort will be complete. The following are important considerations:

- location;
- geologic implications;
- coupling and retrievability issues;
- sensor orientation;
- system issues.

##### ***Location***

Borehole data are needed for source mechanism and wave propagation arrays, local site effects arrays, and as free field input to structural response arrays. The location of the borehole is principally dictated by the needs of the particular project and thus the required array configuration. Borehole location and depth will also depend on the soils and depth to bedrock. It is recommended that external advice or review be obtained for borehole location selection.



**Fig. 7.64** Lowering a borehole accelerometer into the cased borehole.



Ground-borne noise is not the serious issue that it is with high-gain seismic systems, but it is still important to minimize non-earthquake vibrations in a borehole strong-motion installation. Few man-made signals will penetrate tens of meters of soil, so background noise will be reduced in a borehole sensor. However, some boreholes are shallow, and often the borehole accelerometer is collocated with a surface sensor. For this reason, the borehole should be located as far away from cultural (man-made) noise sources as possible. These include large, above ground structures, such as telephone poles, which can be driven by wind, vibration sources such as nearby rock quarries or industrial plants, and roadways bearing large vehicles.

The structure used for housing the recording station itself can be a source of coupled soil-structure vibration and must be designed carefully. The interaction of large structures with the soils can introduce noise into the ground motion. For this reason, the surface accelerometers should be located at least  $1.5H$  distance away from the structure, where  $H$  is the height of the structure.

Within an array of borehole and surface sensors, one must optimize the layout with regard to physical concerns such as cabling and environmental protection. The lengths of surface cables should be minimized for several reasons. First, because of cost. Second, the longer the cable the greater the potential for damage or introduction of noise or induced voltage, even if the cable is shielded and in conduit, and even if there is lightning protection both at the wellhead

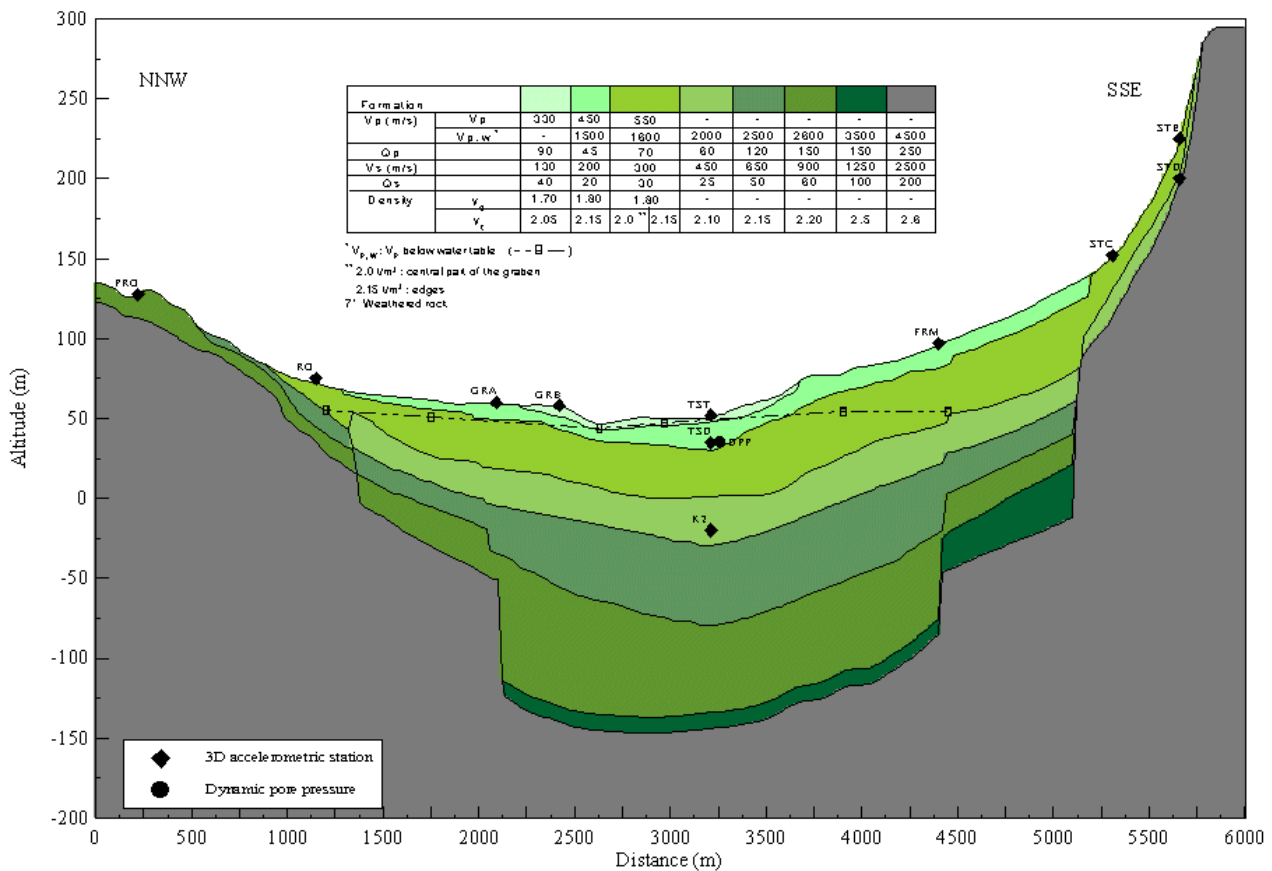
box and at the recording station (as there should be). The recording station should be located near the wellhead boxes to minimize cable lengths.

Finally, it is best if the wellhead box is dry most of the time although it is assumed that the borehole itself is full of water and the wellhead box is designed to be waterproof. The top of the borehole should be positioned with regard to local water drainage and preferably not in a topographic low.

### ***Geologic implications***

Specific knowledge of the geology of a site is extremely useful during planning in order to meet project needs and accurately estimate the costs. The implications of local geology will depend upon the specific purpose of the borehole array. One should at least understand the surface geology, the depth to basement rock, and the local and regional tectonic structure. Fig. 7.65 shows a composite model of the EuroSeisTest site. This is an example of the kind of geologic understanding which should accompany a borehole strong-motion array installation.

The best information will come from both a thorough literature search to find existing information and then pre-installation geophysical studies. Once a site is selected, more detailed geophysical and geotechnical studies will be needed for ground motion and structural response modeling.



**Fig. 7.65** Geologic model of EuroSeisTest showing accelerometer array configuration.

### ***Coupling and retrievability issues***

The coupling of the borehole sensor to the surrounding soil is a critical issue for borehole strong-motion systems. The goal of the measurement is to record the particle motion of the native soil or rock at depth. Care must be taken to ensure that the borehole installation minimizes the disturbance of the soil or rock column. The borehole itself, the casing, the grout used to seal the casing and couple it to the surrounding soils, the borehole accelerometer package, and the method used to couple the package to the wall of the casing, all can have some effect on the recorded motions, especially if the motions approach 1g. The issue of coupling is related to instrument retrievability, which is the ability to pull out a borehole sensor if repair is needed. For some borehole sensor installations, a permanent coupling solution (grouting or cementing the sensor in place ) may be selected. This is not recommended as experience has shown that borehole sensor failure does occur. Failure of a borehole sensor that can not be retrieved not only entails the replacement of the sensor, but of the borehole as well, and the cost of the borehole often well exceeds that of the instrumentation.

If permanent coupling of the sensor is essential, using some sort of grout, it is advisable to design the borehole system to have a “weak point” above the sensor that will break cleanly when the cable is pulled and leave as little of the cable as is possible in the hole. If the sensor fails, it would be possible to abandon the sensor and cement in a replacement in the same borehole at a slightly shallower depth.

Removable coupling (locking) methods include backfilling the annular space around the package with some specified material, wedge-type locking systems, and pneumatic/hydraulic locking systems. Backfill materials used in the past have included sand, gravel, lead shot, and glass beads. Of these, water saturated sands can be expected to liquefy under vibrating conditions, and lead shot has been found to cold form over time, making retrieval difficult and even impossible. This leaves gravel or glass beads as successful alternatives. Kinemetrics recommends the use of a combination of 3mm and 5mm glass spheres as a coupling method; the company can be contacted for further details.

Several commercial wedge-type locking systems are available from borehole sensor manufacturers. Experience has shown that these will work well in shallower (<50m) installations, but may become unreliable in deeper installations. Some borehole installations, such as the 500m borehole sensor installation at the Garner Valley Downhole Array have used custom hydraulic locking systems. Some temporary borehole strong-motion sensor installations have used pneumatic (air-pressure) locking systems with success, but the air pressure must be maintained.

### ***Sensor orientation***

Another important issue in borehole strong-motion studies is the accurate orientation of the horizontal components of the borehole sensors. In the past, practitioners have most often relied on loading poles to manually orient the instrument package. Loading poles are generally square-section tubes which are rigidly attached to the accelerometer package. During installation, the loading poles are joined end-to-end with the painted side facing the same direction as the others, thereby permitting the package to be manually oriented from the top of the hole. This manual method can work well for shallow (< 20 m) borehole installations, but twisting of the poles can introduce large errors for deeper installations.

Fluxgate magnetometer sensors were first used with strong-motion borehole instrumentation in the late 1980's by a firm called Applied Geophysics of Los Angeles. At that time, the compass was used to determine the arbitrary orientation of a slotted end cap installed at the bottom of the PVC casing by the driller. The mating notch on a special borehole package was then rotated and fixed so that, once guided (by gravity) into position in the slot of the cap, the package was oriented as desired. This method was expensive because of the special construction and installation of the cap at the bottom of the hole, and the special packaging required to accommodate the orientation notch.

The current generation of commercial downhole accelerometers (for example, Kinemetrics FBA-23DH) have provisions for an internal fluxgate magnetometer compass to make orientation simpler and much less expensive than either loading poles or borehole bottom devices. With this device, one simply observes the compass orientation via a notebook computer at the surface and rotates the sensor cable until the correct orientation is achieved. Accuracies of 2-3 degrees can be achieved with this method.

It is also possible to simply install a downhole accelerometer with random orientation and then to determine the orientation later by comparing the vector orientation of a surface sensor to that recorded from the borehole sensors. Orientation can be determined after installation by comparing surface and downhole data, either earthquake or microtremor. Note that anisotropy in the near surface soils can produce errors in this type of orientation unless the events are large regional events with significant long-period energy, i.e., with wavelengths much larger

than the soil depths. Using a known source location, the orientations can be determined to  $\pm 5^\circ$  by using linearity of 1<sup>st</sup> motions on radial and transverse components (see Aster and Shearer, 1991).

### ***System issues***

The borehole accelerometer package can not be installed properly without consideration of the overall system, and particularly the recording device. Usually, the recording system will have exceptional dynamic range to take advantage of the low noise qualities of the borehole installation and the accelerometer. Therefore, particular care must be taken avoid system-related noise due to improper grounding and other common problems. Additional issues include system level lightning protection to protect circuitry against large voltage transients. Many of these system issues have been discussed in detail in 7.4.2.1.

### **7.4.6.3 Borehole preparation**

A critical step in a successful borehole accelerometer installation is the preparation of the borehole itself. The borehole should be vertical, carefully drilled, with carefully installed casing grouted to ensure good coupling of ground motions at higher frequencies. Fig. 7.66 shows a typical drilling site. One can see that this can be a major construction effort.

### ***Planning***

Site selection for borehole accelerometers is often dictated by factors other than practicality for drilling operations. If there is room for adjustment of borehole location, try to meet or exceed the following minimum clearance requirements:

- 2m from borehole to any obstructions such as fences, walls, or ditches;
- two 10m paths 4m wide on opposite sides of the borehole for drilling equipment;
- no overhead power lines

Access to the site should be able to support repeated trips with heavy trucks without damage. If a site is located in soft soils, consider the use of wood under the wheels of the drill rig and a four-wheel-drive water truck. Some sites may require the use of track-mounted drill rigs for access, or heavy earthmoving equipment to help position the drill rig on site. Fig. 7.67 below demonstrates why these are requirements for a drilling operation. Important steps in borehole preparation are planning, contractor selection, permitting, drilling, sampling, casing, and grouting. These issues are discussed below.



**Fig. 7.66** Typical borehole drilling operation in an open area.



**Fig. 7.67** Drilling operation in a confined area, showing size of drill rig.

Water availability is critical for drilling operations. A residential-type supply is adequate as long as an on-site water tank or truck is available to provide storage for peak demands. If all water is to be supplied by truck, a water supply capable of filling the truck quickly should be located within a 15 minute drive. Disposal of drill tailings and excess water is usually not a problem in remote sites where the tailings can be spread around the surrounding area to dry. In urban areas, fluids can be channeled sometimes down drains or ditches, but solid tailings must be removed. For shallow holes this can be done in drums and a pickup truck, if a nearby spot can be located to dispose of the tailings. In urban areas, a separation cone can be placed over a waterproof dumpster, and tailings collected in the dumpster, to be removed periodically by a liquid waste hauler with a vacuum-lift truck. In some areas, drill tailings are automatically classified as hazardous waste, which complicates disposal matters tremendously. Be sure to work these issues out before starting the project. In many developed areas, this can be a large cost item, easily as large as the drilling costs.

If multiple boreholes are to be drilled in one location, a separation of 6m between boreholes is generally adequate to prevent damage to a borehole during the drilling of subsequent boreholes. This should preclude the possibility of boreholes drifting laterally into each other during drilling. Shallow holes may be placed closer together by drilling in a line and backing the drill rig over the most-recently completed hole to protect it from damage.

The time of year chosen to begin a drilling project may significantly affect the schedule and cost of a project. Warmer weather is generally desirable, as are long periods of daylight and a lack of rain or snow.

### ***Selection of drilling contractor***

The selection of a drilling contractor must be based upon a number of factors; perhaps the least important of these is cost. If geotechnical sampling is to be performed, the contractor must have the equipment and crews familiar with geotechnical drilling practices. The key item in drilling is experience with the tasks to be performed and the equipment to be used. Drilling deeper (>50m) boreholes is not the place to have an inexperienced crew. The driller's reputation for completion of work and quality of work should be reviewed. It is strongly recommended that one obtain references for a drilling contractor.

When obtaining a bid, consider getting separate bids for a fixed price per meter and for actual time and materials used. For deep boreholes, fixed price per meter bids may appear more expensive but can save enormous amounts of money if a site is difficult. In addition, most companies will send out their best crews on fixed price bids, generally giving faster completion and fewer complications. In the long run, fixed price contracts often save money.

### ***Permits***

The drilling of boreholes near aquifers is carefully controlled in many countries to ensure that ground water sources are not contaminated. This control is generally exercised by local government, often through environmental health departments. Usually the responsible entity will require the submission of a permit application detailing depth, diameter, location, property lines, adjacent structures, wells and septic systems; well construction method, owner, and licensed driller to perform the work, and a fee of several hundred dollars per well. The application may require the signature of the land owner and the drilling contractor. In addition, some regions require copies of the driller's license and a performance bond before the driller is approved for work. In the US, typical permitting work is as follows:

- submission of permit application and fee;
- after permit application and payment of fee an inspector will usually visit the site before issuing approval of the permit;
- during construction, depending upon location, the inspector may visit the site several times, and may require notification 24 hours before grout is to be placed;
- in some municipalities, an inspector must be present during the grouting process;
- notifying the permitting agency of the completion of wells; usually this is in the form of a drill log which shows lithologic information and details of the well construction.

In other countries, and in rural or remote areas, permitting may not be needed. However, permission by the owner of the land will likely be required in all cases.

### ***Drilling***

There are several methods for shallow drilling in soils and rock. The book by Driscoll (1986) on Groundwater and Wells is a good reference for the various methods. Several ASTM (American Society for Testing and Materials) Standards describe drilling methods as well (see <http://www.astm.org>). A good start is ASTM D420-98 "Guide to Site Characterization for Engineering, Design, and Construction Purposes."

Direct rotary or "rotary mud" drilling is the drilling methodology best suited to most downhole accelerometer installation projects. One major advantage of this method is the support that the borehole wall receives from the drilling fluid that always fills the borehole. This method can be performed in both hard and soft formations, and can be done using fairly compact equipment. The drilling is performed by rotating a bit on the end of a heavy pipe or "drill string", which is driven down by its own weight, or by additional downward forces applied by hydraulic cylinders or chain pull-downs on the drill rig. The bit is lubricated, and drill cuttings are removed by drilling fluid or "mud" that is pumped down to the bit through the drill string. The fluid then returns to the surface through the annulus formed by the outer surface of the drill string and the borehole wall. This is possible because the borehole diameter is generally several inches larger in diameter than the drill string. As the fluid moves to the surface it carries with it the particulate debris produced by the cutting action of the bit. At the surface the fluid is directed into a holding area, either a box or dug pit, to let the cuttings settle out before the fluid is pumped down the drill string again.

Speed of completion is important in drilling due to the potential loss of a borehole by collapse. Some soil formations will remain open for long periods of time even if the fluid level drops significantly; other formations will cave in with the slightest provocation, even when filled with thick drilling fluid. In many instances, the premium paid to work around the clock is justified by the savings of not having to remove the drill string each evening, and the reduced risk of borehole collapse during the night.

Drilling is a messy business due to the large volume of water and mud involved. In a confined urban space such as an alley, a 100 m borehole might be drilled by a truck mounted rig only 8 m long, with an 8m tall tower, with only 1m of clearance on either side of the rig. Water and drill rod would be transported by a single 6m truck. All cuttings could be contained and removed from the site in 200 liter drums or a larger container. This would be possible but is far from an ideal situation. In a rural or remote area, a 250 m borehole might use a rig 15 m long, with a tower of 10-15 m, a separate pumping rig 10 m long, a 12 m drill string trailer, a

water storage tank, a mud pit 3 by 6 m, and an area several hundred m<sup>2</sup> to hold cuttings piles and miscellaneous support equipment. Fig. 7.67 shows such a setup.

Installation of a typical 100 mm (4 inch Schedule 40 or 80) PVC casing requires a minimum 200 mm (8 inch) diameter borehole. A larger diameter may be used when needed to maintain a clear hole, but this will increase drilling and grout costs and the potential of damage to the casing from grout cure heat, discussed further in the section on grouting. Most municipalities in the USA require a 50 mm (2 inch) annular seal around the casing, and this diameter meets this requirement. Depending upon the size of the drill rig used, this may be drilled in one pass, or a smaller pilot hole; usually 120 mm (4 7/8 inch) diameter will be drilled first, and all sampling and geophysical testing will be done in the pilot hole before drilling again to the final size.

There is a trade-off between speed of drilling and verticality of the borehole. Since verticality is an important issue for downhole sensor installations, it is important to make the driller aware of this, and stress that slow steady drill rates and perhaps collars attached to the drill string can help keep the borehole vertical. If you can get the driller to agree to a tolerance when doing a contract, this will also help keep the drill rig operator focused on this issue. Deviations of less than 5° from the vertical are acceptable. Larger deviations will affect the dynamic range of most sensors, unless the instrument has some type of auto-leveling device.

### ***Geotechnical sampling***

Generally, installation of a downhole accelerometer, or vertical array, is done to understand the effects of local site response on ground motion. Acquisition of soil or rock formation samples during drilling, as well as geophysical information, provide a great deal of information useful in site characterization studies. Not only can these samples provide clear indications of the formations beneath a site, but can also be used in laboratory studies to determine structural characteristics of the formation.

Soil sampling is described by Kenji Ishihara (1996) and by Driscoll (1986). Several ASTM standards also exist for sampling; two important references are ASTM D1586-99 “Standard Test Method for Penetration Test and Split-Barrel Sampling of Soils” and ASTM D1587-94 “Standard Practice for Thin-Walled Tube Geotechnical Sampling of Soils”, both from <http://www.astm.org>.

In general, one will use five major categories of sample types in borehole strong-motion array studies, as discussed below.

- **Bag samples** are simply a collection of drill cuttings carried to the surface by the drilling fluid and caught as they enter the mud box or pit. The accuracy of this sampling method is influenced by a number of factors including depth of the borehole, rate of circulation of drilling fluid, and size of cutting fragments produced by the drilling process. This method has limitations, due to the introduction of clays from the drilling fluid into the sample, as well as older material falling off the borehole wall. However, when used in conjunction with a detailed record of drilling rate, bag samples can provide extensive information about formation.
- **Drive sampling** (“Split-Barrel Sampling” or “SPT” sampling) produces an intact but disturbed sample 2.5-5 cm (1-2 inch) in diameter by driving a sample tube into the formation at the bottom of the borehole. This is done by removing the drill string and



lowering the sample tube, mounted on the bottom of a sliding hammer, to the bottom on a cable. The hammer is then actuated by lifting and dropping the cable until the sample tube has been advanced the desired distance. Often the number of blows to advance a given distance is recorded to provide blow count (SPT N-value), a measure of formation hardness. This procedure is time consuming because it requires the removal of the entire drill string from the borehole, but provides samples with excellent depth control. This method is useful only in soils. Fig. 7.68 shows typical drive samples.



**Fig. 7.68** Drive samples being collected for later laboratory testing.

**Pitcher samples** (“thin-walled tube” samples) produce an undisturbed sample 75 mm (3 inches) in diameter and up to 750 mm (30 inches) long. This technique is used when large high quality samples are required for laboratory tests, and where formations are too hard to yield results with a drive sampler. Pitcher sampling is performed by removing the drill string from the borehole and replacing the standard bit with a pitcher sample barrel. The barrel supports a thin wall steel tube on a spring loaded plunger. The plunger allows the tube to retract through the center of an annular carbide bit when it reaches the bottom of the borehole. Drilling then proceeds, cutting an annulus. The 75 mm (3 inch) center core is forced into the thin wall tube. When advancement is complete, the entire assembly is removed and the core shears off at the bottom of the thin wall core tube. The sample may then be stored and transported in the tube, or extruded at the drill site. The original bit is then replaced and lowered to the bottom of the borehole to resume drilling. This is a very time consuming procedure, as it requires removing and inserting the entire drill string twice, but it yields very good undisturbed soil samples. This method is not used in very stiff soils or rock. Fig. 7.69 shows some Pitcher samples in the field.



**Fig. 7.69** Pitcher samples.

- **Diamond core samples** are taken in hard formations, usually rock. The procedure for diamond coring is identical to Pitcher sampling, except that a diamond core barrel is used. It also cuts an annulus while retaining the center core inside the core barrel. As with the Pitcher sample, this is a very time consuming procedure, as it requires removing and inserting the entire drill string twice, but is the only way to obtain samples in hard rock. Continuous coring provides a complete undisturbed record of the formations the borehole passes through. There are several ways of performing continuous coring; one method is referred to as “wireline” sampling. This, as well as most continuous methodologies, allow for the mounting of a variety of annular bits at the bottom of the drill string as well as for the exchange of the center portion of the bit from cutter to sample tube via a lightweight wireline cable through the center of the drill string. This permits the retrieval of cores without removing the drill string. This is a superior methodology for deep (>200m) boreholes where the larger drill rigs required to support it are justified.

### ***Casing***

A plastic (PVC, poly-vinyl-chloride) cased borehole is recommended for installations to 250 m depth. Medium-walled casing (“Schedule 40” in US standards) is acceptable to about 50 m depth; thick-walled (“Schedule 80”) should be used for installations between 50 and 250 m. Steel casing is recommended for deeper installations. As discussed elsewhere in this Manual, the Kinometrics FBA-23DH and its associated installation equipment are designed for use with Schedule 40 PVC casing. The bottom of the casing is closed with a slip cap. In the United States, this casing is typically supplied in 20 foot lengths with “bell-end” sockets molded at one end to receive the straight “spigot” end of the next casing section. Other sizes and forms of casing may be used but will require modification of associated installation items. Magnetic casing materials must not be used in conjunction with the flux-gate magnetometer compass option.

Joining of PVC casing sections is done using a solvent glue. A primer is used to clean and etch the surfaces to be joined before the glue is applied. Low temperatures can significantly degrade the quality of a solvent joint, as can the presence of water on the joint surfaces. If temperatures are below 0 degrees C, a low temperature solvent glue should be used. More expensive screw-joint casing can be used as an alternative.

Installation of the casing, except in the shallowest holes, requires filling the casing with water, sometimes drilling fluid, to negate the buoyancy of the casing column thus allowing the casing to be pushed down into the fluid filled borehole, usually by hand. In addition, fluid inside the casing equalizes internal and external pressures, preventing the collapse of the casing due to external fluid pressure in deep boreholes (greater than 100 m).

Attempting to push empty casing into a fluid filled borehole with the weight of the drill rig causes casing to "snake" in the borehole, making accelerometer installation more difficult, as well as increasing the risk of damage to the top of the casing section being forced down or by telescoping an uncured glued joint.

Joining PVC casing by the use of screws in conjunction with gluing is not recommended, due to the potential for protrusion through the interior wall and subsequent damage to the cable during installation, as well as providing a path for leakage of water out of the casing following installation. If screws are used, only stainless steel screws set partially through the casing should be used. Pilot holes should be drilled in the casing after gluing to prevent fracturing of the casing during screw emplacement.

### ***Grouting***

Grouting the well casing involves filling the annular space between the casing and the borehole wall with a suitable slurry of cement or clay. For borehole accelerometer installations it is critical that this process is done with care, to ensure that the casing is well-coupled to the native soil or rock.

The grout is pumped into place through a small diameter pipe, usually a 25mm (1 inch) galvanized steel called a "tremie tube", lowered into the borehole between the casing and the borehole wall. When the end of the pipe reaches the bottom of the borehole, drilling fluid is circulated through the tremie tube to establish a clear path for the grout, and to clean the bottom of the borehole of any settled material. The grout is then pumped down to the bottom of the borehole, where it displaces drilling fluid out of the top of the borehole. The pumping may be done by the pump on the drill rig, or by a separate grout or concrete pump. When the drilling fluid is completely displaced and grout is flowing from the top of the borehole, the pump is stopped and the tremie tube withdrawn, disassembled and cleaned. Many U.S. municipalities require that the volume of grout placed be recorded and that it meet or exceed the volume calculated for the annulus. In some deep boreholes, grout will be placed in several separate loads or "lifts" to reduce the pressure exerted on the pipe by the liquid grout, as discussed later. This is usually scheduled as one lift per day.

Local codes usually require a sanitary seal of cement grout in the top 15m of a well, and seals between all aquifers penetrated by the well. The other areas may be filled with sand or gravel, but it is generally cheaper just to fill the entire annular space with cement grout. This will also make the eventual abandonment of the well (which can require another permit) much simpler.

Common grouting practices primarily center on the use of cement and water ("neat cement"), although the slurry may also contain sand, bentonite clay, or hydrated lime in certain

situations. For downhole installations a mix of 400kg (20 U.S. sacks) of Portland cement type A or B per cubic meter of grout works well. Addition of 5kg of bentonite clay per cubic yard will ease the pumping of the grout into deep boreholes.

It is important to recognize that cement grouts exert greater collapse pressure on casing than water or drilling fluid. Installing grout 60m at a time for Schedule 40 PVC pipe provides a safety margin against casing collapse for the added effect of softening of the pipe by the heat of cure of the grout.

Other methods of placing grout, for example through a one-way valve in the bottom of the casing, are sometimes used but are generally considered to be less reliable. Be sure that the drilling contractor is completely comfortable with whatever method is to be used.

#### **7.4.6.4 Geotechnical/Geophysical measurements**

The primary motivation for installing downhole accelerometers is to increase understanding of the contribution of site response to the earthquake ground motion. Often the measurements of site response will be accompanied by analytical studies. Detailed understanding of the subsurface geology and soil/rock properties is necessary for such analytical studies. A good example of the kinds of site characterization data needed for strong-motion site response studies can be found in the documentation of the project

“Resolution of Site Response Issues in the Northridge Earthquake – ROSRINE” (<http://peer.berkeley.edu/news/2001spring/rosrine.html>).

The basic site geology provides the primary description of the site. Information obtained during drilling (through observation and soil sample collection) will augment any prior geological knowledge of the general area. Normal laboratory testing of soil samples (disturbed or undisturbed) will confirm soil/rock types. Borehole and surface geophysical measurements can also assist in determination of site geology.

In addition to the site geology, dynamic soil and rock properties are needed for modeling of earthquake site response. The primary modeling properties are density, dynamic modulus, and damping (Q-value). The latter two properties are nonlinear functions of strain. These properties are obtained by laboratory testing of undisturbed samples and by one or more surface or borehole geophysical measurements of shear-wave velocity.

This Chapter gives a brief overview of the most common geological and geophysical measurements used in conjunction with borehole accelerometer installations to determine site geology and dynamic soil/rock properties.

#### ***Literature search***

In most populated areas there will have been previous geological studies of the region and perhaps even local environmental, ground water, or planning studies. These can contain a wealth of information that will assist in site response studies. Planning for a borehole accelerometer installation should include a thorough literature search for such previous studies.

Potential sources of information on a regional basis are government geological or natural resources agencies. An example is the U.S. Geological Survey. For local studies, sources of

information are the local government planning agency, local universities, private water companies, and even local water well drilling companies. A literature search can be a very inexpensive source of information on site geology and even subsurface soil and rock properties.

### ***Pre-installation geophysical studies***

Before site selection and borehole drilling, geophysical methods can be used to obtain a more detailed understanding of the site geology and subsurface properties. A good review of methods for site characterization is found in ASTM Standard D420-98 “Guide to Site Characterization for Engineering, Design, and Construction Purposes” and in ASTM Standard D6429-99 “Standard Guide for Selecting Surface Geophysical Methods” (see <http://www.astm.org>). For borehole strong-motion array applications, common methods are:

- seismic reflection;
- seismic refraction;
- resistivity profiling;
- cone penetrometer.

Seismic reflection and refraction are two techniques for using surface measurements to determine the seismic wave velocity structure of the subsurface geology. Both use a surface source of energy (mechanical or explosive) and instrumentation for measuring travel times of seismic waves at various distances from the source. Inverse analysis of these travel times provides an estimate of the seismic wave velocities of soil and rock layers. These methods can provide a cost-effective determination of general soil/rock layer properties, bedrock depth, and water table depth over a wide area.

Resistivity profiling is another surface technique for measuring the electrical properties of the subsurface geology. The electrical field from a surface AC or DC current source is measured at several locations. Inverse analysis is then used to estimate the resistivity of the subsurface soil or rock. This method can assist in both shallow (< 100 m) or deep (> 100 m) geological studies of a site.

The previous methods have all been noninvasive surface techniques. Initial geological studies of a potential borehole accelerometer site can also include the invasive techniques of cone penetrometer studies. These allow detailed soil/rock type determination at a specific location. A cone penetrometer (a metal probe pushed into the soil) can also obtain information about shallow (< 30 m) soils. Exploratory drilling can also be used in these initial site characterization studies.

### ***Lithology logging***

An experienced geologist should be present during drilling to determine the soil and rock classification. This is done by observing the drill cuttings, the samples, and the action of the drill rig. A good procedure for such lithology logging exists in the ASTM Standard D5434-97 “Standard Guide for Field Logging of Subsurface Explorations of Soil and Rock” (see <http://www.astm.org>).

### ***Laboratory testing of soil samples***

Samples obtained during drilling are useful in determining the soil type and soil properties. Basic geotechnical soil properties can be determined by simple laboratory testing, including:

- moisture content;
- dry density;
- LL/PL; and
- void ratio (porosity).

These simple soil measurements can be performed by most commercial or university soil laboratories.

Dynamic laboratory testing, however, requires a much more skilled and specialized laboratory and very high quality undisturbed samples. Dynamic properties of primary interest for earthquake site response analysis are soil shear modulus and material damping ratio (in shear) and their variations with:

- shear strain;
- effective confining pressure;
- loading frequency;
- loading duration; and
- number of load cycles.

Dynamic testing should be performed using triaxial resonant column, simple shear, or torsional shear methods. Appropriate strain ranges for earthquake response studies are 0.0001% to 0.1%. Appropriate frequency ranges are <1Hz to 200Hz. Further details of dynamic testing can be found in the book *Soil Behaviour in Earthquake Geotechnics* by Ishihara (1996) and other textbooks on soil dynamics.

### ***Borehole geophysical measurements***

There are many geophysical measurements available for characterization of the soil and rock properties in a borehole. These can measure chemical, electrical, radiation, and mechanical properties. All require specialized instrumentation and a skilled, experienced field geophysicist.

The chemical, electrical, and radiation properties are generally not of interest in an earthquake site response study, except as they are useful in determining soil and rock types. Sometimes resistivity and natural gamma emission measurements in an uncased borehole (before installing PVC casing) can be useful in determining boundaries of clay, sand, and rock layers

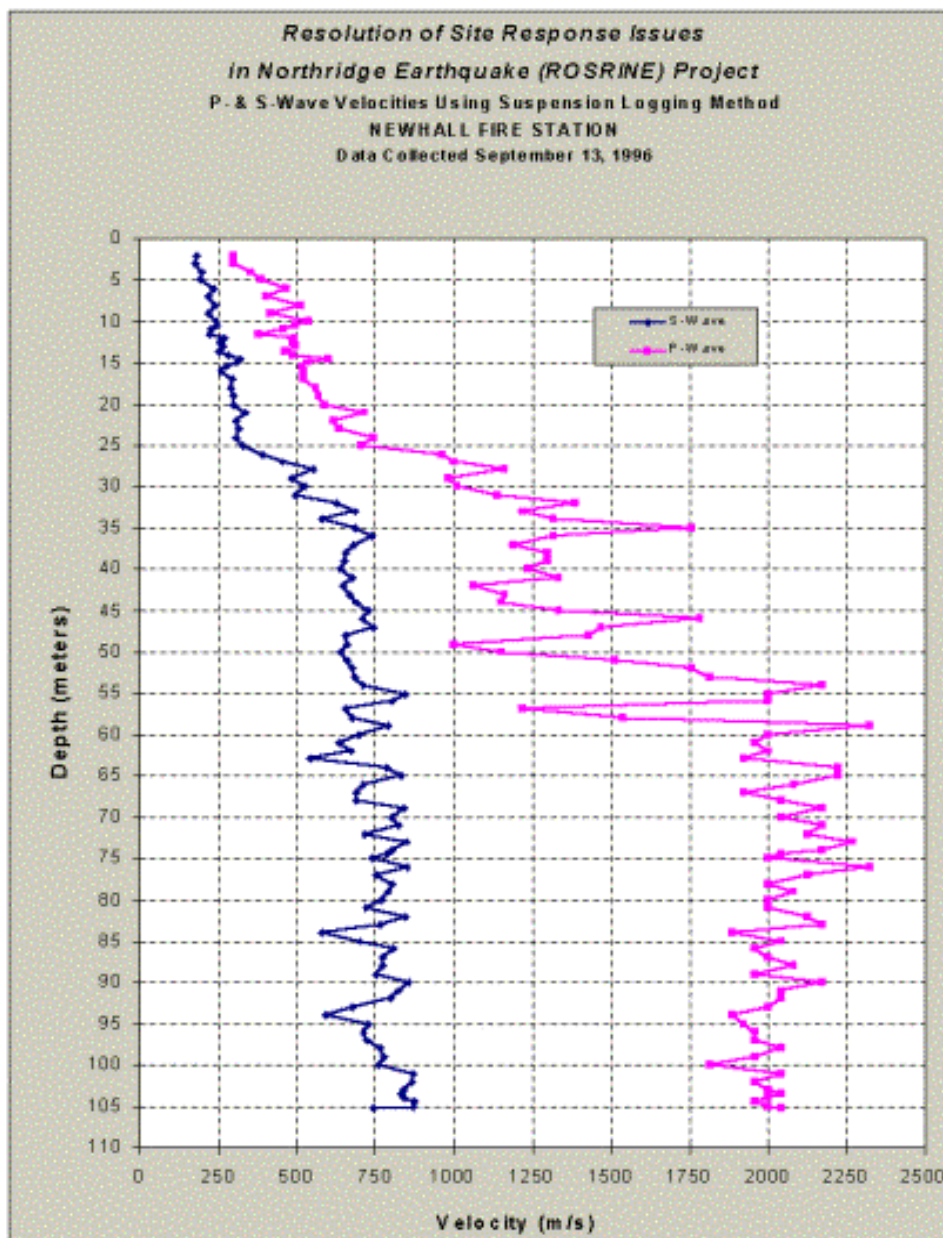
Of particular interest to site response studies are the mechanical properties of the soil and rock, primarily the P-wave and S-wave velocities versus depth (velocity profile). Borehole methods for velocity profile measurement are:

- downhole (vertical seismic profiling);
- crosshole; and
- suspension.



All three methods use a mechanical or electromechanical source to produce seismic waves, and one or more sensors (generally geophones) and a recording system to measure the induced ground motion. Details of these methods can be found in the book by Kenji Ishihara (1996).

For the downhole test, an impulsive energy source at the surface near the borehole top produces seismic waves which propagate radially. These can be either P waves or SH waves, depending upon source configuration. One or more sensors are installed in the borehole at known depths. The source and sensor signals are recorded, and the travel time of the first wave arrival is measured as a function of depth. The instantaneous slope of the travel time vs. depth curve is the reciprocal of the wave velocity at that depth.



**Fig.7.70** Velocity profiles for P waves (red curve) and S waves (blue curve) at a strong-motion site.

The crosshole test requires two or three adjacent boreholes separated by known distances. A source is installed in one borehole and receivers in the others, all at the same depth. The travel time of the generated seismic waves between boreholes is measured, and the velocity calculated by dividing the separation distance (at depth) by the travel time. This method requires careful control over the source and receiver depths, and detailed measurement of borehole separation versus depth. Both the downhole and crosshole methods are normally performed in cased boreholes. The suspension method, however, can be used in either cased or uncased boreholes. It consists of a single probe about 5 meters long, with an impulsive source at the bottom and two sensor sets ("receivers") near the top, separated by 1 meter. The source transmits energy through the borehole fluid to the borehole wall where it is transformed into both P and S waves. These are detected by both receivers, and the difference in arrival times is measured. Dividing the 1 meter separation by the travel time differences gives the P- and S-wave velocities for the 1 meter interval between sensors. Fig. 7.70 is a plot of a velocity profile measured with the suspension method.

#### **7.4.6.5 Installation procedure**

Depending on the goals and the operational environment of strong-motion installations there exist different guidelines for site selection, instrumentation and installation, e.g., those published by the Consortium of Organizations for Strong Motion Observations Systems (COSMOS) for urban or advanced national seismic strong-motion reference stations. These guidelines are available via <http://www.cosmos-eq.org/publications>. This section describes the careful installation of a downhole strong-motion sensor, its orientation, operational checkout, evaluation period, coupling/locking, and documentation/reporting. Accompanying this procedure will be a procedure for the installation of surface sensors, surface cabling, recording station, and other infrastructure, which have been discussed in more detail in earlier sections of 7.4.

##### ***Sensor installation***

After completing functional tests of the sensor, the wellhead box, the cables and recorders, and calibrating the internal compass, the borehole package can be installed. The following procedure assumes a standard installation of a Kinemetrics FBA-23DH sensor with glass beads or gravel, and the internal compass for orientation. Installation for other downhole strong-motion sensors will be similar.

If not already done, the borehole should be filled with clean water to within 6m of the top. Filling completely will make installation easier since the water gives the cable near- neutral buoyancy, allowing a single person to handle the weight. In any case, consideration must be given as to where the displaced excess water will go when the package and cable are lowered into the hole.

This is the main reason a drain should be provided for the wellhead box, even though it is sealed from weather, and also why sufficient slack surface cable should be provided in the wellhead box to allow the contents of the wellhead box to be temporarily moved out of the way. In a sealed system, it would be best if the drain were sealed off after performing this function, to prevent moisture from entering the wellhead box after installation is complete. If the borehole has already been checked, the package can be lowered smoothly using the cable, being careful not to allow the cable to slip away. Fig. 7.64 shows this procedure in action. It is good practice to have a second person feeding the cable to the person lowering the package.



If the borehole has not been checked previously, care should be taken to "feel" for any constrictions in the casing which could signal trapping of the package. If such is felt, it is wise to move slowly, and then try coming back up every meter or so until the obstruction is passed. Check the total depth of installation by using the depth markers provided on the borehole cable (if available). Continue until the sensor is resting on the bottom of the borehole.

### ***Orientation***

This procedure assumes an internal compass in the sensor package. Once the sensor package has reached the bottom of the casing, apply power to the compass and rotate the package using the cable until the desired orientation is reached. Feed slack cable into the casing to help hold the sensor in position. Allow the package to rest on the bottom to get a steady reading, and measure the accelerometer offsets. Record the offsets carefully.

The acceptable DC offset depends on the final gains expected to be used with the system. If possible, it is desirable to keep the offsets to less than 25 millivolts. To accomplish this, compensating offsets will usually need to be applied to counteract the combination of residual factory offsets and vertical misalignment of the casing at the bottom of the borehole. In other words, if an offset of +150 millivolts is recorded for the horizontal sensor oriented east at the bottom of the hole, then when the package is removed to the top, the offset for the same sensor must be mechanically adjusted so that -150 millivolts is added to whatever the offset is before the adjustment. This must be done carefully so that the package orientation doesn't change from the beginning of the adjustment to the end. This is where some sort of test fixture is helpful to hold the package steady. This procedure may need to be repeated in order to get it right. In fact, expect to repeat this procedure once more after 30 or 60 days of operation before installing the backfill material, after the sensor has adjusted completely to the temperature at the bottom of the borehole. This requires opening the sensor package which can be difficult in field situations. Requiring the driller to produce a vertical borehole within a predetermined tolerance can often avoid this.

### ***Operational checkout***

Once the package has been installed and oriented with acceptable DC sensor offsets, one should connect the recording system and check for proper sensor operation. This operational checkout should follow manufacturer's procedures, and results of the system test should be compared with the factory reference test.

### ***Evaluation period***

It is recommended that the array be operated for at least 30 days and as much as 60 days prior to installation of the backfill material. This period is needed to eliminate any initial startup problems, and allow the sensors to achieve steady state temperature response. During this period, the sensor package will rest at the bottom of the borehole. While this is inadequate coupling for large earthquakes, it should prove adequate for small events.

Preamplifier gains in the recording system should be set 10 to 100 times higher than normal in order to record as many small events as possible during this initial test period. Data should be reviewed, and the sensor removed and checked if any problems or questions arise.

### ***Coupling/Locking***

Once the downhole accelerometer has functioned properly for a period of 30 to 60 days, it is time for installation of the coupling or locking device. If a permanent locking method (i.e., grout or cement) is to be used, one should carefully install the locking material without disturbing the sensor orientation. If a wedge-type, pneumatic, or hydraulic locking system is used, it should be tightened to final specifications. If sand, gravel, or glass bead backfill is used, proceed with backfilling. Resist the temptation to pull on the cable afterwards to confirm the security of the system. It is possible to shift the package enough to disturb the orientation, and not be able to get it right without pulling it out, or you could lose backfill material to the bottom of the borehole. It is better to trust the process, and assume the material is in position. The proof is in the results.

### ***Documentation/Reporting***

Excellent documentation is very important to preserve the details of construction and installation for better interpretation of the data. Most important is the proper organization and use of calibration data. It is suggested that a formal Commissioning Report be created to preserve this installation information. Photographic documentation is also important. All documentation should be preserved along with the data for use in future data analyses.

### **7.4.6.6 Costs**

A borehole strong-motion array can be quite expensive when all the needed work is considered. Besides the cost of instrumentation, there are the planning, preparation, site studies and installation costs. One could omit some of the planning and site characterization costs (this is often done), but at a significant cost in understanding of the resulting strong-motion data.

Costs will vary considerably in different parts of the world. With this proviso, below is a commercial cost estimate for a representative borehole strong-motion array with one borehole at 100 m depth and one surface sensor. The year 2000 costs in US\$ in the U.S./California are:

- Instrumentation (one borehole accelerometer, one surface accelerometer, one 6-channel seismograph) - \$20,000;
- Downhole sensor cable - \$4,000;
- Array infrastructure (power, enclosures, communications, etc) - \$10,000 ;
- Planning and initial studies (includes one seismic refraction line) - \$10,000;
- Borehole preparation - \$15,000;
- Geotechnical/Geophysical Studies (includes lab testing of soils) - \$15,000;
- Array construction and installation - \$10,000;

**TOTAL COST = \$84,000.**

### **7.4.6.7 Special references and websites related to strong-motion installations, measurements, data analysis and use**

Reports, Guidelines and Workshop Summaries of the Consortium of Strong Motion Observation Systems (COSMOS), available under <http://www.cosmos-eq.org>, e.g.:

Anonymous (2006). Workshop 2: Guidelines for installation, operation, and data archiving

and dissemination.

Anonymous (2005a). Guidelines and recommendations for strong-motion record processing and commentary. COSMOS Publication CP-2005/1, 36 pp.

Anonymous (2005b). Effects of strong-motion processing procedures on time histories, elastic, and inelastic spectra. COSMOS Publication CP-2005/2, 46 pp.

Anonymous (2004). Invited Workshop on record processing guidelines.

Anonymous (2001a). Guidelines for installation of advanced national seismic system strong-motion reference stations, 45 pp.

Anonymous (2001b). Urban strong-motion reference station guidelines – goals, criteria and specifications for urban strong-motion reference stations. 5 pp.

## **Other sources**

Earthquake Engineering Field Investigation Team (EEFIT) (1993). EEFIT field investigations: Objectives and methods, leaflet.

Guide to Site Characterization for Engineering, Design, and Construction Purposes, ASTM D420-98, <http://www.astm.org>

Guidelines for Determining Design Basis Ground Motions, Report RP3302, March, 1993, Electric Power Research Institute, Palo Alto, CA.

## **7.5 Marine seismic observation (M. Shinohara, K. Suyehiro, and H. Shiobara, 2011)**

### **7.5.1 Deep ocean environment, required logistics, sensors and data recording**

#### **7.5.1.1 Introduction**

In this section, we describe the logistical, legal, environmental and technical requirements and conditions to be considered and met by an observer who attempts to acquire seismic data from the sea floor. Currently, most of earthquakes with magnitude ( $M$ )  $> 4.5$  can be detected anywhere in the world including the oceans (e. g., <http://earthquake.usgs.gov/>). Satellite altimetry data show us the large-scale seafloor topography (bathymetry) of tectonic features such as the oceanic ridges, trenches, and transform faults, which can be compared with global seismicity. Based on these general pictures, an observer can design seismic experiments to look into the heterogeneous structures of different tectonic provinces or locate and analyze background microearthquakes or aftershock sequences in the wake of major off-shore earthquakes, and so forth. We are motivated to use marine observations by the inhomogeneous distribution of land-based stations and insufficient azimuthal coverage of seismic sources and structural-tectonic features to be investigated. Such data resolve systematic biases in derived parameters and models better than data from only land-based seismic observations. Because the oceanic plates recycle through the Earth's crust and mantle through time, there are many fundamental problems to be studied. There are often important gains and it is sometimes necessary to make specific observations on the seafloor rather than from land, or in combination with land-based seismic networks. Examples of gains from including seafloor observations include: achievement of more complete azimuthal coverage, higher spatial resolution and accuracy in earthquake locations and in source mechanisms studies, and in tomographic velocity or attenuation models. Seafloor

observations avoid contamination or complications from data sampled over too large a distance and across major heterogeneities.

#### **7.5.1.2 Access**

For research purposes, most Ocean Bottom Seismograph (OBS) operations are temporary operations spanning from several to a few hundred days across a spatial scale from 10 to 1000-km. To work in the marine environment, one must first be able to bring the OBSs to the desired locations, then install them there, and to recover them after the mission is finished. This is done by ocean going vessels. Dropping OBSs from a helicopter has been done but recovery requires a boat. If the instrument needs to be positioned at a precise location or requires a special installation scheme such as burying the OBS, one needs extra methods such as utilizing an ROV (remotely operated vehicle) to reach the seafloor or a submersible with robotic functions.

##### ***The ship***

Depending on the size and the weight of each OBS, the maximum number of OBSs that can be put on board will be restricted. Experiments close enough to shore that the transit time is short may use a smaller boat and deploy groups of instruments during several short cruises. The availability of a ship is an important condition. Ships may be obtained either through application to organizations managing research vessel operations, by open competition through a proposal review system or if you have your own funding, by charter of a commercial vessel. Usually a management body will do the scheduling of the ship and one will need to adjust one's schedule to fit the ship schedule. In the case of a chartered vessel, one may have more control, but the ship's capabilities must be carefully looked at in the light of requirements and costs.

The ship must have the capacity to store all the OBSs away from severe vibration and temperature changes that could affect the seismograph system performance. Recently, we have sought to keep onboard work to a minimum and to keep the instruments inside their pressure housings. The instruments can be calibrated by using a calibration coil or applying an electronic test pulse (see Chapter 5, sub-chapter 5.6), and the stored data read without opening the cases. However, there could be situations where one needs to refurbish OBSs, so one needs a clean and dry lab with appropriate space and access to the deck. Such a lab could be a container lab that can be brought to ship for the experiment.

Active controlled source OBS experiments using an airgun array and hydrophone streamer to simultaneously monitor the reflected seismic waves require a ship's crew trained in this type of operation, and a vessel able to run in a straight line at a constant speed across the ground. Such methods are common in the oil industry, but less common in the academic research world. Details will not be given here, but in general, there is the mechanical system to tow the airguns and the streamer, a pneumatic system with an air compressor system sufficiently powerful to shoot the airguns at the desired pressure (100-150 kg/cm<sup>2</sup>), and repetition rate (10s of seconds), an electrical system to control the shooting sequence and to log the sequence and to record the data from the streamer. All these have to be precisely controlled in time and location, usually utilizing the Global Positioning System (GPS). In the case to use a large ship, an antenna position of the GPS system must be considered. Handling high air pressure equipment requires extra safety considerations.

### ***Ship capability***

Once one knows the OBSs can be put onboard, the other things to consider are the electrical, mechanical, and Internet Technology (IT) specifications on the ship one plans to utilize to service the instruments. Electrical voltage and frequency (and its accuracy) are some elementary factors to know. If the ship is used for both launching and recovering from the sea a free-fall pop-up package, one will need to discuss with the ship operator how to conduct these operations. One may need more than one method depending on the sea and wind conditions and so widen the margin for successful operations. The release into the water and the capture from the water are the times when things may go wrong such as unwanted impact of the instrument into the ship's hull.

GPS is available on every research vessel, but you will want to know how to obtain the data in digital format including timing information. Keeping accurate time on a ship has been a challenge for many centuries, but is now solved to microsecond accuracy by satellite (GPS) technology. The ship must also be equipped with an accurate depth measurement system. Recent swath bathymetry data are desirable for choosing the drop point of the OBS and so avoiding locally steep slopes that may not let the seismic sensor in the OBS be leveled. Existing bathymetric maps of an area can be used to pick locations for OBSs if the map was made with GPS navigation. GPS based bathymetric maps have an accuracy of better than 100 m, whereas maps based on older data may be too poorly navigated to be useful for choosing OBS locations. Even with a map, checking with a deep echo sounder to be certain of the depth before the OBS is deployed should be attempted.

Communication to OBSs is done by sending and receiving acoustic signals from the ship. Sometimes it becomes difficult to maintain this communication link because of ambient noise, particularly from the ship's engine. Testing and understanding the communication link in various situations is an important practical knowledge to have for successful recovery of OBSs.

### ***Applying for and sharing ship time***

Applying for and sharing ship time is different in each country. There is an international database of ocean research vessels where one may find websites of interest (<http://www.pogo-oceancruises.org>). After getting a successful proposal through peer review and the proposal is accepted, one may find one is required to share a cruise with other research groups or that one has been allotted less than the desired number of cruise days. In such cases, one must prioritize to achieve one's goals within the plan. Usually there will be a chief scientist assigned for such coordination well before the cruise. Unlike on land, there may be scientists from different disciplines such as physical and chemical oceanography or marine biology on a cruise. Even within the solid Earth sciences, there could be geomagneticists, geologists, etc. One may need to explain and demonstrate the importance of your science.

### ***Legal matters***

It is necessary to know under which jurisdiction lies the area of your interest (see [http://www.un.org/Depts/los/convention\\_agreements/texts/unclos/UNCLOS-TOC.htm](http://www.un.org/Depts/los/convention_agreements/texts/unclos/UNCLOS-TOC.htm)).

Generally working in the open sea poses no problem for scientific research except that one needs to comply with international rules protecting the environment. If it is within your territorial waters or within your EEZ (Exclusive Economic Zone), you must notify your own country's authorities. If it is within foreign countries' EEZs, permission must be sought through proper channels (which may be your funding agency). It is important to remember that it may require 8 months or more for permission to be either granted or denied. If EEZs are not well defined due to territorial disputes, you may not be able to conduct any

experiments. Clarifications will require consultation with your Foreign Ministry. There could be different conditions to be met in order to conduct seismic observations in these waters on a case-by-case basis. Difficulties may be encountered in conducting active source experiments due to environmental reasons. Relevant information should be available from the managing body of the research vessel. If the vessel enters foreign waters, then one must follow export/import rules for that country, and also the security export rules for your country of origin.

### ***Environment***

Keeping marine ecosystems healthy is recognized as being more and more important. One should not pollute the ocean nor significantly influence the ecosystem. OBS observations should not pose much of a problem in this respect. When conducting an experiment, one may be required to negotiate with fishermen through respective authorities. Unlike commercial companies who may pay compensation to the fishermen for obstructing their work, scientific researchers have traditionally negotiated agreements allowing them to conduct experiments subject to certain time and space limitations but without paying compensation. This means one must know the fishing seasons for different groups of fishermen working in the area of interest. There could also be military operations in the area of one's interest, and one should consult with pertinent authorities such as the Coast Guard. Shipping lanes can be noisy for seismic observations. The best weather window should be sought to avoid stormy, windy, or foggy seasons.

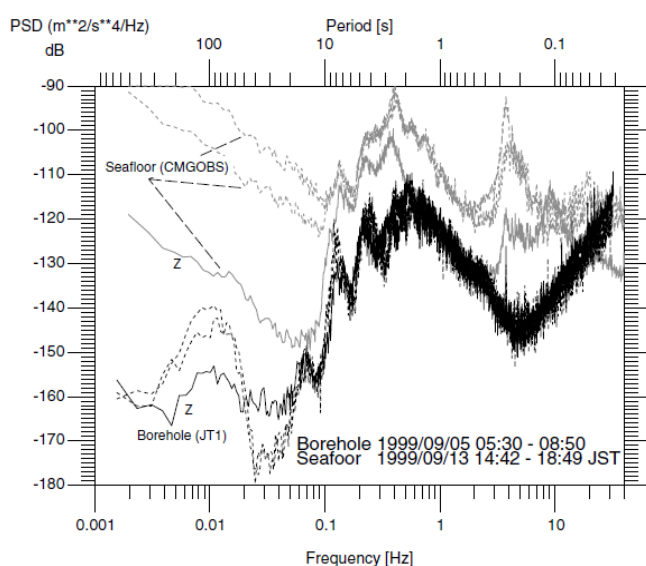
#### **7.5.1.3 Sea water and unconsolidated sedimentary layer**

Any OBS on the seafloor has a sea water layer above it so that corrections for topography are required when analyzing data. The water layer at the seafloor cannot be considered a free surface and needs to be appropriately treated unless one is only interested in traveltimes (e.g. Shinohara *et al*, 1994; Thorwart and Dahm, 2005). For controlled source experiments, where shots occur at the sea surface, how the seismic ray travels in the water column can be important. The sound velocity in the water column has a depth profile that changes geographical area and season and is mainly controlled by lateral and vertical variations of water temperature and salt content. For example, water temperatures may be very different in the same latitudes, e.g. due to the warm Golf stream in the Atlantic or the cold streams on the west coast of North America, or when going along an E-W profile from the Mediterranean Sea with 4.5% salt content to the Azores with about 3.5%. Including these considerations in modeling ray paths rather than assuming a constant velocity layer in the water column leads to more accurate structural models.

An unconsolidated sedimentary layer is almost ubiquitous at the ocean bottom. This layer, which can be from a few hundred meters to a few km (or more in some cases) in thickness, must be treated appropriately in data analysis. The P velocity is only slightly faster than the velocity in the seawater in this layer because the pore water dominates the sediments, but the layer is capable of shear wave transmission. There can be a rapid velocity increase with depth as the pores are squashed. OBS data cannot distinguish this layer very well since all refracted phases travel nearly vertically in this layer to the OBS. Resolving how the shallow sedimentary layer varies in space may require that reflection seismic data be acquired over the OBSs. Usually the top of the igneous crust does not run parallel to the seafloor bathymetry as sedimentation occurs after formation of the crust.

### 7.5.1.4 Noise

The quality of data from the seafloor stations is lower than that of data from most land stations because of the seafloor's much higher noise environment. The ocean surface generates microseisms (Longuet-Higgins, 1950) that at some frequencies may be much stronger than magnitude 6 earthquake arrivals at teleseismic distances (Webb, 1998). Understanding the nature of seismic noise at sea floor is important to understanding seismic observations on and below the sea floor. Unfortunately, noise is also generated by the interaction of seawater flow with the seafloor or with the seismometer at the sea floor. The flow at the bottom of the ocean can degrade long-period performance at 10 sec and longer by directly tilting the seismometer or by deforming the surrounding media (Webb, 1988). The noise amplitudes from tilting can be larger than the amplitudes of most long-period teleseismic phases especially on horizontal-component records. For examples, characteristics of long-period noise in the Atlantic and Tyrrhenian Sea have been obtained using hydrophone, pressure gauge and seismometers (PMD) (Dahm *et al.*, 2006). Seismometers in boreholes into the sea floor, present the best environment to study the nature of seismic ambient noise at the sea floor by avoiding flow noise (Fig. 7.71). Crawford *et al.* (2006) showed a results that the seismic noise in an unsealed borehole is larger than that on the sea floor by removing tilt noise due to fluid flow from the data from the 1998 Ocean Seismic Network Pilot Experiment off Hawaii. Their result indicates that cementing of seismic sensors in a seafloor borehole is essential to reduce a seismic noise in a seafloor borehole.



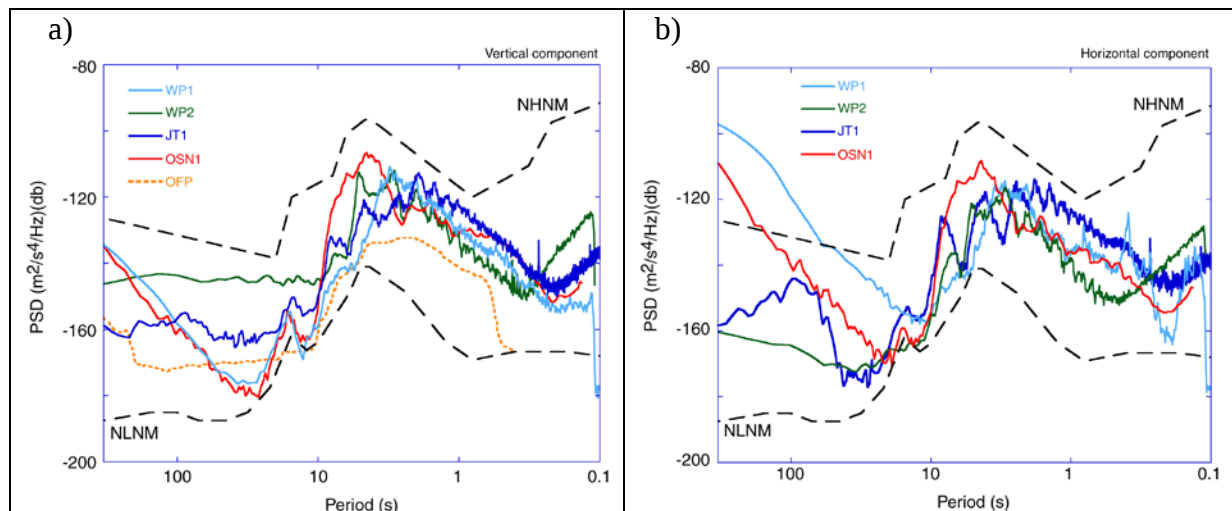
**Fig. 7.71** Difference of typical noise power spectra in a borehole installation and a seafloor installation near the borehole of site JT1 (see Fig. 7.80). The seafloor observation is noisier than that in the borehole and the horizontal components are much noisier than the respective vertical (Z) components (after Araki *et al.*, 2004).

In 2000 and 2001, state of the art borehole broadband seismometers were installed in the northwestern Pacific Basin (WP-2) and the Philippine basin (WP-1) (Shinohara *et al.*, 2006). The depths of sensors are 561m and 460 m below the seafloor. The seismic data from the WP-1 and WP-2 seismological observatories enable the study of broadband seismic noise in the frequency range from 3 mHz to 50 Hz in the northwestern Pacific over an observation term covering more than one year of continuous recording. Temporal changes in ambient seismic noise within deep-sea borehole can be analyzed in depth using this data (Fig. 7.72). The noise level as a function of time and period for vertical and horizontal components of the WP-1 and the WP-2 observatories were obtained. The major advantage of borehole observatories is that there are little or no temporal change in noise levels for periods longer than 10s. It is found that the noise levels in the vertical component at the WP-1 reach -180 db (rel  $(\text{m/s}^2)^2/\text{Hz}$ ) for

time-periods between 10 s and 100 s, much lower than noise levels for instruments on the seafloor. In contrast, small temporal variations (maximum fluctuations of 10 dB) in noise levels for periods about a few seconds are seen, which can be correlated with the seasonal variability of ocean microseisms (see Chapter 4).

The WP-1 observatory recorded relatively large noise levels in summer and fall at periods of 4 - 8 s. The vertical component of the WP-2 observatory has minimum noise levels about -145 db, which is relatively high noise level. The high noise level of the vertical component of the WP-2 may be due to damage during installation of the sensor. The noise levels in the horizontal components of WP-2 seismometers are -160 db for periods between 10 s and 100 s and thus lower than that of the vertical component. The records from the WP-2 observatory reveal high noise levels in a period range of a few seconds during winter, which is thought to be the mean stormy season which also generates the largest ocean microseisms observed on land-based seismic stations. For both the vertical and the horizontal components of both the observatories, the noise levels are close to the New Low Noise Model (Peterson, 1993) near 10 s period. In general, the peak in the noise spectrum near a period of 100 s in measurements taken at the seafloor or in a shallow-sea borehole is due to deformation of the seafloor by ocean gravity waves (Webb, 1998). Araki *et al.*, (2004) reported that installation of a seismometer in the basement rocks minimize the effects of the ocean gravity waves. Because all the seismometers of the WP-1 and WP-2 observatories were installed into basaltic basement, there is no peak due to the ocean gravity waves.

The horizontal component records of the WP-1 observatory have larger noise levels than the vertical component records for periods longer than 20 seconds. There is a possibility that the horizontal sensors in the WP-1 seismometer have a coupling problem. In contrast, the noise level of the vertical component for periods longer than 10 seconds in the WP-2 observatory is larger than those of the horizontal component. The vertical sensor of the WP-2 observatory may be damaged. In Fig. 7.72 we have plotted representative noise power spectra measured at the WP-1 and WP-2 borehole sites, together with respective spectra from 3 other borehole sites in the Western Pacific (JT1; for positions, also of WP-1 and WP-2 see Fig. 7.80), the North Atlantic (OFP), and the Central Pacific off-coast Hawaii (OSN1). The spectra of the WP-1 and WP-2 is estimated from one hour record for quite season of each site. Note that all borehole installations have almost the same noise level.



**Fig. 7.72** Comparison of ambient seismic noises from seafloor borehole installations: a) for vertical component; b) for horizontal component. WP-1, WP-2 and JT-1 denote borehole



observatories in the western Pacific. OFP and OSN-1 indicate borehole observatories in the North Atlantic Ocean (Montagner *et al.*, 1994) and the Pacific Ocean off Hawaii (Stephen *et al.*, 1999), respectively. NHNM and NLNM indicate the New High Noise Model and the New Low Noise Model by Peterson (1993), respectively.

#### **7.5.1.5 Sensors**

Seismic sensors for land stations are selected according to types of observations needed. The possible choices of sensors is not very wide for OBS observations because the requirement for low power consumption and the limited volume available for the sensors within most pressure capsules. For more general information on the theory, principle design and calibration of seismic sensors and their technical specifications the contributions by E. Wieland (Chapter 5 and DS 5.1).

##### ***Geophones***

The 4.5-Hz exploration-type geophone is mainly used for local and regional earthquake observations. This geophone is also useful for seismic exploration using airguns and explosives. The geophone has the advantage of no power-consumption and small size. The geophone measures signals from approximately 1 to 100 Hz, with a corner frequency at 4.5 Hz. They have a flat response to ground velocity for frequencies greater than this corner frequency. A typical example is the Mark Products L-28. The 2Hz geophone such as the Mark Products L-22 is also used, however, the volume of this sensor unit is larger.

##### ***Active type (moving coil)***

There are also active seismic sensors (Lennarts) on the market which extend the flat velocity output frequency range of 4.5-Hz geophones electronically (see DS 5.1). This sensor has the advantage of small size for its corner frequency of 1Hz. Because land observations for local earthquakes often use 1-Hz sensors, this sensor is useful for combined analysis of land and OBS data.

##### ***Active type (liquid)***

The seismometer (PMD) (see DS 5.1) is a three-component seismometer having a fairly broad frequency band response. The velocity amplitude response is essentially flat from 25 to 0.03 Hz with useful output down to 0.01 Hz. The PMD seismometer sensors are robust and require no leveling or clamping. This makes them particularly suitable for sea floor observation. The inertial component is a liquid—water with potassium iodide in solution—in a container which includes permeable grids that serve as cathode and anode. Between the grids a small bias voltage ( $<0.7$  V) is applied, resulting in an ion current. Vibration of the instrument produces changes in the motion of the ionized liquid through the electrodes, resulting in a modulation of the ion current. These modulations are the output signal of the seismometer. From observation, the power spectral density instrument noise level is about -150 dB [ $(\text{m/s}^2)^2/\text{Hz}$ ] at frequencies between 30 and 10 mHz. This is  $\sim 13$  dB higher than for a CMG-1T broadband seismometer. Below 30 mHz, the performance deteriorates (Shiobara and Kanazawa, 2009).

##### ***Servo-type accelerometer and MEMS***

The up-to-date servo accelerometer or MEMS is also suitable as a seismic sensor for sea floor observation, because they have high resolution and stability. The primary advantage for this type of accelerometer is its compactness, however the power consumption is large and the noise floor of these sensors is slightly higher than the New High Noise Level (Peterson, 1993;

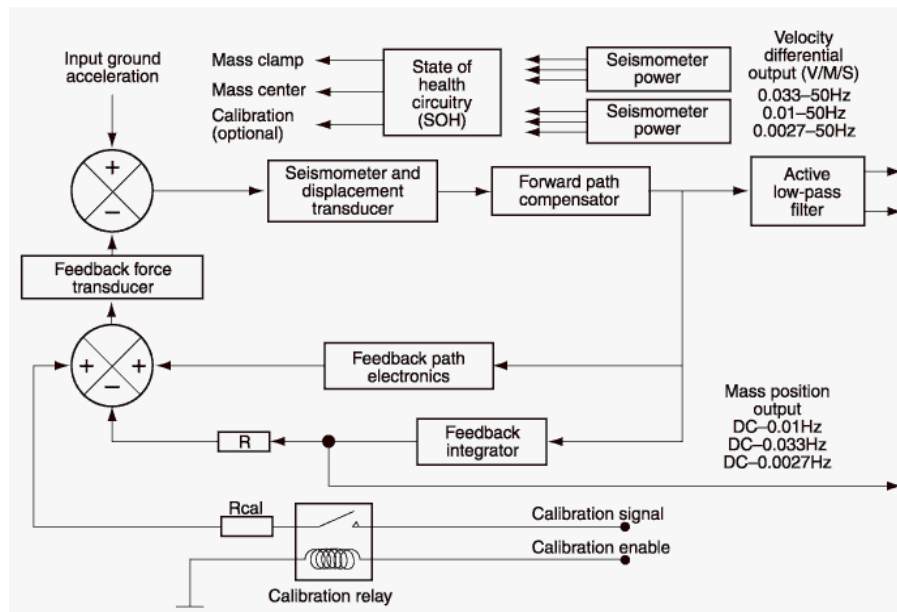
see also Chapter 4). However, because the accelerometers can measure accelerations larger than 1 g, these are useful for studying the source processes of large nearby earthquakes.

### ***Broadband sensors (CMG-1T, CMG-3T)***

Broadband (BB) sensors have a flat response to ground velocity from approximately 0.01 to 50 Hz. Typical examples are the Guralp CMG-1T or CMG-3T seismometer (for specifications see DS 5.1) with frequency range from 0.03 to 50 Hz. The BB sensor provides more information compared to conventional short period (SP) seismometers. However, BB seismometers are more expensive, more fragile and more difficult to deploy, operate, and maintain. Data retrieval and analysis can be more complicated than for geophone sensors. BB seismometers use active feedback to the sensor mass and so require a stable, large power supply. In addition, the seismometers may require control of mass centering so that the recorder must continuously monitor the status of the sensor and autonomously control the sensor. Since BB sensors record microseisms over a much wider period range their raw output signal contains much more seismic noise than signals from a short-period geophones. This means that the recorder needs to have a larger dynamic range for A/D conversion.

The vertical sensor in the Guralp sensors is a modified Lacoste type, and the horizontal sensor is an inverted pendulum. The inertial mass is a boom (a solid machined beam) supporting a transducer coil. The vertical sensor mass is supported by a prestressed triangular spring to support its weight and has a natural period of  $\sim 0.5$  s. The horizontal sensor mass is centered by an unstressed flat triangular spring and has a natural period of  $\sim 1$  s. The effective mass of each sensor is  $\sim 180$  g. The springs are connected to the frames with a temperature compensating wire that minimizes the effect of temperature variation. A compact design is achieved chiefly by the short stiff springs and short boom. The adjustments required for operation consist of leveling the boom of the vertical sensor and tilting the bases of the horizontal sensors to center the mass movements in their equilibrium positions. These are made by small (1 cm in diameter and 3 cm long) direct current (DC) motors operating gear mechanisms to tilt the horizontal sensor bases and to apply a small extra force to the vertical sensor's boom. Before and during the installation, the instrument may be subjected to severe motion that can damage the mass support hinges. Therefore, the masses must be locked securely in their frames so that the hinges can be released after installation. This operation is performed by a small motor-driven clamp, which is controlled by a recorder.

The sensors employ feedback to expand their bandwidth and dynamic range. The sensor's response is determined by the characteristics of the feedback loop. The mass position is sensed by the capacitive position sensor. The voltage from the sensor, which is proportional to the displacement of the mass from its equilibrium position, is amplified and fed to a coil on the mass. The current in the coil forces the mass to its equilibrium position. With a high loop gain, motion of the mass is essentially prevented; the feedback voltage is then a measure of the force and, thus, the acceleration applied to the mass. The block diagram of the feedback system is shown in Fig. 7.73. The system velocity response is identical to that of a conventional long-period sensor with a velocity transducer whose natural resonant period is 360 s with a damping factor of 0.707. The velocity output (flat to 100 Hz) is low-pass filtered ( $< 50$  Hz) before digitization. The mass position output can be used for periods longer than 360 s.



**Fig. 7.73** Example of block diagram of broadband sensors. Frequency ranges depend on product model (Guralp model).

### ***Levelling of sensors***

It is more difficult to level seismic sensors on the sea floor because humans can not access the sensor in the OBSs directly. Although some sensors do not require leveling, most seismic sensors do. There are two types of leveling systems; passive and active leveling. The passive leveling system uses a gimbal mechanism that levels under the force of gravity and strong damping from placing part of the gimbal mechanism in a high viscosity liquid such as silicone grease. Although this system has a simple mechanism, it is difficult to level sensors with natural periods of more than one second with sufficient accuracy. For broadband sensors, an active leveling system is used. The active system also has a gimbal mechanism, however, the gimbal mechanism is driven by micro-motors and fixed by brakes. The leveling system has two tiltmeters and sometimes also a compass. The Central Processing Unit (CPU) reads the tilt of the sensors and controls the motors to level the sensors. The system levels the sensors before starting the observation and after deployment on the sea floor,. In addition, the system checks the level daily. Both passive and active leveling systems typically work over angles of up to 20 degrees.

### **7.5.1.6 Data recording**

#### ***The necessity of low-power compact systems***

All autonomous OBS systems require a low-power recording unit, because the power-supply will be limited. For example, the power consumption of a recorder developed by Earthquake Research Institute (ERI), the University of Tokyo, is less than 0.3 Watts. The system must be compact due to the limited capacity of most pressure vessels. All system units including the sensors, recording system and battery pack must be accommodated within a sphere type pressure vessel of about 50cm diameter.

### ***A/D conversion***

Modern data acquisition systems are based on continuous integration and oversampling in combination with carefully designed digital low-pass filters. In most modern seismic recording systems, a Delta-Sigma-Modulator is used for analog to digital conversion (ADC). The Delta-Sigma ADC needs much electric power, so that it is difficult to use in the conventional way for OBS systems. For some systems the ADC is turned on intermittently to reduce the electric power consumed by the OBS recording system. Some systems still use ordinary successive approximation register A/D converters and sample-and-hold techniques to reduce power-consumption.

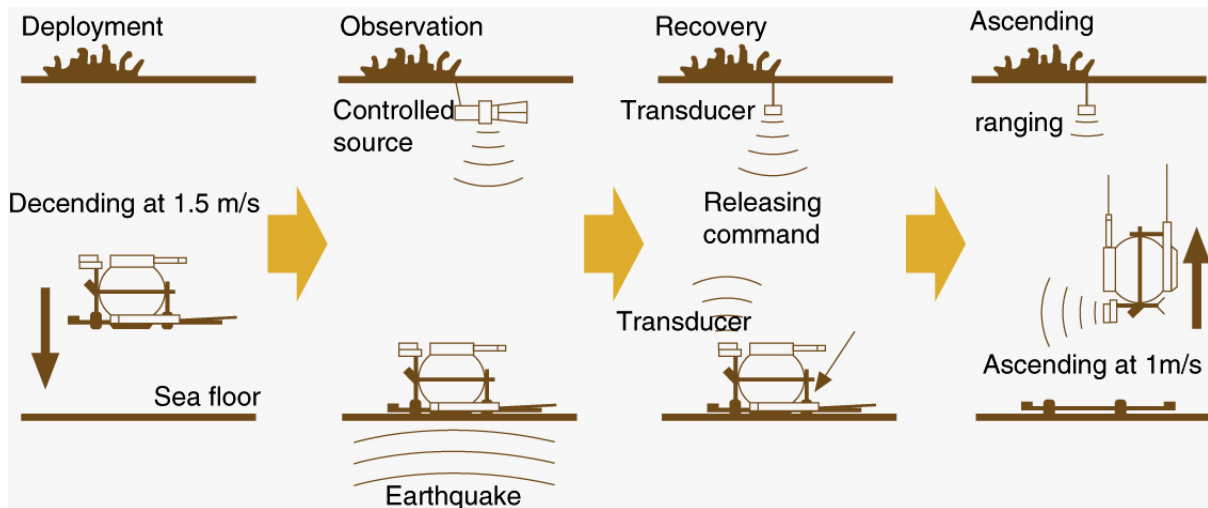
#### ***Data storage media (Tape, MO, HD, Memory)***

In OBS observations with continuously recording instruments, an OBS collects about 30-50 M Bytes of data per day for a 3-component station with 100 Hz sampling rate. Early OBS recorders were based on the Analog Direct Recording method in order to record this huge amount of data. Later quarter inch tape and other cassette tapes and in the early 1990s, Digital Audio Tape (DAT) was often used (Shinohara *et al.*, 1993). At present hard disks (HD) are common as mass storage technologies in OBS systems. HDs are reliable, low-cost and robust and allow for much faster data access than other media for mass data storage like tapes. However, the power consumption is significant when the disk is spinning up. An advantage of hard disks as opposed to other magnetic- and opto-magnetic media is that they are encapsulated and less sensitive to rough and dusty environments. For connecting the hard disk to the recorder, many systems use the Small Computer System Interface (SCSI) others use the Integrated Drive Electronics (IDE) interface. With respect to data transfer, SCSI systems are more reliable than IDE systems, however the latter are less costly. Memory systems such as Compact Flash card (CF) and Secure Digital memory card (SD) now have sufficient capacity for data storage and are rapidly replacing other media in OBSs. For example, the capacity of SD now exceeds 32 GB, which is enough for continuous recording for a few months. These memory systems have no moving parts, and are quite robust.

## **7.5.2 Ocean Bottom Seismograph (OBS)**

### **7.5.2.1 Introduction**

There are two types of equipment for seismic observations in marine environments: pop-up type OBSs and cabled OBSs. The pop-up type OBS is a stand-alone system. These are deployed on the sea floor by free-fall from a research vessel, and are recovered by pop-up after releasing an anchor. The advantages of the pop-up OBS are low-cost for the equipment and observations and that they can be deployed anywhere. However, the data are obtained only after recovery of the OBSs (Fig. 7.74). At the present, pop-up OBSs can be classified by type of sensor, and observation term. Cabled OBS can send data in real-time and power is supplied from land. This means there are fewer limits for power consumption by the OBS. Real time data is required for earthquake monitoring in the context of alarm and disaster mitigation schemes. The disadvantage of cabled OBSs are the high cost. As examples both types of OBSs, as developed and used in Japan, are described in the following in more detail. For technical specifications see section 7.5.4.



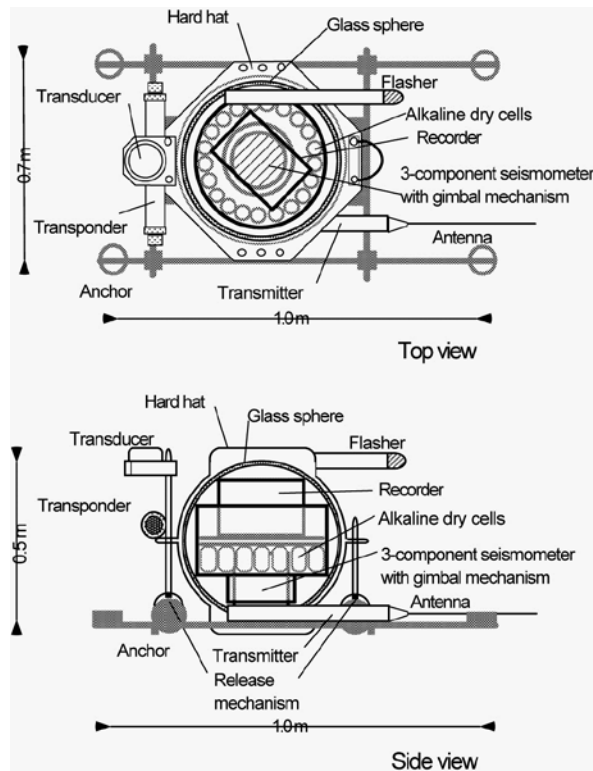
**Fig. 7.74** Schema of seafloor observation using pop-up type OBS.

### 7.5.2.2 Pop-up type OBS

#### *Short-term OBS*

In recent years, ocean bottom seismometers (OBSs) have been used to obtain crustal and upper mantle seismic structure and the distribution of earthquakes. Before the 1990's, the OBSs used in these experiments worked with analog recording tape. One of the advantages of analog recording OBSs was that they could record data continuously for more than two weeks. A disadvantage of the analog OBSs was their limited dynamic range per recording channel.

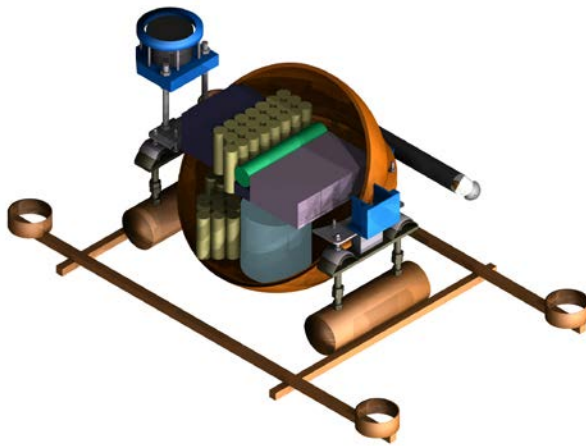
A digital recording short-term OBS (ST-OBS), which has the same recording length of time as the analog OBSs, was developed to obtain higher quality data. The ST-OBS consists of one pressure vessel, an acoustic release system, a radio transmitter, and a strobe light (Fig. 7.75). The pressure vessel (glass sphere) contains sensors, recorder and battery cells. The design of ST-OBS, except for the digital recorder, is the same as the analog recording OBS. First, a commercial portable digital audio tape (DAT) recorder was used to store seismic data to obtain long-term recording at low power and low cost. Next, 2.5 inch SCSI hard disk was used to store data. The use of hard disks has allowed a longer recording period. The latest recorder has memory media such as Compact Flash memory (CF) or SD memory. The first practical digital recorders had 16-bit A/D converters and thus a dynamic range of 96 dB. Nowadays, 24-bit A/D converters are common which allow a dynamic range of up to 144 dB. Recorders usually record 3 or 4 channels of data continuously at 100 or 128 Hz sampling rate. A real-time clock (RTC) with an accuracy of order  $10^{-8}$  controls timing of the recording system including the timing of A/D conversion, the starting of data acquisition at a predetermined time, and so on. The electric setup and monitoring of the recorder inside the sealed vessel, including adjustment of this RTC, can be performed by connecting a computer to the recorder through the pressure vessel. A timing clock pulse is output to outside of a pressure vessel allowing comparison of the RTC of the recorder to a time standard provided by a GPS receiver. The data on the recording media is retrieved after recovery an OBS.



**Fig. 7.75** Top view (upper) and side view (lower) of ST-OBS layout. The glass sphere housing instruments has a positive buoyancy and is released by acoustic command. External mechanical configuration has not changed from that of the analog recording OBS (after Shinohara *et al.*, 1993)

### Long-term OBS

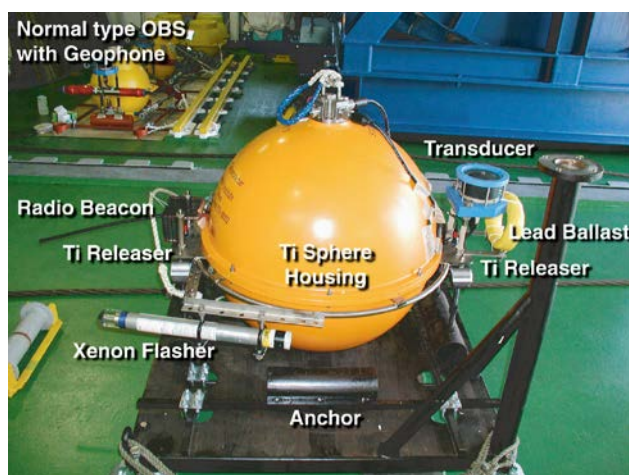
Long-term OBSs (LT-OBS) were developed at ERI to observe earthquakes continuously on the sea floor over more than one-year. For the housing of the LT-OBS, a titanium-alloy sphere made in Russia with 50 cm diameter was selected, so as to avoid problems from corrosion of the pressure vessel (Fig. 7.76). The titanium-alloy vessel was tested for use at 6,000m depth before shipment from the maker. Lennartz's LE-3Dlite is used for the short-period sensors in this OBS operating with a velocity-proportional response in the frequency range between 1 and 80 Hz. Three component sensors are mounted on a leveling system. The leveling unit is equipped with two tiltmeters which have 0.1 degree accuracy. The outputs of the three seismometer components are amplified by a low-noise and low-power analog unit, and are converted to 22-bit digital signals continuously. The sampling frequency is either 200Hz or 128Hz. Data are stored in memory temporarily. When the memory is full, the data on the memory is transferred to Hard Disk (HD). Power to all electric circuits (CPU, sensors, leveling system, and HD) are supplied by lithium battery cells. Each cell has a power capacity of 30Ah. Although the number of battery cells depends on the intended duration of recording, approximately 50 battery cells are needed for one-year of observations. The LT-OBS is equipped with an acoustic release and communication system. The most important function of the acoustic system is to release a weight for recovery. After an acoustic release command is sent from a shipboard unit, the acoustic system on the OBS provides electric power to a weight release unit causing rapid electrical corrosion of the strap holding the weight. Another function of the acoustic system is communication between the sea floor and sea surface. The recording unit on the OBS can be controlled from the on-board unit, for example, starting or stopping recording, leveling the sensor, and checking the status of the recording unit. This type of LT-OBS is widely used for monitoring earthquake activity in Japan. At the present, approximately 200 LT-OBSs are in operation in Japan.



**Fig.6** Perspective cutaway of the LT-OBS. The titanium-alloy vessel contains seismic sensor, recorder and batteries. The duration of recording of the LT-OBS reaches one year.

### ***Broadband OBS***

Based on long experience with the short-period OBS, a broadband OBS (BBOBS) has been developed at ERI as well. This OBS is equipped with broadband sensors. This system is also mobile, compact, reliable and easy to operate and enables long-term observations. These BBOBSs have been used since 1999, yielding good scientific results and offering more opportunities for observations, especially in the long-period range. The design of the BBOBS (Fig. 7.77) is quite simple, as all units are inside of a housing made of titanium. These BBOBS also use self pop-up for recovery. The key design points were low power consumption of the sensor and recorder including the precise RTC clock, and low weight for the OBS so as to maintain positive buoyancy despite the limited volume of the housing. The titanium-alloy pressure vessel has a diameter of 65 cm, which is slightly larger than the LT-OBS, as required to contain the BB seismic sensor with its larger volume and additional batteries needed to power the BB sensor. The sensor was specially made for our active leveling unit and it is possible to physically separate the mechanical part and the main electronic boards. When the BBOBS is deployed at the deep sea floor, the inside of the housing is an ideal operating environment for the sensor (and also the crystal-oscillator clock) because of small temperature variations experienced and absence of human activity compared to land observatories. The amount of the data stored is about 4 GB for one month if the sampling rate is 200 Hz for three components and when using a completely recoverable compression method. The quality of the data is described in a later section.

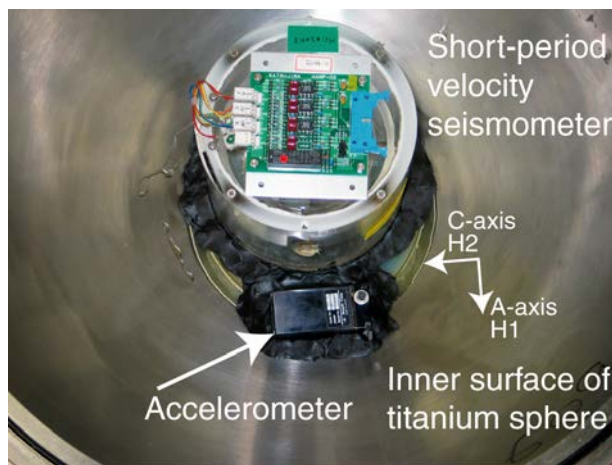


**Fig. 7.77** Photograph of the BBOBS before deployment. The shape of the BBOBS is similar to that of the LT-OBS, however, the pressure vessel is larger than that of the LT-OBS due to sensor size.



### ***Ocean bottom accelerometer (OBA)***

To understand the characteristics of large earthquakes which occur in a subduction zone, it is necessary to study the asperities, where large earthquakes may occur repeatedly, from a close distance. Since a conventional OBS is designed for high-sensitivity observations, OBS records of large earthquakes which occur nearby are often saturated. A servo-type accelerometer is well-suited for recording large amplitude seismic waves. However it was difficult to use such an accelerometer in OBSs due to the large electric power consumption. Recently, a servo-type accelerometer with a large dynamic range and low-power consumption has been under development. Alternatively, one can use a larger size titanium sphere pressure vessel for the OBS and so contain many more batteries. For long-term sea floor observations of large earthquakes, a small three component accelerometer is installed into conventional long-term OBSs for acquisition of low-sensitivity (strong motion) accelerograms of the sea floor. For example, we use a compact three-component servo-type accelerometer with a weight of 53 grams as the seismic sensor. Measurement range and resolution of the sensor are 3 g and  $10^{-5}$  g. The sensor is directly attached to the inside of the pressure vessel (Fig. 7.78). Signals from the accelerometer are digitally recorded to Compact Flash memory with 16 bit resolution and a sampling frequency of 100 Hz. The OBAs have been deployed several times and accelerograms obtained on the sea floor. One earthquake with magnitude of 7 was recorded by the OBA at an epicentral distance of approximately 20 km. The maximum acceleration reached  $5.5 \text{ m/s}^2$ . Continuous improvements are being made to the system to obtain better quality data.



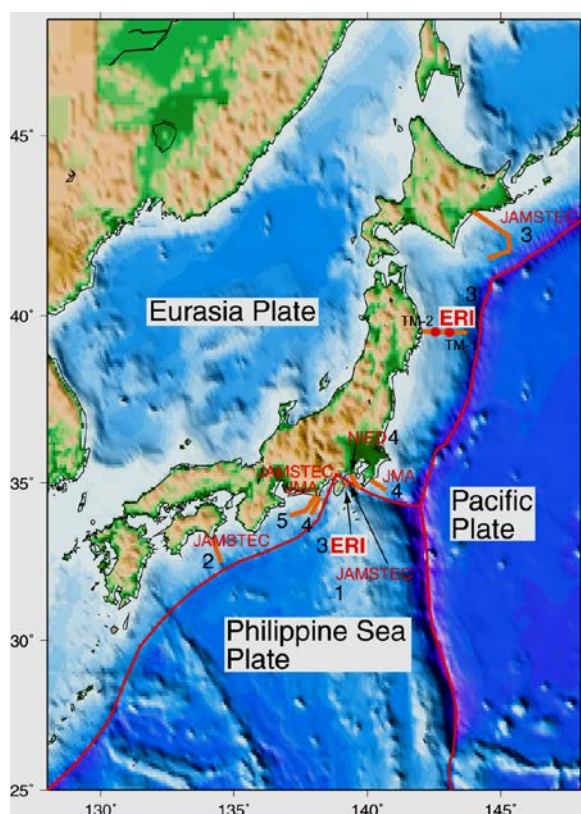
**Fig. 7.78** Installation of the accelerometer to the inside of the pressure vessel (Type 1). The accelerometer was fixed beside the conventional three-component seismometer with gimbal mechanism with the same orientation (after Shinohara *et al.*, 2009).

### **7.5.2.3 Cabled OBS system**

In this section some of the existing and planned OBS cabled systems are described. In Japan, a number of optical-cable linked ocean bottom seismic stations and tsunami stations have been deployed since the 1990's (Fig. 7.79). Cabled geophysical observation systems are deployed off Hokkaido, Sanriku, Boso, Izu, and Shikoku. In 1994, the Earthquake Research Institute, University of Tokyo deployed a cabled OBS network system around an earthquake swarm area off Izu. Data from three seismic sensors on sea floor are transmitted on-line by using optical fibers to a land station. In 1996, an ocean bottom geophysical observatory using a sea floor optical fiber cable was installed off Sanriku. The Sanriku system has three seismometers and two tsunami gauges. These data are transmitted to a land station in real-time by individual fibers. The sea floor optical fibers are buried in areas where the water



depth is less than 1,000 meters to protect the cable from fishery activities. These systems provide integrated observation from both on land and under the sea. The disadvantages of the existing cabled OBS systems are their high cost and limited adaptability. Once the system is deployed, it is very difficult to add sensors for new observation. The Japan Agency for Marine-Earth Science and Technology (JAMSTEC) is constructing a new cabled OBS system using wet-mate-able connectors. Sensors can then be replaced using a Remotely Operated Vehicle. The system will be deployed off central Japan where a huge earthquake is expected to occur. These existing systems are based on the technology of trans-ocean communication cables which need high reliability. However, since the existing system is so expensive, it is difficult to deploy many observation nodes for experimental use, for example when it is important to observe earthquakes using a spatially high density seismic network. Therefore a new cabled OBS system is being developed using Internet-Technology to lower the cost for spatially high density seismic observations (Kanazawa and Shinohara, 2009). This system is planned to be deployed in the Japan Sea during 2010.

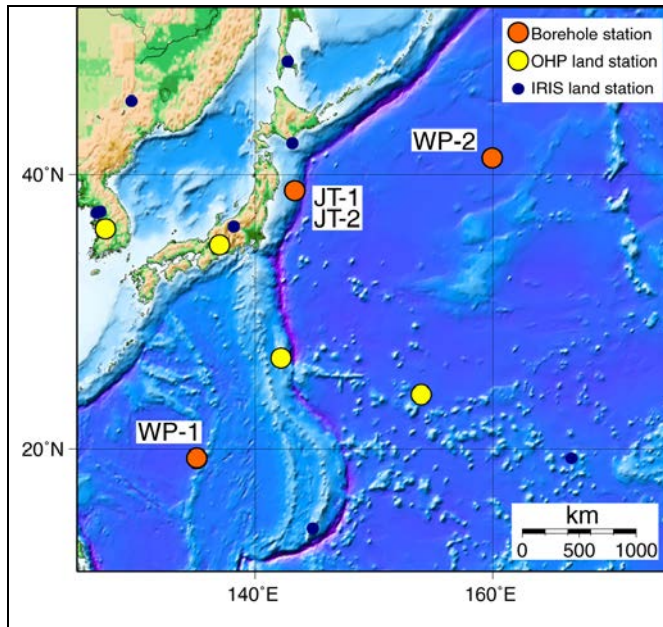


**Fig. 7.79** Locations of cable OBS system for local earthquake observation around Japan. Numerals indicate the number of observation nodes for each system. Operating agents are also shown. Note that the existing cable OBS systems have a small number of observation nodes. TM-1 and TM-2 indicate positions of real-time tsunami-meter.

#### 7.5.2.4 Ocean floor borehole system

The Japanese Ocean Hemisphere Network Project installed four borehole geophysical stations at three sites covering a period from 1999 to 2001 (Fig. 7.80). In 1999, two borehole stations (JT-1 and JT-2) were installed during Ocean Drilling Program (ODP) Leg 186 (Sacks *et al.*, 2000) and are located immediately above the inter-plate earthquake generation zone on the landward side of the Japan Trench. The JT-1 and JT-2 stations were installed approximately 50 km apart and have two broadband seismometers, a strainmeter and a tiltmeter for each station. In August 2000, during ODP Leg 191, the borehole seismological observatory WP-2 was successfully installed at ODP Site 1179 in the northwestern Pacific basin (Kanazawa *et al.*, 2001). Borehole seismic observatory WP-1 was installed in the west Philippine basin during ODP Leg 195 in April 2001 (Salisbury *et al.*, 2002). The WP-1 and WP-2

observatories each have two identical broadband seismometers (Guralp CMG-1TD). Together, these observatories can cover a large expanse of the northwestern Pacific in areas that comprise major gaps in the coverage provided by pre-existing global seismological observatories. At present, the data from these four borehole observatories are retrieved by ROV visits at an interval of about 0.5 - 1.5 year (Suyehiro and Mochizuki, 2002). Araki *et al.* (2004) have already reported on the JT-1 and JT-2 observatories and the characteristics of the geophysical data obtained from the JT-1 observatory.



**Fig. 7.80** Location map of broadband seismic station coverage in the north-west Pacific. Small blue and yellow circles indicate land seismic stations, orange circles are seafloor borehole observatories, which were installed by drilling vessel. The JT-1 and JT-2 were deployed in 1999. The WP-1 and WP-2 observatories were installed in August 2000 and April 2001, respectively. These sea floor observatories successfully captured broadband seismic data for a few years.

#### 7.5.2.5 Dealing with a multiple-OBS long-term dataset

As LT-OBS and BBOBS data became more available, a huge amount of the data is being obtained from these OBSs. One ST-OBS generates less than 100 Mbytes of data during a few months of observations. For one year of observation using LT-OBSs, the amount of the data becomes approximately 20 GB for each LT-OBS. A single array deployment for earthquake observations can use more than 20 LT-OBSs, so that the total amount of the data from the OBS network will be about 0.5 T bytes. The data collected by the OBSs are stored on 2.5 inch HD in a format which depends on the recorder. After the recovery, the data on each 2.5 inch HD must be copied to an ordinary computer using a data reproduction system. The data reproduction system requires a fast speed for data transfer and a huge capacity disk array. Usually, these features are obtained by using an up-to-date CPU and a disk array with a capacity of more than 1 T bytes with a fast interface (e.g. fiber connection).

For OBS observations, each OBS has its own clock. After the transfer of the data, time adjustments for each OBS record are required. The time difference between each OBS clock and a GPS clock is measured just before deployment and just after recovery. Because the temperature of sea water is constant on the deep sea floor, the crystal oscillator should also be at constant temperature on the seafloor. A linear drift in time to the timing offset of the crystal oscillator is assumed, and used to adjust the time stamp of the record to the correct GPS time. After the time adjustment, each OBS record is combined to construct multi-station files for the whole observation period. During this procedure, multi-station files with a time length of one (or a few) minutes are created for easy handling of the data. Time of multi-station files is continuous so that it is easy to make more longer multi-station file for picking up arrivals.

Because a huge number of files are generated as a result, a powerful computer with a large capacity disk array is also needed for this procedure. After that point, the data processing (e.g. event detection) is the same as for land seismic networks.

#### 7.5.2.6 State of OBS activities

In Japan, the development of OBSs was started at the Geophysical Institute, and ERI of the University of Tokyo in 1970's, and now continues at ERI. These groups developed various types of OBSs for surveys of Earth structure and earthquakes observation (e.g. Mochizuki *et al.*, 2008). A few universities have ST-OBS for seismic surveys and local earthquake observations. At present a few governmental institutes have ST-OBSs and are operating cabled OBSs. JAMSTEC has a few hundred ST-OBSs and a smaller number of LT-OBSs and BBOBSs. Their ST-OBSs are mainly used for seismic surveys (e.g. Kodaira *et al.*, 2004).

The U. S. National Ocean Bottom Seismograph Instrument Pool (OBSIP) has provided instrumentation to support US funded research programs requiring OBSs beginning in 2000. The Lamont-Doherty Earth Observatory at Columbia University, the Scripps Institution of Oceanography at the University of California San Diego and the Woods Hole Oceanographic Institution each have developed and now maintain OBSs for the US OBSIP, including ST-OBSs, LT-OBSs and BBOBSs. The ST-OBSs in the OBSIP can record for about 2 months and use for sensors both geophones and hydrophones. Principal investigators at research universities can request the use of instruments as part of the standard NSF proposal process. Many experiments supported by OBSIPs have been carried out (e.g. Pozgay *et al.*, 2009 for the MARIANA experiment).

The Leibniz Institute of Marine Sciences at Kiel University (IFM-GEOMAR) in Germany has both OBSs and Ocean Bottom Hydrophones (OBHs). OBHs were first developed and then OBSs were developed based on the experiences with the OBH. In 2002 a completely revised design was developed for the OBS system to enable the operation of either short period or broadband seismometers. The recording term of BBOBS is up to one year. In Germany, the Institute for Geophysics at Hamburg University and Alfred Wegener Institute in Bremerhaven have also developed and operate OBSs. These institutes are actively carrying out experiments (e.g. Dahm *et al.*, 2002). The German Instrument Pool for amphibian seismology (DEPAS), which is operated by the Alfred Wegener Institute für Polar und Meeresforschung, Bremerhaven, uses Guralp CMG 40T (60 sec) sensors in titanium pressure cylinders ([http://www.awi.de/en/research/research\\_divisions/geosciences/geophysics](http://www.awi.de/en/research/research_divisions/geosciences/geophysics)). At present, the DEPAS has a large scale facility of broadband OBS (80 instruments) (Friederich and Meier, 2008). Hamburg University and GEOMAR use SEND Geolon MLS data logger with a power consumption of 0.25 W, 21bit resolution at 50 Hz, and time accuracy of 0.5 ms/day.

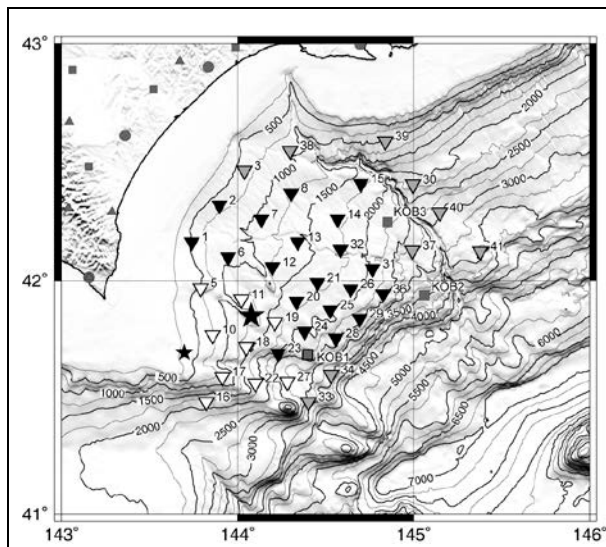
The French Research Institute for Exploitation of the Sea (Ifremer) has approximately 20 ST-OBSs for wide angle seismic surveys. The OBS at Ifremer consists of identical recorders based on the GEOMAR data logger and also on their own mechanical design. The Ocean Bottom Instrumentation Consortium (OBIC) in UK comprises three well established academic institutions: the Universities of Durham and Southampton and Imperial College, London. The OBIC has about fifty LT-OBSs, which are mainly used for seismic structure surveys. In 2008, 50 OBSs recorded airgun signals in the source region of the 2004 Sumatra earthquakes, and 10 OBSs were re-deployed to record earthquakes for approximately one year (Barton *et al.*, 2008).

## 7.5.3 Examples of seafloor seismic observations

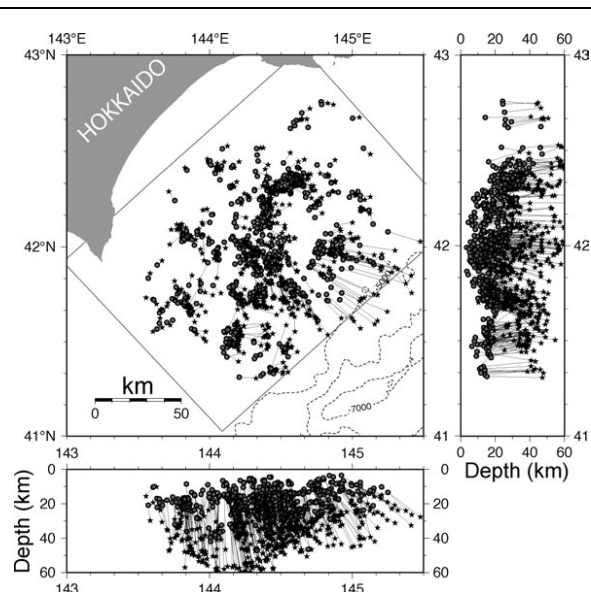
### 7.5.3.1 Local earthquakes

One of the best examples of sea floor observations is that of the aftershocks of large earthquakes occurring in marine areas. On 25 September 2003 the great Mw8.3 Tokachi-oki earthquake occurred off Hokkaido, Japan. Four days after the mainshock, a deployment of twenty-nine OBS in the source region was started to study the aftershock activity of this event using a chartered ship with a size of 700 tons. From October 19-21, nine OBSs that had been deployed near the epicenter of the mainshock were recovered by a research vessel and the data used to obtain the distribution of aftershocks. Ten OBSs were redeployed at the same sites as the previous deployment to continue the aftershock observations. On October 18 and 19, eight OBSs were additionally deployed to enlarge the observation area. Consequently, the aftershocks were observed at forty-one sites including three cable OBS sites, and forty-seven pop-up type OBSs were used in total. All the OBSs at the sea floor were recovered by a chartered ship with a size of 700 tons from November 17 to 20. The observation area was 150 km x 100 km that covered the high aftershock activity, as estimated from the land seismic network. The station spacing of the OBSs was approximately 15 km in the trench region. In the landward area, OBS were deployed with a spacing of about 20 km because the aftershocks were estimated to occur in this region at depths deeper than 20 km (Fig. 7.81).

Two types of the digital recording OBS system were used. The 46 OBSs had both vertical and horizontal velocity sensitive electro-magnetic geophones with a natural frequency of 4.5 Hz and one BBOBS (Guralp CMG-3T). Accurate timing, estimated to be within 0.05 s, was provided by a crystal oscillator. All the OBS were a pop-up type with an acoustic release system. The OBS positions at the sea floor were estimated with an accuracy of several meters using acoustic ranging and ship GPS positions. The water depths at the OBSs were determined by the acoustic ranging and hypocenters were determined by using a three-dimensional velocity structure based on a seismic refraction study. According to the obtained epicentral distribution, the aftershocks occurred within the small slip region of the main shock. In addition, the epicentral distribution of aftershocks, except near the large slip region, was similar to that of the earthquakes occurring before the main shock. Although most land-based hypocenters, which included the data from the cabled OBSs, had depths of greater than 20 km, all hypocenters determined by the OBS network had focal depths around 20 km. P and S wave arrivals at seismic stations above a hypocenters are essential in order to obtain precise hypocenter positions. The OBS network satisfies this condition. All hypocenters located by the OBS network were shallower than those determined by the land network (Fig. 7.82) with most of the aftershocks occurring around 20 km depth. The aftershocks form a dipping plane toward the land. It is inferred that this plane shows the upper boundary of the subducting Pacific Plate. The position of the plate boundary estimated by the aftershock distribution is consistent with estimates by the past seismic surveys (Yamada *et al.*, 2005).



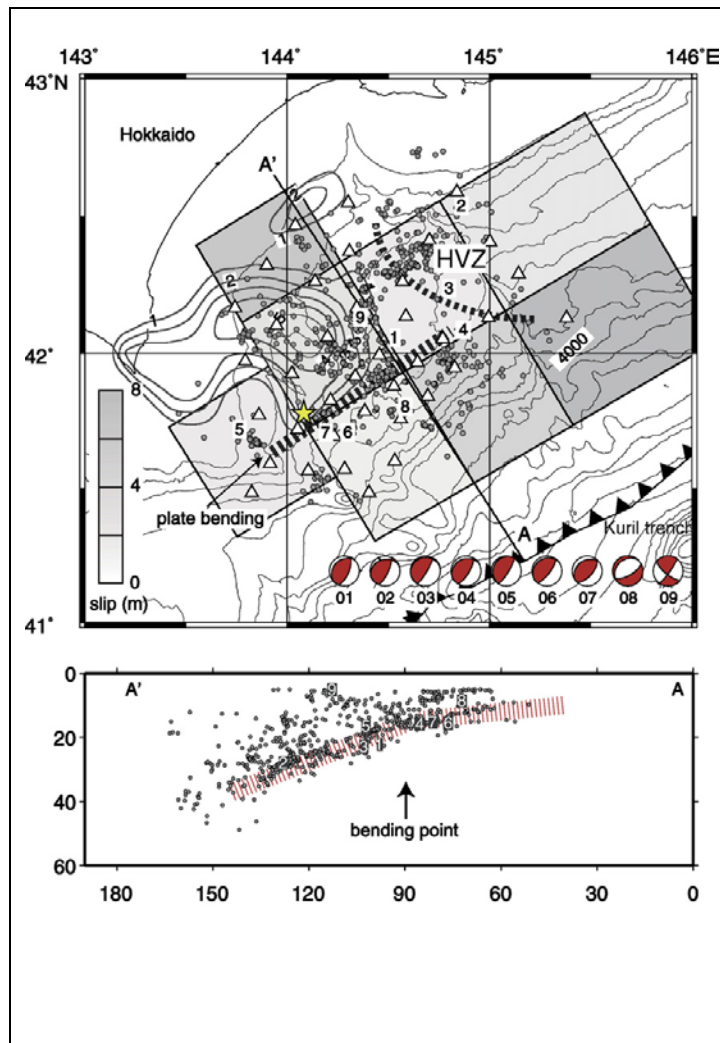
**Fig. 7.81** Bathymetric map off-shore of SE Hokkaido with the positions of OBS used for the aftershock observation of the 2003 Mw8.3 Tokachi-oki earthquake. Inverted triangles indicate the positions of pop-up type OBSs, squares the positions of the cabled OBSs, numerals the site numbers, large and small stars the epicenters of the mainshock and the largest aftershock. Gray symbols in the land region show the positions of the permanent seismic stations (after Shinohara *et al.*, 2004).



**Fig. 7.82** Comparison of the distribution of Tokachi-oki earthquake aftershock epicenter (upper left) and hypocenter positions in N-S and E-W projection (upper right and lower left) as determined by the OBS network (gray circles) and by the dense land network on the Japanese islands. Arrows relate the land network locations to the OBS locations (after Yamada *et al.*, 2005)

Simultaneous estimation of the hypocenters and 3-D seismic velocity models from the P- and S-wave arrivals of the aftershocks recorded by OBSs was performed (Fig. 7.83). The subducting plate is clearly imaged as a northwest dipping zone in which  $V_p$  is higher than 7 km/s. The aftershock distribution reveals that the dip angle of the plate boundary increases around 90 km from the Kuril Trench towards Hokkaido from less than 5 degree to 16 degree. The bending of the subducting plate corresponds to the southeastern edge of the rupture area. The island arc crust on the overriding plate has P-wave velocities of 6–7 km/s and a  $V_p/V_s$  of 1.73. A region of  $V_p/V_s$  greater than 1.88 was found north of the epicenter of the main shock. The depth of the high  $V_p/V_s$  region extends about 10 km upward from the plate interface. The plate boundary just below the high  $V_p/V_s$  region has the largest slip at the main rupture. A high  $V_p$  anomaly ( $\sim 7.5$  km/s) is found in the island arc crust in the northeast part of the study area. The rupture of the main shock terminated at this high  $V_p$  region (Machida *et al.*, 2009).





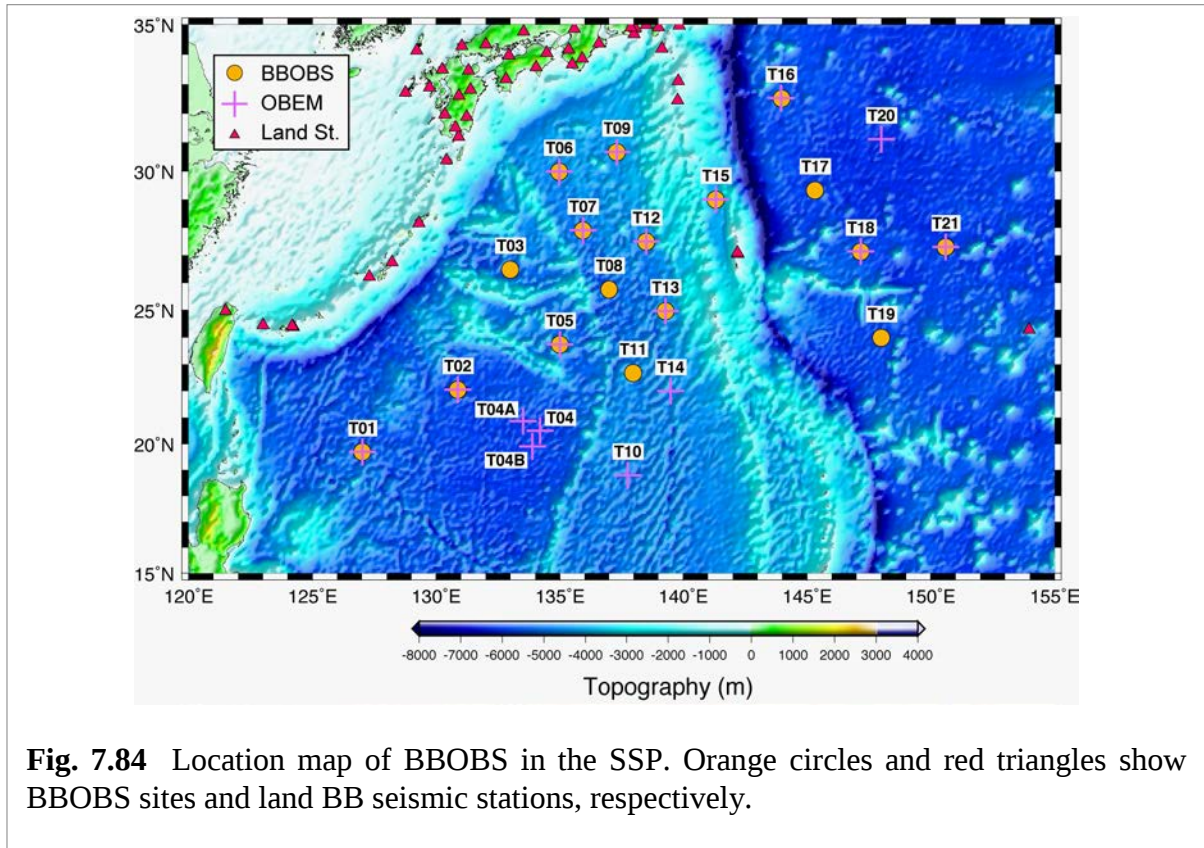
**Fig. 7.83** **Top:** Distribution of 589 relocated aftershocks determined beneath the OBS network. The yellow star and the gray circles represent the epicenters of the 2003 Tokaichi-oki mainshock and its relocated aftershock hypocenters, respectively. Triangles denote positions of the OBSs. Focal mechanisms determined by JMA for the numbered aftershocks are shown at the lower left. Lower hemisphere projection is used. HVZ means the high P-wave velocity zone by the tomography with OBS data. The slip distribution of the main shock after Yamanaka and Kikuchi (2003) is shown by thicker contour lines. The tsunamigenic asperity of 1952 Tokachi-oki earthquake (Hirata *et al.*, 2003) is also indicated (gray regions). **Bottom:** Distribution of the relocated hypocenters projected on the vertical cross section along A-A'. (after Machida *et al.*, 2009)

### 7.5.3.2 Teleseismic events

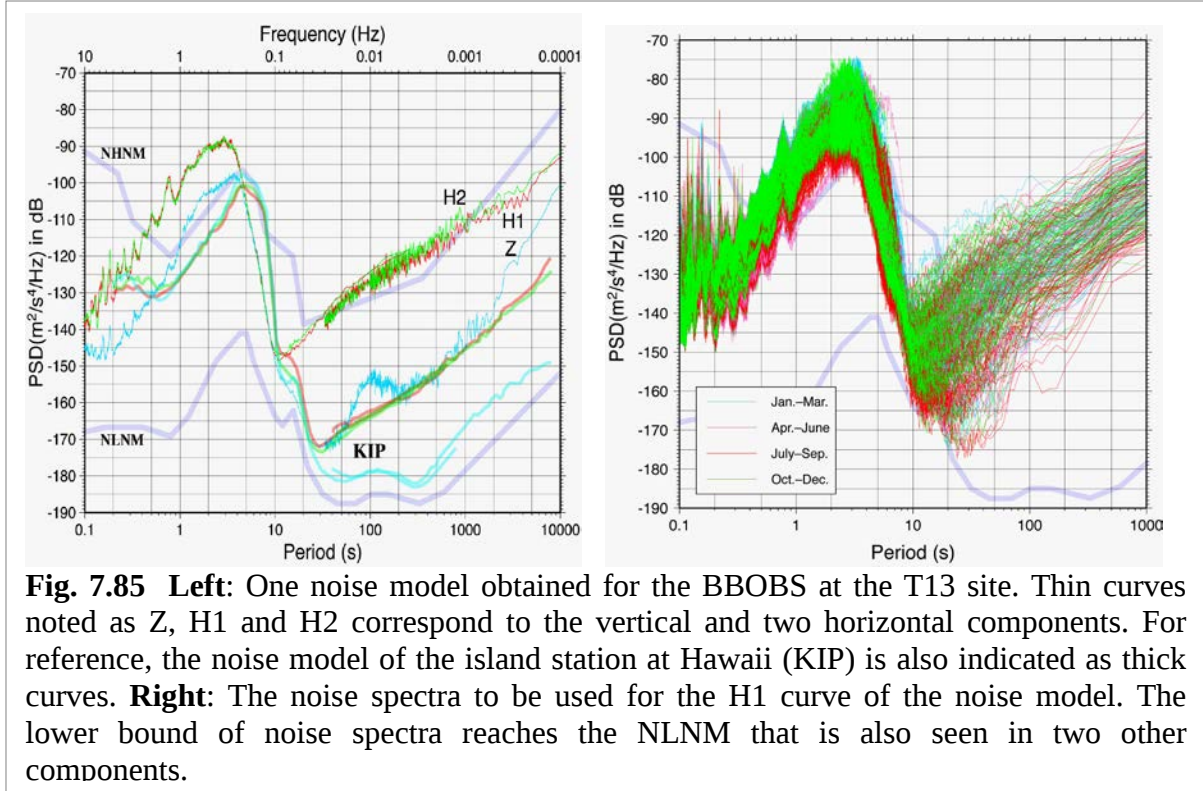
To investigate the stagnant slab beneath the northern Philippine Sea, we have deployed a large OBS array over the 3 years between 2005 and 2008 (Shiobara *et al.*, 2009). This was a key part of the "Stagnant Slab Project" (SSP) started in 2004 and running for 5 years, because these were the first direct and dense observations above this target area aimed at revealing the fine physical structure of the stagnant slab. The transition of the slab morphology along the Izu – Ogasawara (Bonin) – Mariana arc as shown by a global tomography is also a target of interest to be resolved in high resolution by this experiment. The observation area extended about 1000 km (N-S) X 2000 km (E-W), as shown in Fig. 7.84. The first, second and fourth year's cruises were performed by the R/V Kairei (JAMSTEC), and the third year's cruise was done by a chartered ship. The number of BBOBS deployed was 12, 12, 15 for the first, second and third year observation periods, respectively. All of the BBOBS and OBEMs have been recovered successfully except for one BBOBS that did not come up to the sea surface.

From the noise study of all BBOBS stations, all components showed noise levels at periods around 10 s in the middle between the New Low Noise Model (NLNM) and the New High Noise Model (NHNM) according to (Peterson 1993). The vertical noise data were found to lie

in this same middle level at longer periods, but the horizontal noise levels became then higher than the NHNM (see Fig. 7.85 for site T13). The main reason for this high noise level for the horizontal components is assumed to be caused by variations in the BBOBS's tilt due to the sea bottom currents. But, the lower bound of individual noise spectra used for the noise model reaches the NLNM between 10 s and 30 s in the horizontal component that shows the high noise level in the noise model (Fig. 7.85).

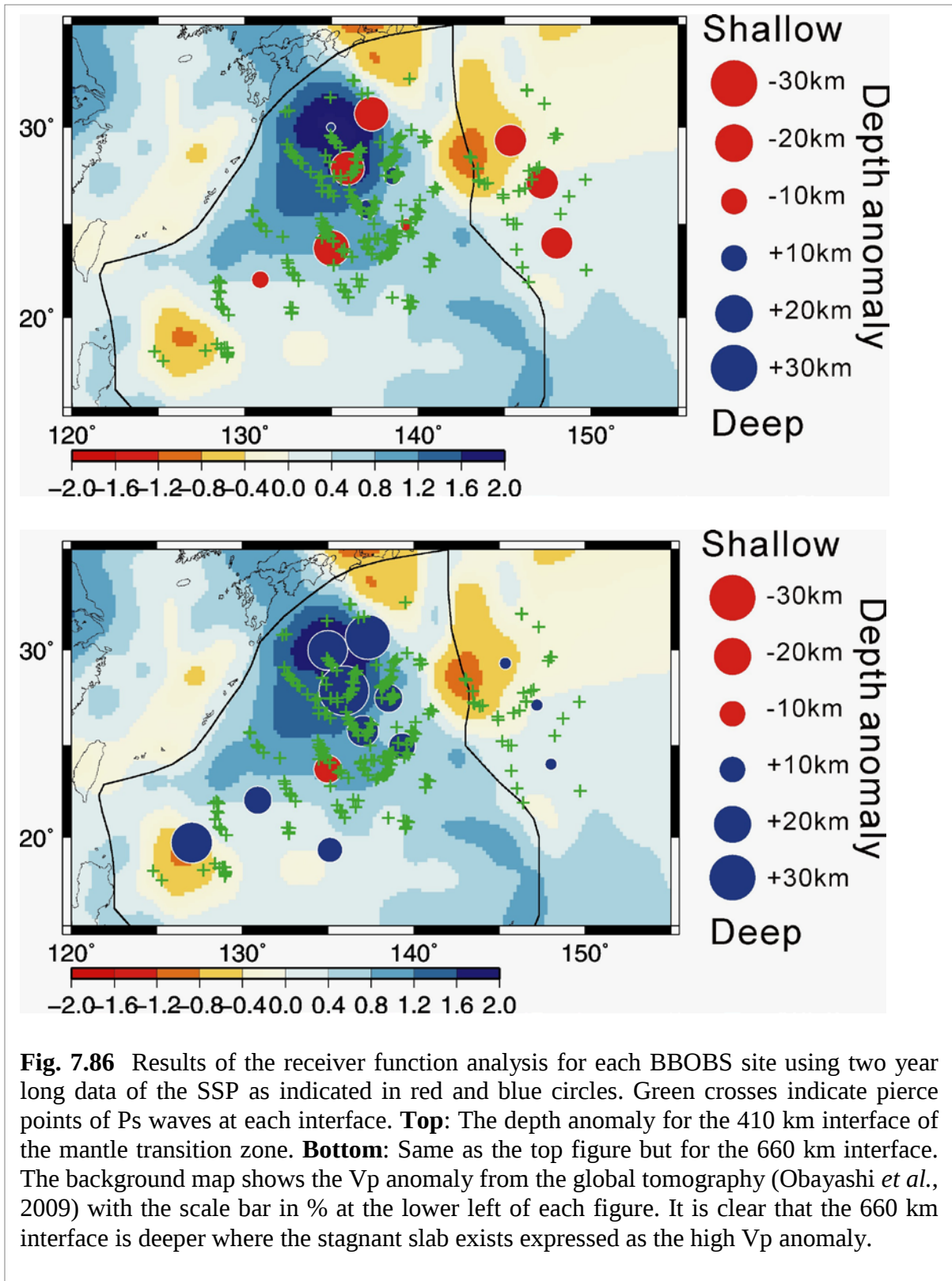


**Fig. 7.84** Location map of BBOBS in the SSP. Orange circles and red triangles show BBOBS sites and land BB seismic stations, respectively.



Results from the dataset from the first and second years of observation include surface wave tomography (Isse *et al.*, 2009) and receiver function analysis (Suetsugu *et al.*, submitted). The former revealed a better resolved upper mantle model for this area than has been seen before. The later study demonstrates clear differences in the mantle transition zone at the edge of the stagnant slab (Fig. 7.86).

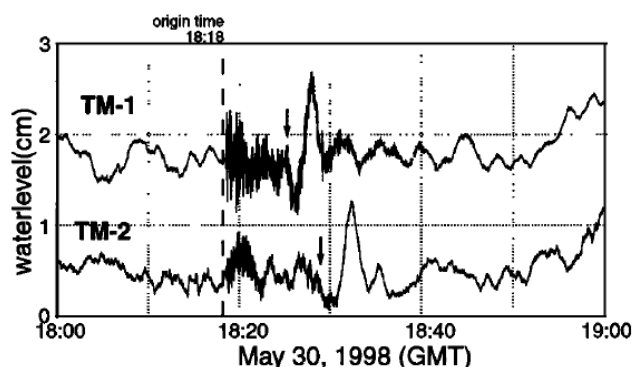




### 7.5.3.3 Tsunami events

Three ocean bottom seismographs (OBSs) and two ocean bottom tsunami-meters (OBTMs) were deployed off Sanriku, northeastern Japan by the Earthquake Research Institute of the University of Tokyo. All data from the OBSs and OBTMs are transmitted to land in real-time. On 30 May 1998, a thrust-type earthquake (Mw 6.1) occurred off the Sanriku region. The

two OBTMs, TM-1 and TM-2, which were installed at depths of 990 and 1563 m, respectively, about 80 and 100 km apart from the hypocenter, detected a small tsunami generated by this earthquake. with a peak-to-peak amplitude of only about 1.5 cm and a wavelet period of about 300 s (Fig. 7.87). The OBTMs had quartz-type pressure sensors with pressure resolution and full scale of 0.5 mm and 3,000 m, respectively, and a sampling interval of 100 ms. Hino *et al.* (2001) estimated reliable tsunami source parameters and showed that ocean bottom tsunami-meters are useful for studying plate subduction.



**Fig. 7.87** Tsunami records obtained by the OBTM stations, TM-1 and TM-2. See Fig.9 for positions of TM-1 and TM-2. Ocean tide components were removed and moving average of 60 s length were applied (after Hino *et al.*, 2001).

## 7.5.4 Technical specification examples

### **Specifications of ST-OBS**

Sensor : Tri-axis 4.5Hz~ (-3dB) (Sercel L-28 for ERI)  
Recorder : 24bit, 100/200Hz sampling, 0.29W in average (HDDR5)  
Data media : Firewire HDD (ex. 40GB, max. 4 drives)  
Clock : Within the recorder, MCXO ( $<\pm 0.1$ ppm, QEM77-AH)  
Power : 6 Li cells (for 90 days obs.), DD size (3.9V, 30AH)  
Transponder : Slant ranging, anchor release control  
Anchor releasing : Forced electric corrosion of two thin Ti/Stainless plates  
Housing : Glass sphere ( $\varnothing$  432mm), 25kgw of buoyancy  
Recovery aids : Radio beacon and xenon flasher (Novatech)  
Dimension : 1.0m/1.0m/0.6m (width/depth/height)  
Weight in the air : 80/40 kgw (with/without anchor)

### **Specifications of LT-OBS**

Sensor : Tri-axis 1~80Hz (-3dB), 400V/m/s, 0.1W (Lennartz LE-3D for ERI)  
Gimbal : Active leveling ( $\leq 20^\circ$ )  
Recorder : 24bit, 100/200Hz sampling, 0.29W in average (HDDR5)  
Data media : Firewire HDD (ex. 40GB, max. 4 drives)  
Clock : Within the recorder, MCXO ( $<\pm 0.1$ ppm, QEM77-AH)  
Power : 50 Li cells (for 356 days obs.), DD size (3.9V, 30AH)  
Transponder : Slant ranging, anchor release/recorder control  
Anchor releasing : forced electric corrosion of two thin Ti plates  
Housing : Titanium sphere ( $\varnothing$  500mm), 33kgw of buoyancy  
Recovery aids : radio beacon and xenon flasher (Novatech)  
Dimension : 1.0m/1.0m/0.7m (width/depth/height)  
Weight in the air : 120/80 kgw (with/without anchor)

### **Specifications of BBOBS**

Sensor : Tri-axis 360s~50Hz (-3dB), 1500V/m/s, 0.28W (Guralp CMG-3T for ERI)  
 Gimbal : Active leveling ( $\leq 20^\circ$ ) with fluxgate compass  
 Recorder : 24bit, 100/200Hz sampling, 0.29W in average (HDDR5)  
 Data media : Firewire HDD (ex. 40GB, max. 4 drives)  
 Clock : Within the recorder, MCXO ( $< \pm 0.1$ ppm, QEM77-AH)  
 Power : 78 Li cells (for 400 days obs.), DD size (3.9V, 30AH)  
 Transponder : Slant ranging, anchor release/recorder control  
 Anchor releasing : Forced electric corrosion of two thin Ti plates  
 Housing : Titanium sphere ( $\varnothing$  650mm), 70kgw of buoyancy  
 Recovery aids : Radio beacon and xenon flasher (Novatech)  
 Dimension : 1.0m/1.0m/0.8m (width/depth/height)  
 Weight in the air : 240/160 kgw (with/without anchor)

### ***Specifications of OBA***

Sensor : Tri-axis DC~ (-3dB),  $0.2\text{V/m/s}^2$ , 0.72W (JAE JA-28GA for ERI)  
 Sensor installation : Fixing to pressure vessel  
 Recorder : 24bit, 100/200Hz sampling, 0.29W in average (HDDR5)  
 Data media : Firewire HDD (ex. 40GB, max. 4 drives)  
 Clock : Within the recorder, MCXO ( $< \pm 0.1$ ppm, QEM77-AH)  
 Power : 120 Li cells (for 365 days obs.), DD size (3.9V, 30AH)  
 Transponder : Slant ranging, anchor release/recorder control  
 Anchor releasing : Forced electric corrosion of two thin Ti plates  
 Housing : Titanium sphere ( $\varnothing$  650mm), 70kgw of buoyancy  
 Recovery aids : Radio beacon and xenon flasher (Novatech)  
 Dimension : 1.0m/1.0m/0.8m (width/depth/height)  
 Weight in the air : 240/160 kgw (with/without anchor)

## **Acknowledgments**

The authors of Chapter 7 are thankful to Ahmed Elgamal, C. R. Hutt, Dieter Mayer-Rosa, Jamison Steidl, Rod Stewart and Peter Zweifel, for their reviews and suggestions which have helped to improve the original text. The non-native English-American contributors to this Chapter owe particular thanks to Rod Stewart for his careful English proof-reading of the sub-chapters 7.1 to 7.4. W. Hanka, author of section 7.4.4, also thanks M. Brunner (†) for many excellent ideas and carrying out the shielding experiments and K.-H. Jaeckel for very fruitful discussions around this topic.

The authors of section 7.5 on OBS wish to express thanks to Toshihiko Kanazawa, retired professor of marine seismology, Earthquake Research Institute of the University of Tokyo. Valuable comments from the editor of the NMSOP, Peter Bormann during two careful reviews and many suggestions made on the first and second draft helped to improve the paper. The authors also thank Torsten Dahm, Azusa Nishizawa and Spahr Webb for their critical reviews and recommendations and are particularly grateful to Spahr Webb for the troubles he has taken in polishing up the English draft.

## **Recommended overview readings**

Havskov and Alguacil (2004 and 2006)  
 Isihara (1996)

## References

- Ad Hoc Group of Scientific Experts to Consider International Cooperative Measures to Detect and Identify Seismic Events (1991). *Sourcebook for International Seismic Data Exchange. Conference Room Paper/167/Rev.2*, Geneva, April 1991.
- Ambraseys; N. N., Simpson, K. A., and Bommer, J. J. (1996). Prediction of horizontal response spectra in Europe. *Earthq. Eng. Struct. Dynam*, 25, 371-400.
- Araki, E., Shinohara, M., Sacks, S., Linde, A., Kanazawa, T., Shiobara, H., Mikada, H., and K. Suyehiro (2004). Improvement of seismic observation in the ocean by use of seafloor boreholes. *Bull. Seism. Soc. Am.*, 94, 678-690.
- Aster, R. C., and Shearer, P. (1991). High-frequency borehole seismograms recorded in the San Jacinto fault zone, Southern California. Part 1. Polarizations. *Bull. Seism. Soc. Am.*, 81, 1057-1080.
- Baumbach, M. (1999). SEIS89 - A PC tool for seismogram analysis. In: Bormann (Ed.) International Training Course 1999 on Seismology, Seismic Hazard Assessment and Risk Mitigation, Lecture and exercise notes, Vol. 1, *GeoForschungsZentrum Potsdam, Scientific Technical Reports STR99/13*, 423-452.
- Bartal, Y., Somer, Z., Leonhard, G., Steinberg, D. M., and Horin, Y. B. (2000). Optimal seismic networks in Israel in the context of the Comprehensive Test Ban Treaty. *Bull. Seism. Soc. Am.*, 90, 1, p 151-165.
- Barton, P. J., Dayuf Jusuf, M., and *et al.* (2008). *RV Sonne Cruise 198-1, 03 May-14 Jun 2008. Singapore - Merak, Indonesia*. Southampton, UK, National Oceanography Centre Southampton, *National Oceanography Centre Southampton Cruise Report*, 31, 100 pp.
- Bormann, P., Wylegally, K., and Klinge, K. (1997). Analysis of broadband seismic noise at the German Regional Seismic Network and search for improved alternative station sites. *J. Seism.*, 1, 357-381.
- Bycroft, G. N. (1978). The effect of soil-structure interaction on seismometer readings, *Bull. Seism. Soc. Am.*, 68, 823.
- Crawford, W. C., Stephen, R. A., and Bolmer, S. T. (2006). A second look at low-frequency marine vertical seismometer data quality at the OSN-1 site off Hawaii for seafloor, buried, and borehole emplacements. *Bull. Seism. Soc. Am.*, 96, 1952-1960.
- Dahm, T., Thorwart, M., Flueh, E. R., Braun, Th., Herber, R., Favali, P., Beranzoli, L., Danna, G., Frugoni, F., and Smiraglio, G. (2002). Ocean Bottom Seismometers Deployed in Tyrrhenian Sea, *EOS Transactions*, 83(29) 309, 314-315.
- Dahm, T., Tilmann, F. and Morgan, J. P. (2006). Seismic broadband ocean-bottom data and noise observed with free-fall stations: Experiences from long-term deployments in the north Atlantic and the Tyrrhenian sea. *Bull. Seism. Soc. Am.*, 96, 647-664.
- Driscoll, F. G. (1986). Groundwater and wells, Second Ed., published by *Johnson Filtration Systems, Inc.*, St. Paul, MN, ISBN 0-9616456-0-1.
- Friedrich, A. (1996). Untersuchung der breitbandigen seismischen Bodenunruhe an GRF- und GRSN-Stationen. Diploma Thesis, Department of Physics, *University of Erlangen-Nürnberg*.
- Hanka, W., and Kind, R. (1994). The GEOFON Program. *Annali di Geofisica*, 37, 5, 1060-1065.
- Hardt, M., and Scherbaum, F. (1994). The design of optimum networks for aftershock recordings. *Geophys. J. Int.*, 117, 716-726.
- Henger, M. (Reporter) (1995). Abschlussbericht Breitbänderfassung seismischen Wellenfeldes im Bereich der Bundesrepublik Deutschland. Phase A: Errichtung und

- Betrieb der seismischen Stationen und des Datezentrums. DFG- Forschungsvorhaben Du 36/9-1,2 und Ba 1276/1-1, *BGR Archiv* Nr. 113 510.
- Hino, R., Tanioka, T., Kanazawa, T., Sakai, S., Nishino, M., and Suyehiro, K. (2001). Micro-tsunami from a local interplate earthquake detected by cabled offshore tsunami observation in northeastern Japan, *Geophys. Res. Lett.*, 28, 3533-3536.
- Hirata, K., Geist, E., Satake, K., Tanioka, Y., and Yamaki, S. (2003). Slip distribution of the 1952 Tokachi-oki earthquake (M 8.1) along the Kuril Trench deduced from tsunami waveform inversion. *J. Geophys. Res.*, 108, 2196. doi:10.1029/2002JB001976.
- Holcomb, G. L., and Hutt, C. R. (1992). An evaluation of installation methods for STS-1 seismometers. *U. S. Geological Survey, Albuquerque, NM, Open File Report* 92-302.
- Ishihara, K. (1996). Soil Behaviour in Earthquake Geotechnics. Oxford Engineering Science Series, No 46, *Clarendon Press*, ISBN: 0198562241.
- Isse, T., Shiobara, H., Tamura, Y., Suetsugu, D., Yoshizawa, K., Sugioka, H., Ito, A., Shinohara, M., Mochizuki, K., Araki, E., Nakahigashi, K., Kawakatsu, H., Shito, A., Kanazawa, T., Fukao, Y., Ishizuka, O., and Gill, J. B. (2009). Seismic structure of the upper mantle beneath the Philippine Sea from seafloor and land observation: implications for mantle convection and magma genesis in the Izu-Bonin-Mariana subduction zone, *Earth Planet. Sci. Lett.*, 278, 107–119; doi:10.1016/j.epsl.2008.11.032.
- Kanazawa, T., Sager, W. W., Escutia, C., and Shipboard Scientific Party (2001). Northwest Pacific seismic observatory and Hammer Drill tests, Sites 1179-1182. *Proc. ODP, Init. Rep.*, 191 (CD-ROM).
- Kanazawa, T., and Shinohara, M. (2009). A new, compact ocean bottom cabled seismometer system - Development of compact cabled seismometers for seafloor observation and a description of first installation plan. *Sea Technology*, July 2009, 37-40.
- Kijko, A. (1977). An algorithm for the optimum distribution of a regional seismic network - I. *Pageoph*, 115, 999-1009.
- Kodaira, S., Iidaka, T., Kato, A., Park, J.-O., Iwasaki, T., and Kaneda, Y. (2004). High pore fluid pressure may cause silent slip in the Nankai Trough. *Science*, 304, 1295–1298.
- Longuet-Higgins, M. D. (1950). A theory on the origin of microseisms. *Philos. Trans. R. Soc. Lond.*, A 243, 1-35.
- Luco, J. E., Anderson, J. G., and Georgevich, M. (1990). Soil-structure interaction effects on strong motion accelerograms recorded on instrument shelters. *Earth. Eng. Struct. Dyn.*, 19, p 119.
- Machida, Y., Shinohara, M., Takanami, T., Murai, Y., Yamada, T., Hirata, N., Suyehiro, K., Kanazawa, T., Kaneda, Y., Mikada, H., Sakai, S., Watanabe, T., Uehira, K., Takahashi, N., Nishino, M., Mochizuki, K., Sato, T., Araki, E., Hino, R., Uhira, K., Shiobara, H., and Shimizu, H. (2009). Heterogeneous structure around the rupture area of the 2003 Tokachi-oki earthquake (Mw=8.0), Japan, as revealed by aftershock observations using Ocean Bottom Seismometers, *Tectonophysics*, 465, 164-176.
- Mochizuki, K., Yamada, T., Shinohara, M., Yamanaka, Y., and Kanazawa, T. (2008). Weak Interplate Coupling by Seamounts and Repeating M~7 Earthquakes. *Science*, 321, 5839, 1194-1197; doi: 10.1126/science.1160250)
- Montagner, J. P., Karczewski, J. F., Romanowicz, B., Bouaricha, S., Lognonné, P., Roult, G., Stutzmann, E., Thiot, J. L., Brion, J., Dole, B., Fouassier, D., Koenig, J.C., Savary, J., Louri, L., Dupond, J., Echardour, A., and Floc'h, H. (1994). The French pilot experiment OFM-SISMOBS: first scientific results on noise level and event detection. *Physics Earth Planet. Inter.*, 84, 321-336.
- Obayashi, M., Yoshimitsu, J., and Fukao, Y. (2009). Tearing of stagnant slab. *Science*, 324, 1173; doi: 10.1126/science.117296.
- Peterson, J. (1993). Observations and modeling of seismic background noise. *U.S. Geol. Survey Open-File Report* 93-322, 95 pp.

- Rabinowitz, N., and Steinberg, D. M. (1990). Optimal configuration of a seismographic network: A statistical approach. *Bull. Seism. Soc. Am.*, 80, 1, 187-196.
- Sacks, I. S., Suyehiro, K., Acton, G.D., and Shipboard Scientific Party (2000). Western Pacific Geophysical Observatories, Sites 1150 and 1151, *Proc. ODP, Init. Rep.*, 186, (CD-ROM).
- Salisbury, M.H., Shinohara, M., Richter, C., and Shipboard Scientific Party (2002). Seafloor observatories and the Kuroshio current. *Proc. ODP, Init. Rep.*, 195, (CD-ROM).
- Shinohara, M., Hirata, N., and Takahashi, N. (1994). High resolution velocity analysis of ocean bottom seismometer data by the  $\tau$ -p method. *Marine Geophys. Res.*, 16, 185-199.
- Shinohara, M., Yamada, T., Kanazawa, T., Hirata, N., Kaneda, Y., Takanami, T., Mikada, H., Suyehiro, K., Sakai, S., Watanabe, T., Uehira, K., Murai, Y., Takahashi, N., Nishino, M., Mochizuki, K., Sato, T., Araki, E., Hino, R., Uhira, K., Shiobara, H., and Shimizu, H. (2004). Aftershock observation of the 2003 Tokachi-oki earthquake by using dense ocean bottom seismometer network. *Earth Planets Space*, 56, 295-300.
- Shinohara, M., Araki, E., Kanazawa, T., Suyehiro, K., Mochizuki, M., Yamada, T., Nakahigashi, K., Kaiho, Y., and Fukao, Y. (2006). Deep-sea borehole seismological observatories in the western Pacific: temporal variation of seismic noise level and event detection. *Annals of Geophysics*, 49, 2/3, 625-641.
- Shinohara, M., Suyehiro, K., Matsuda, S., and Ozawa, K. (1993). Digital recording ocean bottom seismometer using portable digital audio type recorder. *J. Japan Soc. Mar. Surv. Tech.*, 5, 21-31 (in Japanese).
- Shinohara, M., Yamada, T., and Kanazawa, T. (2009). Development of ocean bottom accelerometer for observation of strong motion on sea floor. *J. Japan Soc. Mar. Survey Tech.*, 21 (2), 15-24 (in Japanese with English abstract).
- Shiobara, H., and Kanazawa, T. (2009). Development of a light weight and autonomic sensor system for ocean bottom seismometer. *Zisin* 2, 61 (3), 137-144 (in Japanese).
- Shiobara, H., Baba, K., Utada, H., and Fukao, Y. (2009). Ocean bottom array probes stagnant slab beneath the Philippine Sea. *Eos, Transactions AGU*, 90(9), 70-71.
- Steinberg, D. M., Rabinowitz, N., Shimshoni, Y., and Mizrachi, D. (1995). Configuring a seismograph network for optimal monitoring of fault lines and multiple sources. *Bull. Seism. Soc. Am.*, 85, 6, 1847-1857.
- Suetsugu, D., Inoue, T., Obayashi, M., Yamada, A., Shiobara, H., Sugioka, H., Ito, A., Kanazawa, T., Kawakatsu, H., Shito, A., and Fukao, Y. (2011). Depths of the 410-km and 660-km discontinuities in and around the stagnant slab beneath the Philippine Sea: Is water stored in the stagnant slab ?, Submitted to *Phys. Earth Planet. Inter.*
- Suyehiro, K., and Mochizuki, K. (2002). Marine seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam*, 421-436.
- Trnkoczy, A., and Živčić, M. (1992). Design of Local Seismic Network for Nuclear Power Plant Krsko, *Cahiers du Centre Europeen de Geodynamique et de Seismologie*, Luxembourg, 5, 31-41.
- Uhrhammer, R. A., Karavas, W., and Romanovicz, B. (1998). Broadband Seismic Station Installation Guidelines, *Seism. Res. Lett.*, 69, 15-26.
- Webb, S. C. (1988). Long-period acoustic and seismic measurements and ocean floor currents. *IEEE J. Oceanic Eng.*, 13, no. 4, 263-270.
- Webb S. C. (1998). Broadband seismology and noise under the ocean. *Rev. Geophys.* 36, 105-142.
- Webb, S. C. (2002). Seismic noise on land and on the sea floor. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam*, 305-318.

- Wielandt, E. (1990). Very- broad-band seismometry. In: Boschi, E., Giardini, D., Morelli, A. (Eds.), *Proceed. 1<sup>st</sup> Workshop on MEDNET*, Sept. 10-14, 1990, CCSEM, Erice, Il Cigno Galileo Galilei, Roma, 222-234.
- Wielandt, E., and Streckeisen, G. (1982). The leaf-spring seismometer: Design and performance. *Bull. Seism. Soc. Am.*, 72, 2349-2367.
- Wielandt, E., and Steim, J. M. (1986). A digital very-broad-band seismograph. *Annales Geophysicae*, 4, 227-232.
- Wielandt, E., and Zürn, W. (1991). Messungen der kurzperiodischen Bodenunruhe in Schiltach (BFO). In: Henger (1995).
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report SE-20, September 1979, Boulder, Colorado, 165 pp.
- Yamada, T., Shinohara, M., Kanazawa, T., Hirata, N., Kaneda, Y., Takanami, T., Mikada, H., Suyehiro, K., Sakai, S., Watanabe, T., Uehira, K., Murai, Y., Takahashi, N., Nishino, M., Mochizuki, K., Sato, T., Araki, E., Hino, R., Uhira, K., Shiobara, H., Shimizu, H., (2005). Aftershock distribution of the 2003 Tokachi-oki earthquake derived from highdense network of ocean bottom seismographs (in Japanese), *Zisin* 2, 57, 281–290 (in Japanese).
- Yamanaka, Y., and Kikuchi, M. (2003). Source processes of the recurrent Tokachi-oki earthquake on September 26, inferred from teleseismic body waves, *Earth Planets Space*, 55, e21-e24.

| Topic   | What to prepare and provide if seismic site selection is purchased?                                                          |
|---------|------------------------------------------------------------------------------------------------------------------------------|
| Author  | Amadej Trnkoczy (formerly Kinometrics SA);<br>E-mail: <a href="mailto:amadej.trnkoczy@siol.net">amadej.trnkoczy@siol.net</a> |
| Version | Sept. 1999; DOI: 10.2312/GFZ.NMSOP-2_IS_7.1                                                                                  |

If seismic station site selection procedure is purchased as a part of services along with the seismic network equipment, the purchaser should prepare several logistic things to assure efficient work of the manufacturer's experts. Note that these services are usually paid by the time the experts work on site selection for a new seismic network and that site selection is a 'stretchable' process. The more time (read "money") one spends on it, or the more efficiently one works during a given time period, the better the station sites and, consequently, the network performances will result. Therefore, it is of direct benefit to the customer to consider carefully all the required issues and to get together all the necessary information and working material, as complete as possible under the given conditions in the particular country. The seismic network purchaser should prepare the following:

- a preliminary and approximate proposal of seismic network layout based on the goals of the network;
- a general-purpose "high school type" topographical map of the whole region of the future network with color representation of terrain altitude (basic topographical display of the region);
- regional (and local, if available) geological maps covering the region of the network;
- map of past seismic activity in and around the region where the network is planned, with instrumental (if any) and historic data be included;
- seismo-tectonic map of the region (if available);
- 1:50.000 or 1:25.000 scale topographic maps covering the entire network region for RF profiling purposes for telemetry seismic systems (1:50.000 scale maps are the best; 1:25.000 maps are better for fieldwork if there is no RF telemetry planned in the network). Get permission to export such maps if they are under export restriction, as these will be needed by the site selection provider for initial studies before fieldwork starts, particularly if the network is an RF telemetry system;
- a state-of-the-art roadmap of the country for finding easy access to potential sites during fieldwork. Try to find the latest edition of such map. Road infrastructure changes fast in many developing countries;
- 1:5.000 scale maps (or at least 1:25.000 if 1:5.000 are not available) of the area surrounding the sites in case shallow seismic profiling of potential seismic sites is planned;



- climatic data in the form of maps or tables published in annual or decade reports from the country's meteorological survey (data should include precipitation, wind, insolation - if seismic stations will be powered by solar panels -, and lightning threat information such as isokeraunic maps or number of storm days per year).
- knowledgeable staff members from the institution that will operate the network as well as well informed local people acquainted with local conditions at each potential station site. The member(s) of the responsible institution working in the field, together with manufacturer's experts, should have full competency to make 'on the spot' decisions regarding acceptability of access difficulties, land ownership issues, and other issues that may have financial consequences during network establishment and future network operation. This person should be full time and continuously with the manufacturer's experts until the site selection procedure is finished. If the region of the network is large, several local people may be needed. They can be members of local authorities (municipalities, land-use planning authorities, etc.) and should be familiar with local development conditions and present and future land use;
- one or two four-wheel-drive vehicles in technically perfect condition, one of which should be big enough to comfortably transport four people together with measuring equipment its original packing (two PC notebooks, seismometer, seismic recorder, cables and, in case of telemetry system, RF spectrum analyzer, provisory antenna mast, and Yagi antennae). Two or three customer's staff members (plus driver and enough cash, coupons or whatever documents are required to purchase gasoline) are the best size team to work with usually two manufacturer's experts;
- air-conditioned working room with three tables, main power, and safe storage place for measuring equipment. If the network is an RF telemetry system, one of the tables must be large enough, minimum 1.5 x 3 m (5 x 10 feet), to allow working with several topographical maps stuck together while taking topographical profiles; and
- permits to enter restricted areas (army camps and training land, private land, natural reserves, state border regions, etc.) for local staff and foreign experts.

The maps sent to the site selection provider and used in the field are working copies. They are normally not returned to the customer. The maps are used when preparing the final report. If color maps are code protected against copying, two copies are needed (one for fieldwork and one for the final report).

Expect from one to three days of work for each station site of the network. Any extra time needed will depend on the dimensions of the network, infrastructure in the country, and general site accessibility. An efficient day of fieldwork usually lasts from sunrise to sunset.

**Hint:** Print this form and put check marks in appropriate bullets while preparing on-site selection procedure.

| Topic   | Using existing communication tower sites as seismic sites                                                                    |
|---------|------------------------------------------------------------------------------------------------------------------------------|
| Author  | Amadej Trnkoczy (formerly Kinemetrics SA);<br>E-mail: <a href="mailto:amadej.trnkoczy@siol.net">amadej.trnkoczy@siol.net</a> |
| Version | Sept. 1999; DOI: 10.2312/GFZ.NMSOP-2_IS_7.2                                                                                  |

Very often less experienced newcomers in seismometry consider mountain peaks with existing communication towers as potential seismic station sites, particularly if they are building an radio frequency (RF) telemetry seismic network. Such places appear to be an easy and inexpensive solution. Access problem is solved, RF communication paths to the central recording site, which is usually situated in the capital or another big city, is supposedly free, main power lines, and even phone lines are readily available.

Unfortunately, such sites also have several serious drawbacks and are in fact rarely suitable for seismic stations. The most important reasons are that:

- existing high towers that sway during windy periods cause high-amplitude, low-frequency seismic noise and may cause large numbers of false triggers with triggered seismic systems and deteriorate low frequency seismic signals. Consequently, a diminished seismic station gain is used resulting in a low detectability of the station;
- there is usually a very high probability of RF interference between seismic RF telemetry system and other users. RF interference may easily impair seismic data transmission and consequently seismic system reliability. Several ‘high power’ parties (compared to one watt or less of RF power used in seismic telemetry) are potentially polluting the RF space at such places. In addition, if other users do not maintain their RF equipment properly, the RF energy radiated within uncontrolled side lobes worsens this danger (this happens quite frequently in developing countries.
- if such sites are inhabited, it is likely there will be too high man-made seismic noise due to human activities);
- the topography of such mountain peaks is rarely suitable for a seismic station. Communication antennae towers usually try to cover an area as large as possible, therefore, as a rule, they are placed on the highest mountains in a country or region;
- nearly all such sites have powerful diesel generators to support communication equipment during power outages. When in operation, these generators are a major source of man-made, high frequency seismic noise. Of course, these generators will surely be running after a strong earthquake because that is precisely when it is most likely that the main power lines will fail. Since the periods during strong earthquakes and following aftershock sequence are the most important for the seismic network, the existing communication towers definitely are not at all suitable for seismic sites.

| Topic   | Recommended minimal distances of seismic sites from sources of seismic noise                                                 |
|---------|------------------------------------------------------------------------------------------------------------------------------|
| Author  | Amadej Trnkoczy (formerly Kinemetrics SA);<br>E-mail: <a href="mailto:amadej.trnkoczy@siol.net">amadej.trnkoczy@siol.net</a> |
| Version | Sept. 1999; DOI: 10.2312/GFZ.NMSOP-2_IS_7.3                                                                                  |

Recommended minimal distances from sources of seismic noise to a seismic site (according to Willmore, 1979) are:

|                                                                                                      |   |                                                   |      |                                                                             |      |                    |      |      |
|------------------------------------------------------------------------------------------------------|---|---------------------------------------------------|------|-----------------------------------------------------------------------------|------|--------------------|------|------|
| STATION SITE NAME: _____<br><br>COORDINATES:<br><br>N    0    '    " _____<br>W    0    '    " _____ |   | SITE #: _____                                     |      | DATE OF ANALYSIS:<br>____/____/____<br><br>DATE OF VISIT:<br>____/____/____ |      | ACTUAL<br>DISTANCE |      |      |
|                                                                                                      |   | HARD MASSIVE<br>ROCK, GRANITE,<br>QUARTZITE, ETC. |      | HARDPAN<br>HARD CLAY, ETC.                                                  |      |                    |      |      |
|                                                                                                      |   | RECOMMENDED MINIMAL DISTANCES<br>[km]             |      |                                                                             |      |                    |      |      |
|                                                                                                      |   | A                                                 | B    | C                                                                           | A    | B                  | C    | [km] |
| 1. Oceans, with coastal mountains system                                                             |   | 300                                               | 50   | 1                                                                           | 300  | 50                 | 1    |      |
| 2. Oceans, with broad coastal plains                                                                 |   | 1000                                              | 200  | 10                                                                          | 1000 | 200                | 20   |      |
| 3. Inland seas, bays, very large lakes, with coastal mountain system                                 |   | 150                                               | 25   | 1                                                                           | 150  | 25                 | 1    |      |
| 4. Inland seas, bays, very large lakes, with broad coastal plains                                    |   | 500                                               | 100  | 5                                                                           | 500  | 100                | 5    |      |
| 5. Large dams, high waterfalls, large cataracts                                                      | a | 40                                                | 10   | 1                                                                           | 50   | 15                 | 5    |      |
|                                                                                                      | b | 60                                                | 15   | 5                                                                           | 150  | 25                 | 10   |      |
| 6. Large oil or gas pipelines                                                                        | a | 20                                                | 10   | 5                                                                           | 30   | 15                 | 5    |      |
|                                                                                                      | b | 100                                               | 30   | 10                                                                          | 100  | 30                 | 10   |      |
| 7. Small lakes                                                                                       | a | 20                                                | 10   | 1                                                                           | 20   | 10                 | 1    |      |
|                                                                                                      | b | 50                                                | 15   | 1                                                                           | 50   | 15                 | 1    |      |
| 8. Heavy reciprocating machinery, machinery                                                          | a | 15                                                | 3    | 1                                                                           | 20   | 5                  | 2    |      |
|                                                                                                      | b | 25                                                | 5    | 2                                                                           | 40   | 15                 | 3    |      |
| 9. Low waterfalls, rapids of a large river, intermittent flow over large dams                        | a | 5                                                 | 2    | 0.5                                                                         | 15   | 5                  | 1    |      |
|                                                                                                      | b | 15                                                | 3    | 1                                                                           | 25   | 8                  | 2    |      |
| 10. Railway, frequent operation                                                                      | a | 6                                                 | 3    | 1                                                                           | 10   | 5                  | 1    |      |
|                                                                                                      | b | 15                                                | 5    | 1                                                                           | 20   | 10                 | 1    |      |
| 11. Airport, air ways heavy traffic                                                                  |   | 6                                                 | 3    | 1                                                                           | 6    | 3                  | 1    |      |
| 12. Non-reciprocating power plant machinery, balanced industrial machinery                           | a | 2                                                 | 0.5  | 0.1                                                                         | 10   | 4                  | 1    |      |
|                                                                                                      | b | 4                                                 | 1    | 0.2                                                                         | 15   | 6                  | 1    |      |
| 13. Busy highway, mechanized farms                                                                   |   | 1                                                 | 0.3  | 0.1                                                                         | 6    | 1                  | 0.5  |      |
| 14. Country roads, high buildings                                                                    |   | 0.3                                               | 0.2  | 0.05                                                                        | 2    | 1                  | 0.5  |      |
| 15. Low buildings, high trees and masts                                                              |   | 0.1                                               | 0.03 | 0.01                                                                        | 0.3  | 0.1                | 0.05 |      |
| 16. High fences, low trees, high bushes, large rocks                                                 |   | 0.05                                              | 0.03 | 5 m                                                                         | 0.06 | 0.03               | 0.01 |      |

**LEGEND:**

- A SP seismic station with a gain of about 200,000 or more at 1 Hz
- B SP seismic station with a gain from 50,000 to 150,000 at 1 Hz
- C SP seismic station with a gain of approximately 25,000 or less at 1 Hz
- a Source and seismometer on widely different geological formations or that mountain ranges or valleys intervene
- b Source and seismometer on the same geological formation and with no intervening alluvial valley or mountain range

**Instructions for use of the form:**

1. Get the information about all potential sources of seismic noise around the site and write the distances to them in the extreme right column of the table.
2. From geological maps and by visiting the site decide on the quality of the bedrock at the site. Decide either for 'good' rock (left three columns A, B, and C with minimal recommended distances) or for 'less suitable' ground (right three columns A, B, and C with the minimal recommended distances).
3. For each seismic noise source (where applicable) decide about seismic coupling between seismic site and the noise source. Select the appropriate horizontal line a) or b) with minimal recommended distance.
4. Mark appropriate cells in the table based on the steps #2 and #3 and compare their content with the actual distances in the extreme right column.
5. Shade all cells of the selected A, B, and C columns where the recommended minimal distances to a noise source is bigger than the actual distance in the extreme right column. Find that of the columns A, B, or C where no shaded cells appear. If this is the column A, the site is appropriate for a sensitive SP station having gain 200,000 or more, if this is column B the site is appropriate for a medium sensitive station having the gain somewhere in between 50.000 and 150.000, if it is column C, only a moderately sensitive station with gain around 25.000 or less can be established.
6. Make such a table for all potential seismic sites studied and compare the results among alternatives.

| Topic   | Detectability and earthquake location accuracy modeling of seismic networks                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|---------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p><b>Mladen Živčić</b>, Environmental Agency of the Republic of Slovenia, Seismology and Geology Office, Dunajska 47/VII, SI-1000 Ljubljana, Slovenia; Tel: +386 1 4787270, Fax: +386 1 4787295, E-mail: <a href="mailto:mladen.zivcic@gov.si">mladen.zivcic@gov.si</a></p> <p><b>Jure Ravnik</b>, Faculty of Mechanical Engineering, University of Maribor, Smetanova 17, SI-2000 Maribor, Slovenia; Tel: + 386 2 2207745, Fax: +386 22207996, E-mail: <a href="mailto:jure.ravnik@uni-mb.si">jure.ravnik@uni-mb.si</a></p> |
| Version | September 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_7.4                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |

This information sheet focuses on the accuracy of determination of earthquake hypocenter location with respect to the locations of seismic stations in a local network and the estimation of network detection thresholds.

Contemporary methods of determination of hypocenter location and earthquake origin time are based on modeling the travel time, which is needed for seismic waves to travel from the hypocenter to the station of a seismic network. For this we need to know:

- location and height above sea level of the seismic stations;
- accurate time on all seismic stations;
- velocity structure of the Earth, through which the seismic waves propagate.

If these parameters are known, we can calculate by means of numerical methods the theoretical travel time of seismic waves from an arbitrary hypocenter to the seismic station. The calculated travel times are then applied to the actual arrival times which were picked from seismograms on all available seismic stations and thus the hypocenter location and earthquake origin time is calculated.

The accuracy of such earthquake locations depends on the three points listed above and on the accuracy of phase picking. Additionally, the theoretical accuracy of hypocenter locations is also controlled by the spatial distribution of seismic stations. Nowadays, with GPS receivers being readily available at reasonable cost, it is not difficult to know the station location and the correct time exactly. The velocity structure of the Earth is fixed, however, and often enough not well known. By studying it, the accuracy of location can be improved. Unfortunately, the determination of the velocity structure requires either extensive specialized deep seismic refraction surveys or an already operating and sufficiently dense seismic network.

Therefore, in the phase of seismic network planning, we can improve its accuracy of event location only by reasonable distribution of the stations.

A computer program LOK has been developed which estimates the accuracy of hypocenter location based on a given spatial distribution of stations. The following assumptions are made:

- station co-ordinates are known exactly;

- for locations of seismic stations an RMS value of noise in the frequency band within which the STA/LTA trigger algorithm will operate (for digital stations) or for the frequency band of the recording equipment (for analogue stations) is known;
- for both digital and analogue stations the frequency response of the seismographs is flat and proportional to ground velocity in the frequency band of interest for modeling the station and network capabilities. This bandwidth depends on the task but also on the network geometry and sensitivity. For local networks it is usually in the range between 1 and 10 Hz;
- P and S arrival times are picked with a known a priori uncertainty (e.g., 0.1 s);
- P and S velocities within the layers and the positions of layer boundaries are known with some known uncertainty;
- travel times are computed for a flat Earth model consisting of homogeneous layers;
- the size of the network area is such that flat Earth approximation can be used.

The results obtained with LOK crucially depend on these assumptions. However, even with poor choice of input parameters (e.g. velocity model) one can get relative performance of different network geometry.

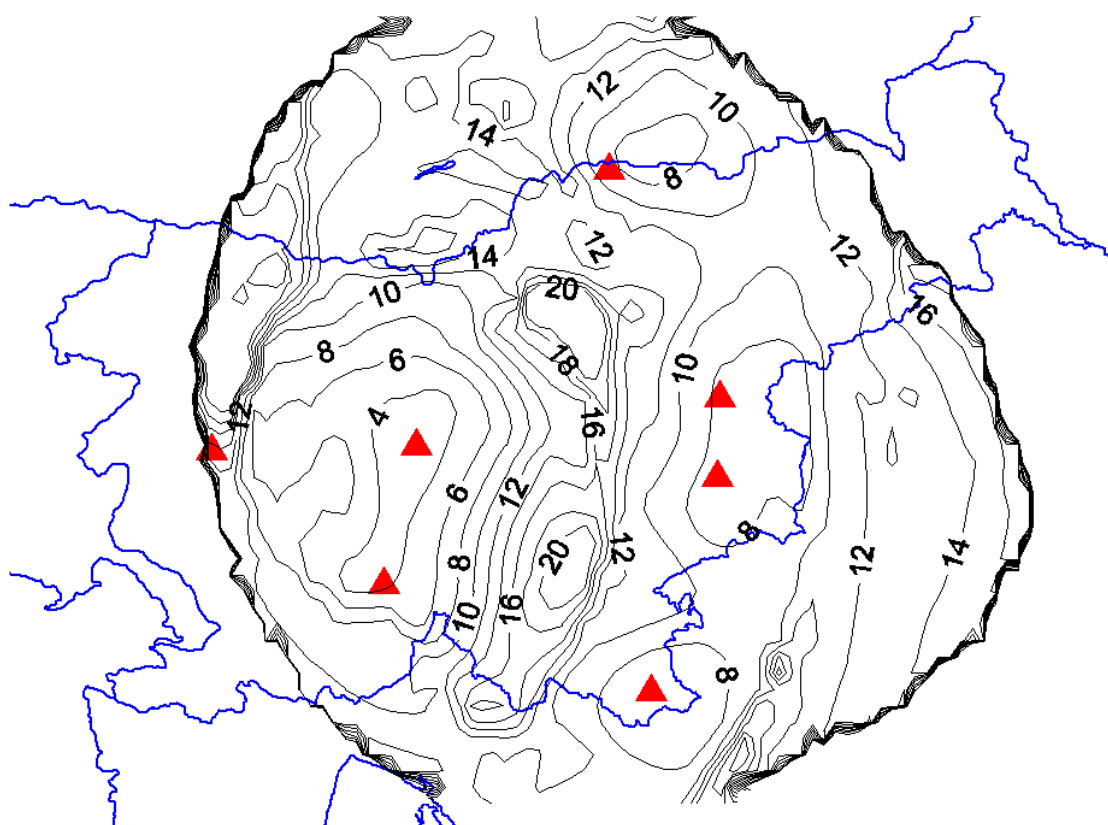
In developing the program LOK we followed mainly the method described in Peters and Crosson (1972). It uses the fact, that the errors of the travel time solution depend on partial derivatives of the travel time function by the unknowns we are looking for. These unknowns are the hypocenter location and the earthquake origin time. The derivatives can be calculated for every point within the seismic network. The area is divided into squares in terms of longitude and latitude. When LOK is run, a hypocenter error ellipsoid is constructed for every grid point. The largest semi-axis of the error ellipsoid is named the hypocenter determination error while the largest of the projections of the ellipsoid semi-axes on the horizontal plane is named the epicentre determination error.

For the computation of the error ellipsoid one should include only stations on which the expected signal is above the noise threshold as defined in the station file (amplification for analogue stations and RMS noise values and STA/LTA trigger ratio for digital stations). Thus the program also gives some information on the differences in expected detectability of events for different geometries of the network. Absolute level of detectability is impossible to predict without detailed knowledge of the attenuation in the region.

Figure 1 below shows the results of respective model calculation for the Stareslo network in Slovenia. The input and output files for this example are included in the distributed version of LOK (see below). An area of  $3.5 \times 1.75$  degree was modeled. An earthquake of  $M_l = 1.0$  was assumed to occur at 15 km hypocentral depth. The network consists of 7 stations, denoted by red triangles in the figure. The border of Slovenia is shown in thick blue. The thin black lines, which are the actual result of the modeling, are isolines of constant hypocenter location error. The numbers in the labels are in kilometers.

As one can see, the error increases outside the network, and the network also has a few blind spots within, where the hypocenter determination error is rather large. The detectability of the network for earthquakes of  $M_l = 1.0$  (at least 4 stations must record the event to obtain the earthquake location) can also be seen. Other examples of magnitude threshold as well as epicenter and hypocenter error calculations using an earlier version of LOK, are shown in Figs. 7.6 to 7.8. of Chapter 7.

The software enables calculation with different hypocenter depths and earthquake magnitudes. It also includes a routine that determines the stations that recorded a particular event. LOK was written in FORTRAN 77 and tested under Linux. The original LOK code has recently been upgraded. A Makefile has been included, written for Intel Fortran Compiler. Users invoke the “make” command in the source directory and LOK is compiled and built. If, however, another compiler is used, the Makefile has to be modified accordingly. Yet, variables were defined in a way that the code does no longer rely on implicit Fortran definition of the variables. For downloading the source code see on the NMSOP-2 cover page in the **Overview** listing *Download Programs & Files* for **lok-sept2011.tar** and related general instructions.



**Figure 1** Result of model calculations for the Stareslo network in Slovenia for  $M_l = 1.0$  earthquake. Red triangles: station positions, blue lines: borders of Slovenia, thin black lines: isolines of hypocenter location error in km; thick black outer boundary: outer limit of the network's location capability for earthquakes of  $M_l = 1.0$ .

## Reference

Peters, D. C. and Crosson, R. S. (1972). Application of prediction analysis to hypocentre determination using local array. *Bull. Seism. Soc. Am.*, **62**(3), 775-788.

# Chapter 8

## Seismic Networks

(Version August 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch8)

Jens Havskov<sup>1)</sup>, Lars Ottemöller<sup>1)</sup>, Amadej Trnkoczy<sup>2)</sup>, and Peter Bormann<sup>3)</sup>

<sup>1)</sup> University of Bergen, Department of Earth Science, Allégaten 41, N-5007 Bergen, Norway  
E-mail: [Jens.Havskov@geo.uib.no](mailto:Jens.Havskov@geo.uib.no); [Lars.Ottemoller@geo.uib.no](mailto:Lars.Ottemoller@geo.uib.no);

<sup>2)</sup> Formerly Kinematics; Bovec, Slovenia; E-mail: [amadej.trnkoczy@siol.net](mailto:amadej.trnkoczy@siol.net)

<sup>3)</sup> Formerly GFZ German Research Center for Geosciences, Telegrafenberg, 14473 Potsdam, Germany  
E-mail: [pb65@gmx.net](mailto:pb65@gmx.net)

|                                                                  | <b>page</b> |
|------------------------------------------------------------------|-------------|
| <b>8.1 Introduction</b>                                          | 3           |
| <b>8.2 Seismic network purpose</b>                               | 3           |
| <b>8.3 Seismic sensors</b>                                       | 5           |
| 8.3.1 General considerations                                     | 5           |
| 8.3.2 Seismometers and/or accelerometers?                        | 5           |
| 8.3.3 One- and three-component seismic stations                  | 6           |
| 8.3.4 Sensitivity of seismic sensors                             | 6           |
| 8.3.5 Frequency range of seismic sensors                         | 7           |
| 8.3.6 Short-period (SP) seismometers                             | 8           |
| 8.3.7 Broadband (BB) seismometers                                | 9           |
| 8.3.8 Very broadband (VBB) seismometers                          | 9           |
| 8.3.9 Long-period (LP) passive seismometers                      | 10          |
| <b>8.4 Seismic network configuration</b>                         | 10          |
| 8.4.1 Real and virtual seismic networks                          | 10          |
| 8.4.2 Seismic networks                                           | 11          |
| 8.4.2.1 Stand alone and central-recording                        | 11          |
| 8.4.2.2 Examples                                                 | 12          |
| <b>8.5 Seismic data acquisition and processing</b>               | 15          |
| 8.5.1 Digital versus analog data acquisition                     | 15          |
| 8.5.1.1 Analog seismic systems                                   | 15          |
| 8.5.1.2 Mixed analog/digital systems                             | 15          |
| 8.5.1.3 Digital seismic systems                                  | 16          |
| 8.5.2 Trigger algorithms and their implementation                | 17          |
| 8.5.2.1 Continuous versus triggered mode of data acquisition     | 17          |
| 8.5.2.2 Trigger algorithms                                       | 18          |
| 8.5.2.3 Coincidence trigger principle                            | 19          |
| <b>8.6 Seismic data transmission</b>                             | 19          |
| 8.6.1 General considerations                                     | 19          |
| 8.6.2 Types of data transmission links used in seismology        | 20          |
| 8.6.3 Simplex versus duplex data transmission links              | 21          |
| 8.6.4 Protocols for continuous data transmission                 | 21          |
| 8.6.5 Data transmission protocols and some examples of their use | 22          |
| 8.6.6 Compression of digital seismic data                        | 25          |
| 8.6.7 Error-correction methods used with seismic signals         | 26          |



|            |                                                                                         |    |
|------------|-----------------------------------------------------------------------------------------|----|
| 8.6.8      | Seismic data transmission and timing                                                    | 27 |
| 8.6.9      | Notes on dial-up phone lines and selection of modems                                    | 27 |
| <b>8.7</b> | <b>Some network examples</b>                                                            | 27 |
| 8.7.1      | International Monitoring System (IMS)                                                   | 28 |
| 8.7.2      | Southern California Seismic Network (SCSN) (E. Hauksson)                                | 28 |
| 8.7.3      | Japanese seismic networks (Hi-net, F-net and K-NET/KiK-net) operated by NIED (Shin Aoi) | 31 |
| 8.7.4      | Chinese seismic networks (Liu Ruifeng, P. Bormann)                                      | 32 |
| 8.7.5      | German Regional Seismic Network (GRSN) (K. Stammer)                                     | 34 |
| 8.7.6      | Norwegian National Seismic Network                                                      | 36 |
| <b>8.8</b> | <b>Seismic shelters</b>                                                                 | 37 |
| 8.8.1      | Purpose of seismic shelters and lightning protection                                    | 37 |
| 8.8.2      | Types of seismic shelters                                                               | 38 |
| 8.8.3      | Civil engineering works at vault seismic stations                                       | 38 |
| <b>8.9</b> | <b>Establishing and running a new physical seismic network</b>                          | 38 |
| 8.9.1      | Planning and feasibility study                                                          | 38 |
| 8.9.1.1    | Goal setting                                                                            | 38 |
| 8.9.1.2    | Financial reality                                                                       | 40 |
| 8.9.1.3    | Basic system engineering parameters                                                     | 43 |
| 8.9.1.4    | Determining the layout of a seismic network                                             | 44 |
| 8.9.1.5    | Number of stations in a seismic network                                                 | 44 |
| 8.9.1.6    | Laying out a new seismic network                                                        | 45 |
| 8.9.2      | Site selection                                                                          | 46 |
| 8.9.3      | VHF, UHF and SS radio-link data transmission study                                      | 49 |
| 8.9.3.1    | The need for a professional RF network design                                           | 49 |
| 8.9.3.2    | Problems with RF interference                                                           | 50 |
| 8.9.3.3    | Organization of RF data transmission network design                                     | 50 |
| 8.9.4      | Purchasing a physical seismic system                                                    | 50 |
| 8.9.4.1    | The bidding process                                                                     | 51 |
| 8.9.4.2    | Selecting a vendor                                                                      | 52 |
| 8.9.4.3    | Equipment selection                                                                     | 52 |
| 8.9.4.4    | The seismic equipment market is small                                                   | 53 |
| 8.9.4.5    | Processing software                                                                     | 53 |
| 8.9.5      | System installation                                                                     | 54 |
| 8.9.5.1    | Four ways of physical seismic system installation                                       | 54 |
| 8.9.5.2    | Organization of civil engineering works                                                 | 55 |
| 8.9.5.3    | Summary recommendations for setting up a new network                                    | 55 |
| 8.9.6      | Running a physical seismic network                                                      | 56 |
| 8.9.6.1    | Tuning of physical seismic networks                                                     | 56 |
| 8.9.6.2    | Organizing routine operation tasks                                                      | 57 |
| 8.9.6.3    | System maintenance                                                                      | 58 |
| 8.9.6.4    | Sensor calibration                                                                      | 60 |
| 8.9.6.5    | Archiving seismic data                                                                  | 61 |
| 8.9.6.6    | Dissemination of seismic data                                                           | 62 |
|            | <b>Acknowledgments</b>                                                                  | 63 |
|            | <b>Recommended overview readings</b>                                                    | 63 |
|            | <b>References</b>                                                                       | 63 |

## 8.1 Introduction

In this Chapter, a brief description of seismic systems will be given. It is intended to provide an overview on basic ideas in seismometry and describes the existing possibilities in the market (year 2009). This Chapter should be of help during the planning stage of a seismic network. For more thorough information about particular elements and concepts in seismometry and seismic recording systems see Chapters 5 and 6, respectively. Note that this Chapter shares most of the figures and some paragraphs with Havskov and Alguacil (2006). Since one of the authors is the same in both, no related acknowledgment is given.

Before 1960, there were generally only individual seismic stations operating independently. Each station made its observations, which were usually sent to some central location. If several stations were operating in a country or region, it was possible to talk about networks. However the time lag between recording and manual processing were very large compared to modern seismic networks. In the 1960s, seismic networks started operating. These were mainly networks made for microearthquake recording, and the distances between stations were a few kilometers to a couple of hundred kilometers. The key feature used to define them as networks was that the signals were transmitted in real time by wire or radio link to a central recording station where all data was recorded with central timing. This enabled very accurate relative timing between stations and therefore also made it possible to make more accurate locations of local earthquakes. Recording was initially analog and, over the years, it has evolved to be nearly exclusively digital. Lee and Stewart (1981) provide a good general description of the pre-digital transmission networks.

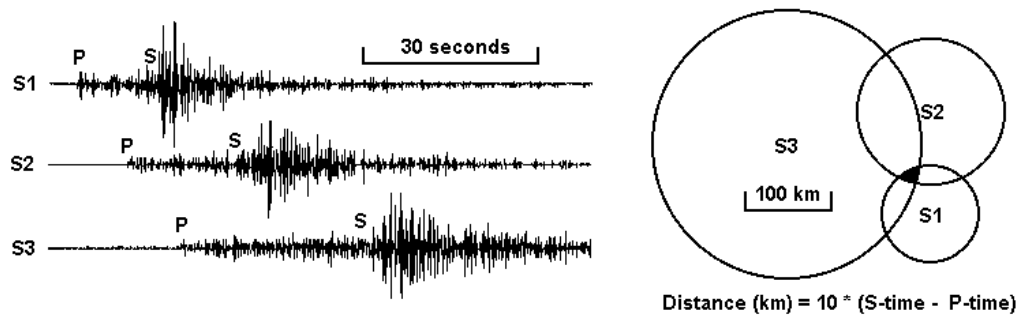
With the evolution of communication capabilities to cover the whole world, seismic networks can now be local, regional or global. The main elements of a modern seismic station are the vault, the seismometer, the digitizer, storage and communication. The distinction between networks is primarily no longer due to differences in data transfer, accuracy of timing, or time lag between data acquisition and analysis, but rather the scope of investigation, spatial resolution, and quality of data in terms of frequency content and dynamic range.

Establishing a new network can be quite challenging and there may be some networks that have not fulfilled their expectations. The main reason for this was probably a lack of knowledge about networks, instrumentation and data processing techniques. Yet such specialized knowledge is unquestionably required if one expects to establish and operate a truly beneficial seismic network. For that reason, in addition to the general description of networks, this document will also outline the basic steps to follow in order to establish a new seismic network.

## 8.2 Seismic network purpose

The main purpose of seismic networks is to determine earthquake locations and magnitudes, to issue alarms, general or specific seismic monitoring and to provide data for research on the interior of the Earth. However, the most basic goal is the determination of accurate earthquake locations. For calculating the three hypocenter coordinates one needs at least the P onset times measured at three stations, or from 4 stations if also the origin time has to be determined as the fourth unknown parameter (see IS 11.1). If, however, the hypocentral distance can be determined for each station independently from the difference of onset times of different seismic phases, e.g., S-P, then 3 stations are generally sufficient (Fig. 8.1). Local, regional

and global research into the Earth's interior is the oldest goal of seismology. Seismic networks are and will be probably forever the only tool that enables study of the detailed structure and physical properties of the deeper Earth's interior.



**Fig. 8.1** Location by the circle (or arc) method. To the left is shown the seismograms at stations S1, S2 and S3 recording a local earthquake. Note that the amplitude scales are different. The stations are located at S1, S2 and S3 (right). The time separation between the P- and S-wave arrivals multiplied by the ratio  $v_P \cdot v_S / (v_P - v_S)$  of the P- and S-wave velocities gives us the hypocentral distance (distance from the station to the earthquake's focus at depth  $h$ ). The epicenter, which is the projection of the focus to the surface, is found within the black area where the circles cross with some “overshoot”. The circles will rarely cross at one point, which indicates errors in the observations, errors in the model, and/or a subsurface source depth. With only two stations, we see that there are either two possible locations, or no possible location if the two circles do not intersect. With more than three stations, the uncertainty in location decreases. Minimization of large circle overshoot can be used to estimate the source depth (see Demo to EX11.1). Note that the “rule-of-thumb” formula given for the distance calculation in the lower right of the figure is for Sn-Pn only. It would be  $8 \cdot (S\text{-time} - P\text{-time})$  for Sg – Pg.

The seismic alarm function, which requires an immediate response after strong earthquakes, serves civil defense purposes with the goal of mitigating the social and economic consequences of a damaging earthquake. Governments, which often finance new seismic networks, emphasize this goal. A related self-organizing seismic early warning information network for Istanbul (SOSEWIN) has been described by Fleming et al. (2009). But seismic monitoring also aides in the long-term mitigation of seismic risk in a region or country as well as in resolving the seismotectonics. Seismic hazard maps of the region may be made which enable the development and implementation of proper building codes. In the long term, building codes are very effective in mitigating seismic risk.

Some cases of seismic monitoring related to seismic risk caused by human activity are of special political concern. This includes monitoring for seismicity induced by large dams or around large mines. Monitoring of seismicity in a volcanic region (see Chapter 13) is also dedicated to volcanic risk mitigation through the prediction of eruptions. Another important function of seismic networks is for explosion monitoring, particularly underground nuclear explosions. Seismic networks are essential to the monitoring of the international nuclear test ban treaty (see Chapter 15).

The purpose of a new seismic network largely defines the optimal technical design for it. Not every design serves equally well for different goals, and many fail completely for some

particular goals. However, modern networks are more capable of dealing with several goals than older networks, which were more narrowly focused due to technical limitations.

## **8.3 Seismic sensors**

### **8.3.1 General considerations**

The choice of an appropriate sensor depends on the application, be it local, regional or global monitoring. The most important factors to consider for a particular application are:

- type of the sensor - accelerometer versus seismometer (compare, e.g., Chapters 11 and 15).;
- number of sensor components per seismic station;
- sensor's sensitivity and dynamic range;
- sensor's frequency range of operation; and
- sensor handling (i.e., how demanding are its transportation, handling, installation, calibration, maintenance etc.; see, e.g. Chapter 7).

### **8.3.2 Seismometers and/or accelerometers?**

During most damaging earthquakes, weak-motion records made with seismometers installed close to the epicenter are clipped. Seismometers are very sensitive to small and distant events and are thus too sensitive for strong-motion signals. This was a very relevant aspect at the time of analog recordings. Traditionally, accelerometers have been considered for strong motion only and seismometers for weak motion. However, the latest generation accelerometers are nearly as sensitive as standard short-period (SP) seismometers down to frequencies around 1 Hz and also have a large dynamic range (up to more than 110 dB; e.g., the Episensor ES-T in DS 5.1). Consequently, for most traditional SP networks, accelerometers would work just as well as 1-Hz SP seismometers and they are similarly priced. At the same time, broadband sensors are now more affordable and have a larger dynamic range. In area where only moderate size earthquakes are expected, broadband seismometers may be sufficient. However, their gain towards higher frequencies (>10 Hz) may be less than for SP sensors. In terms of signal processing, there is no difference in using a seismometer or an accelerometer and correcting digital data for differences in the instrument response can make the signals look identical.

In high seismic risk areas where the main goal of networks is future seismic risk mitigation, strong-motion recordings play an important role, and two sets of sensors will have to be installed so that the system never clips. Although there are significant differences in strong and weak-motion network designs, today both types of sensors are frequently integrated into a single system. Six-channel data loggers with three weak and three strong-motion channels are cost effective and are the current state-of-the-art. They are capable of covering the whole dynamic range of seismic events, from the lowest seismic noise to the largest damaging events. The relative merits of these systems, as well as specific technical details of strong-motion networks, are addressed in the NMSOP\_2 Chapter 7, section 7.4.6 on borehole strong-motion array installations, as well as in several reports and papers of the Consortium of Organizations for Strong Motion Observation Systems (COSMOS), which can be downloaded via the COSMOS website <http://www.cosmos-eq.org>.

### 8.3.3 One- and three-component seismic stations

Historically, many seismic stations and networks used single-component sensors - usually vertical seismometers. Many of them still operate. This was the case because the equipment was analog and the record was often on paper. If three components had been used, three times the amount of equipment would have been required but the information generated would not have been three times more valuable. It was also very difficult, if not impossible, to generate a ground-motion vector from three separate paper seismograms.

Today, in the era of digital recording and processing of seismic data, the situation is different. The price/performance ratio is much more favorable for three-component stations. Most data recorders and data transmission links are capable of accepting at least three channels of seismic data. The cost of upgrading the central processing facilities to accommodate an increased number of channels is relatively small and ground-motion vectors may be generated easily through data processing.

Since ground motion is essentially a vector that contains all of the seismic information, and considering the fact that many modern seismological analyses require this vector as input information, one-component stations are no longer a desirable choice for new installations (not considering seismic arrays which are discussed in Chapter 9). On the other hand, one-component seismic stations are still a choice where communication capability and economy are limiting factors.

### 8.3.4 Sensitivity of seismic sensors

Strong-motion accelerometers are relatively insensitive since they are designed to record the strongest events at small hypocentral distances. Their maximum on scale acceleration is usually expressed as a fraction of the Earth's gravity,  $g$  ( $9.81 \text{ m/s}^2$ ). Accelerometers with 0.25, 0.5, 1, 2, and 4  $g$  full-scale sensitivity are available today. However, modern accelerometers have excellent dynamic range and good signal resolution. They will produce valuable records of smaller events within the close-in epicentral region as well, where seismometer records may still be clipped unless a high-dynamic range recording system is used. Of course, one should order full-scale sensitivity, fitting to the maximal expected acceleration at the sites of the new network. Ordering too sensitive accelerometers may result in clipped records of the strongest and most important events in the region. Accelerometers with too high full-scale range cause diminished sensitivity and needlessly reduce data acquisition resolution of all future records. If the network only uses accelerometers, the most sensitive must be used in order to record small earthquakes if the purpose of the network is not exclusively strong motion.

Weak-motion sensors - seismometers - are usually orders of magnitude more sensitive, however, they can not record as large an amplitude as an accelerometer. They can record very weak and/or very distant events, which produce ground motion of comparable amplitudes to the background seismic noise. Some seismometers can measure ground motion smaller than the amplitudes of the lowest natural seismic noise found anywhere in the world. If one plans to purchase especially sensitive sensors, one must be willing and able to find appropriate, low seismic noise sites for their installation. Standard SP seismometers are in fact so sensitive that they will be able to resolve the ambient Earth's noise above 0.5 Hz in nearly all networks where they are installed. The self-noise of broadband seismometers is typically lower than the

Earth noise down to 0.01 Hz. If the sites are not appropriately chosen and /or have high seismic noise (natural and/or man made), a modern, highly sensitive seismometer is of little use, and a much cheaper sensor, like an accelerometer or a geophone, might be used. However, generally due to affordability and capabilities in terms of frequency and dynamic range, broadband instruments are becoming the standard.

### 8.3.5 Frequency range of seismic sensors

Today's weak-motion sensors are roughly divided into three categories.

The short-period (SP) seismometers measure signals from approximately 0.1 to 100 Hz, with a corner frequency at 1 Hz. They have a flat response to ground velocity for frequencies greater than this corner frequency. Typical examples are the Kinometrics SS-1, the Geotech S13, and the Mark Products L-4C. The 4.5-Hz exploration-type geophone also belongs in this group. This sensor provides reasonably good signals down to about 0.3 Hz at a fraction of the cost of the 1.0-Hz sensor.

Broadband seismometers (BB) have a flat response to ground velocity from approximately 0.01 to 50 Hz. Typical examples are the Guralp CMG3T seismometer with frequency range from 0.008 to 50 Hz, the Wieland-Streckeisen seismometer STS2 with a frequency range from 0.008 to 40 Hz and the similar Chinese CTS-1 (see DS 5.1).

The very broadband seismometers (VBB) can resolve amplitudes of frequencies from below 0.001 Hz to approximately 10 Hz. The best performance at low frequencies is given by the Wieland-Streckeisen STS1 seismometer with frequency range from 0.0028 to 10 Hz (no longer produced). The Chinese JCZ-1 operates in an even wider range from 0.0028 to 50 Hz. However, also the Nanometrics Trillium, the Guralp CMG3T and the STS2 can, proper installation and shielding provided, resolve signals with frequencies below 0.001 Hz (see 5.5.3, 5.5.4, 7.4.4., DS 5.1 and IS 7.5). They are all able to resolve Earth's tides.

On all these different seismometers more information is given in DS 5.1.

The frequency limits shown above are the corner frequencies of the seismometer response functions. This means that analysis below the low-frequency corner and above the high-frequency corner is usually still possible, however, with reduced gain. How much we can extend this range depends on the sensor design, the instrumental self-noise (see Chapter 5, sections 5.5.5 to 5.5.7), the signal-to-noise ratio and beyond the upper corner frequency also on the sampling rate and the steepness of the anti-alias filter (see Chapter 6). The choice of the right sensor depends on its seismological application. In general, the flat portion of the frequency response function should cover the relevant range of frequencies in terms of displacement, velocity or acceleration amplitudes of interest in the investigation of a particular type of seismic event or seismic effect (see Fig. 5.3).

Strong-motion sensors (accelerometers) measure seismic signals between DC and 200 Hz (a typical example is the Kinometrics' EpiSensor; see DS 5.1). However, they differ from the weak-motion sensors in that their output voltage is proportional to ground acceleration and not to ground velocity as it is usual for modern feedback-controlled broadband seismometers. For this reason, as compared to seismometers, accelerometers emphasize high frequencies and reduce low frequency amplitudes. Some strong-motion sensors in the market have no DC

response but a low-frequency, high-pass corner at around 0.1 Hz. These sensors have an important drawback: their records can not be used for residual displacement determination, either of the ground in the near field of very strong earthquakes, or of permanently damaged civil engineering structures after strong events. They are considered as less appropriate for seismic applications where low-frequency signals are important too. The following table should help in the selection of appropriate sensors. It shows some typical seismological applications and their approximate frequency range of interest.

**Tab. 8.1** Application description and approximate frequency range of interest.

| <b>Application</b>                                           | <b>Frequency range (in Hz)</b> |
|--------------------------------------------------------------|--------------------------------|
| Seismic events associated with mining processes              | 5 - 2000                       |
| Very local and small earthquakes, dam induced seismicity     | 1 - 100                        |
| Local seismology                                             | 0.2 - 80                       |
| Strong-motion applications                                   | 0.0 - 100                      |
| General regional seismology                                  | 0.05 - 20                      |
| Frequency dependence of seismic-wave absorption              | 0.02 - 30                      |
| Energy calculations of distant earthquakes                   | 0.01 - 10                      |
| Scattering and diffraction of seismic-waves on core boundary | 0.02 - 2                       |
| Studies of dynamic processes in earthquake foci              | 0.005 - 100                    |
| Studies of crustal properties                                | 0.02 - 1                       |
| Dispersion of surface waves                                  | 0.003 - 0.2                    |
| Free oscillations of the Earth, silent earthquakes           | 0.0005 - 0.01                  |

### 8.3.6 Short-period (SP) seismometers

The SP sensors were historically developed as 'mechanical filters' for mitigating distracting natural seismic noise in the range 0.12 - 0.3 Hz. This noise heavily blurred small events on paper seismograms. However, with today's digital and high-resolution data recording and processing, this rigid 'hardware' filtering can easily be replaced by much more flexible signal processing. A need for sensors that filter seismic signals by themselves does not exist anymore. In addition, when filtering the seismic signal with sensors, we irreversibly lose a portion of seismic information and introduce undesired signal phase distortion. Nevertheless, the SP seismometers, as well as the cheaper geophones, are still a valid selection for several seismological applications, particularly for local seismology where low frequencies of seismic signal are not of major interest or do not exist at all. The traditional 1 Hz sensor might however disappear due to high construction cost compared to modern active sensors.

Most SP electromagnetic seismometers (EDS) are passive sensors with a flat response to velocity above the natural frequency. They are easy to install and operate and require no power, which allows use of smaller backup batteries for the rest of the equipment at remote station sites. They are relatively stable in a broad range of temperatures, which allows less adjustment (and inexpensive) vault designs. The electronic drift and mass position instability usually associated with active sensors are typically not a problem. They are, in short, a very

practical solution for all applications where seismic signals of interest are not expected to contain significant components below 0.1-0.3 Hz.

There are also active SP seismometers in the market, which are either electronically extended 4.5-Hz geophones or accelerometers with electronically generated velocity output. These sensors are often cheaper and smaller. Their drawback is that they require power and are more complicated to repair. An example of such a seismometer is the Lennartz LE-1D.

### **8.3.7 Broadband (BB) seismometers**

Today, the broadband seismometers are a very popular choice. They provide complete seismic information from about 0.01 Hz to 50 Hz and therefore allow a much broader range of studies than the SP records. Even if the prime objective of a network is to locate earthquakes that could be done with SP sensors, other usage of the data in scientific studies benefits from BB sensors.

However, the BB seismometers demand more efforts for installation and operation than SP seismometers. The BB seismometers require a higher level of expertise with respect to instrumentation and analysis methods. They are active feedback sensors and require a stable single- or double-polarity power supply. They also require very careful site selection in a seismological-geological sense as well as a better-controlled environment in seismic vaults in order to benefit from the low frequency part of their output ( $< 0.1$  Hz). Older models are sometimes a bit tricky to install while newer models require less adjustments. Since they do not attenuate the 0.12 - 0.3 Hz natural seismic noise peak (see Fig. 4.7), their raw output signal contains much more seismic noise than signals from a SP seismometer. Consequently, useful seismic signals are often buried in seismic noise and can be resolved and analyzed only after filtering to remove the background noise. So, for all but the largest earthquakes, filtering is required even for making simple phase picks. BB sensors are often perceived as the 'best choice', however at high frequencies ( $> 50$  Hz) their gain is typically less than for SP sensors, and for example for microseismic monitoring the SP sensors will be a better choice.

### **8.3.8 Very broadband (VBB) seismometers**

The VBB seismometers are utilized in global seismology studies. They are able to resolve the lowest frequencies resulting from Earth's tides and free oscillations of the Earth. Their primary purpose is the research of the deep interior of the Earth. Their only important advantage, however, as compared to BB seismometers, is their ability to record frequencies around and below 0.001 Hz. They are expensive, require very elaborate and expensive seismic shelters, and, as a rule, are tricky to install. They are ineffective for seismic risk mitigation purpose and some also lack frequency response high enough for local/regional seismology.

However, data from a VBB station are very useful to the international scientific seismological community. They are also excellent for educational purposes. For a large national project, installation of at least one VBB station is recommended and perhaps two to three in a very large country or region. Site selection and preparation for a VBB station requires extensive study and often expensive civil engineering work (e.g., Uhrhammer et al., 1998 and 7.4.4). The cost of preparation of a single good VBB site can exceed US\$ 100,000.



### 8.3.9 Long-period (LP) passive seismometers

The long-period passive sensors are not a suitable choice for new installations and are not sold anymore. These sensors have a corner frequency of 0.05 to 0.03 Hz and, in that respect, are inferior to most (but not all) BB sensors. Their dynamic range is in the order of 120 dB. An LP sensor with a 24-bit digitizer still makes an acceptable low-cost BB station provided the sensors and the vault are already available. However, nonlinear distortion of such an installation may be problematic. Nevertheless, in the scope of new installations, long-period seismometers are of historical value only. Instruments based on the traditional design are becoming a popular choice for education in school seismology projects.

## 8.4 Seismic network configuration

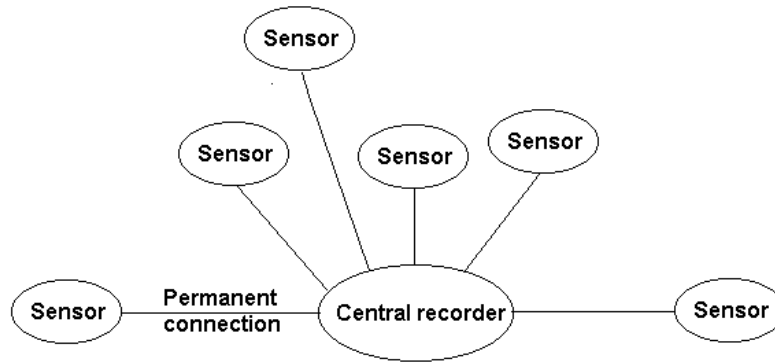
### 8.4.1 Real and virtual seismic networks

When the hardware connection among seismic stations is established, the next question is how the data are sent along the connection and what protocols are used for the units to communicate. This will define, to a large extent, the functionality of the seismic network.

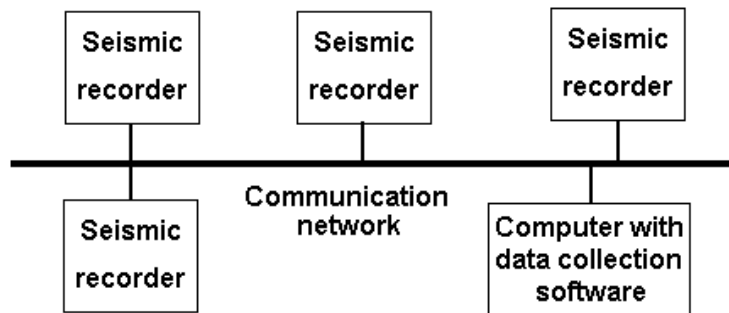
In the days of only microearthquake networks and one-way data transmission (from stations to central-recording site), it was quite clear how a seismic network was defined. Today, the situation is more complex. Nowadays, more and more seismic stations are connected to the Internet and data is available from world-wide distributed stations in near real-time. The stations usually also have a local recording capability. Typically, acquisition, communication and storage is provided by a single unit. Seismic networks are now mostly of this type of stations and the distinction between local, regional, and global networks does not exist any more in terms of hardware, data transmission and acquisition, but is merely a question of how the data are processed.

Due to the improvements in global communication which allows for easy access to data from around the world, the concept of virtual networks has emerged. The term virtual merely indicates that not all stations in the network belong to a single operator or a “real” network. In practice, most networks are virtual. Access to data from a station by another operator could be directly from the station, but often the data is received from the central system. However, in practice it makes little difference in the way stations are connected into either “real” or virtual networks. Examples of virtual networks are the Virtual European Broadband Seismic Network (VEBSN; see <http://www.orfeus-eu.org/Data-info/vebsn.html>), the GEOFON Extended Virtual Network (GEVN-GEOFON; see <http://www.gfz-potsdam.de/geofon/>) the Global Seismographic Network stations and arrays (code \_GSN), the Federation of Digital Broad-Band Seismograph Networks stations (\_FDSN). The “\_” in front of the network code stands for virtual but many national networks become virtual by including data from stations by another operator within or outside the country. Examples of “real” networks are the Global Seismograph Network (GSN - IRIS/IDA) (code II) and the GEOSCOPE network code (G).

As most networks today are virtual networks, it will be assumed in the following that a network is virtual unless specifically noted otherwise.



**Fig. 8.2** Example of a real seismic network. The sensors are connected to a central recorder through a permanent connection like a wire or radio link. The latter may also be via satellite when the network covers large territories. The transmission may be analog with digitization taking place centrally, or an analog-to-digital converter (ADC) could also be placed at each sensor and the data transmitted digitally.



**Fig. 8.3** Example of a virtual seismic network. The thick line is the communication network, which can have many physical implementations. The data collection computer collects data from some or all of the recorders connected to the network. This network could also be a real network if all stations belonged to a single operator.

## 8.4.2 Seismic networks

### 8.4.2.1 Stand-alone and central-recording based seismic systems

In the simplest case, a seismic network is a group of stand-alone seismic stations with a local recording medium. Many of the older networks, particularly analog ones, are still of this type. The information is gathered in person, either by collecting paper seismograms or by locally downloading digital data from stations into a laptop computer. There exist no communication links from the remote stations to the data center. Data can be stored locally on a removable memory medium like memory cards, or removable hard disk. Temporary networks are typically deployed in this way. Such networks can either record in continuous mode or in triggered mode. Earlier, triggering mode was the most common due to limitation in the storage capacity, while today, continuous mode is the most common. Triggering mode is now mostly used in strong-motion seismology where recordings are rare. As a permanent, national, or regional observatory seismic network, this design is rarely suitable.

Most modern networks involve near real-time data transfer from the remote stations to the central processing site. The real-time data can be received over a radio link, satellite or some form of Internet communication. It is quite common to have various means of communication within one network. Data is received by the client at the receiving end from the server at the field stations or another data centre. Most such networks now store continuous data. At present, most networks operate in this fashion. There can be a trade-off between the indispensable remoteness of the seismic stations and the availability of a communication link. However, satellite and possibly mobile phone based solutions are applicable in remote locations. It is also possible to bridge the distance to a point with permanent communication using short-distance spread spectrum RF links.

Most real time networks have a central trigger function and can, therefore, use coincidence detection algorithms in near real time (see 8.5.2).

It is common that networks use more than one system to acquire and process the data. It is also possible that some seismic stations are linked, but not sending data in real-time. It is, therefore, often necessary to make use of software that combines the different systems into one, and brings the data onto a single platform for interactive data processing.

The decision on which type of network to use depends on the availability of communication and the requirement for alarm functionality. For seismic networks with the exclusive purpose of monitoring general seismicity and/or to serve research purposes, there is no need for real-time data. The two main factors in deciding which network is the most appropriate are cost of ownership and quality of data. For research purposes, flexibility is also a very important issue. If an Internet based system is available at seismically quiet sites, this may be the least expensive way to construct a network. Since communication costs are quickly decreasing and links are becoming universally available, real-time networks are now most common.

#### **8.4.2.2 Examples**

*Example 1:* IRIS is the largest global data centre and receives 1,000s of seismic channels in near real-time. The Global Seismograph Network (GSN) itself consists of more than 150 stations. The remaining data is provided by other networks from around the globe. Subsets of the data are combined into virtual networks, such as the GSN and FDSN networks. Data are made freely available as continuous near-real time data streams, or on request as continuous data files and event-based data sets. Very importantly, the GSN is currently re-equipped (completed by 2013) with standardized data collection systems based on the Quanterra Q330HR (high resolution) systems and has embarked on supplementing the yearly relative calibration procedures by network operators' in situ calibrations with portable equipment, thus verifying sensor and system responses as well as sensor orientation and location. Moreover, complete instrument description is available for all data. A number of mechanisms are supported to provide the data. (<http://www.iris.edu/data/request.htm>). The real-time data is made available through LISS and SeedLink (see 8.6.4 and Chapter 10) and event data through tools such as Wilber, BreqFast and WebRequest. Fig. 8.4 shows the GSN network and Fig. 8.11 the type of communications used.

*Example 2:* The public domain SEISNET system is another software enabling establishment of semi-real time virtual seismic networks. SEISNET operates with other types of stations in addition to the GSN stations and also performs network detection and preliminary location

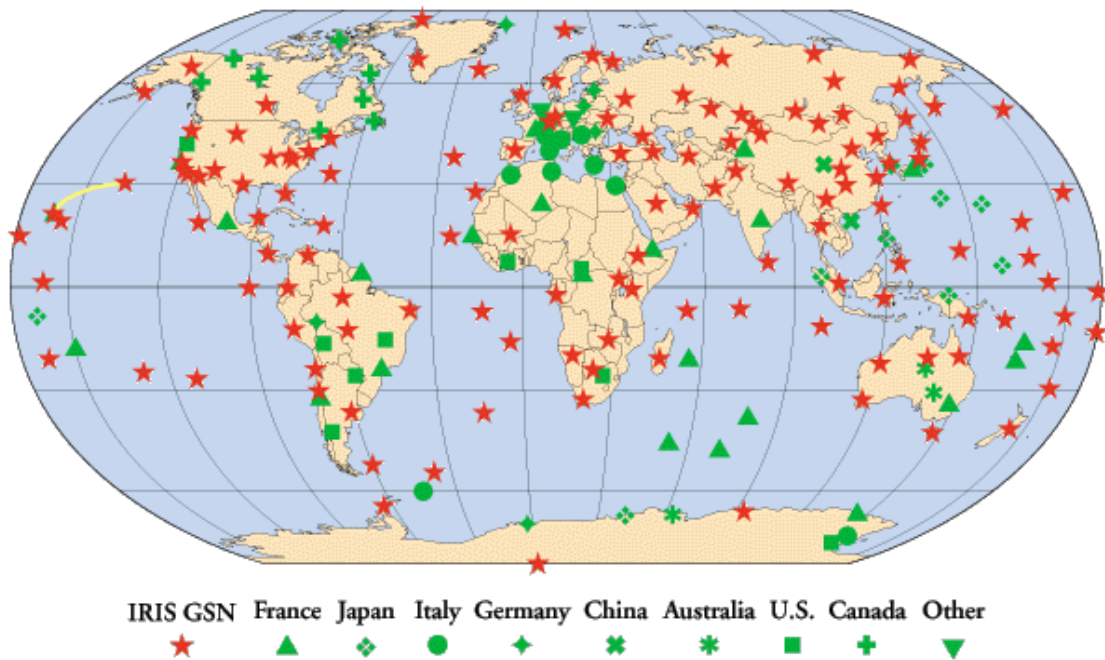
(Ottemöller and Havskov, 1999). It was developed for the Norwegian National Seismic Network and is also used in several other places. SEISNET is very flexible and can be adopted for virtually any type of field station.

*Example 3:* Another widely used and publicly available software is EARTHWORM (EW) (runs on Solaris, Linux, Mac OS X, and Windows), also see <http://www.isti2.com/ew/> . It allows users to run real-time seismic networks of different purpose with emphasis on either real-time seismic data processing or data storage and user interaction. EW includes real-time data transport, automated event detection, phase picking, seismic event association and location, archiving, and other modules, and was originally developed by the U.S. Geological Survey's (USGS) Northern California Seismic Network (NCSN). EW is used by a world-wide community of ~150 institutions that operate the system or its derivatives and contribute to its development and maintenance. Instrumental Software Technologies, Inc. (ISTI) coordinates and leads this worldwide effort. EW modules are also used in other software systems such as HYDRA, used by NEIC, Earlybird used at the Alaska Tsunami Warning Center, and ANSS Quake Monitoring System (AQMS) used by the ~450-station NCSN, the ~400-station SCSN California networks, five other ANSS funded US regional networks, as well as by many other earthquake and volcano networks world-wide.

*Example 4:* The proprietary ANTELOPE software is yet another real-time seismic network software package on the market. It supports a wide range of seismic stations as well as other environmental monitoring equipment. ANTELOPE's open-architecture, modular, UNIX-based, real-time acquisition, analysis, and network management software supports all telemetry using either standard duplex serial interfaces or standard TCP/IP protocol over multiple physical interfaces. In addition to data acquisition, the seismic network functionality includes real-time automated event detection, phase picking, seismic event association and location, archiving, system state-of-health monitoring, interactive control of remote stations, automated distribution of raw data and processed results, batch mode seismic array processing and a powerful development toolkit for system customizing. It can handle continuous and event file-based data and uses relational database management formalism and the CSS v. 3.0 scheme for information organization. It runs on Sun Microsystems' Solaris OS on SPARC and Intel architectures. It was developed by the BRTT Company and Kinemetrics Inc. and is currently used by IRIS networks, the US Air Force, several seismic networks in the US, and about eight national seismic networks in Asia and Europe.

*Example 5: Geofon network.* The most used software in Europe is the public domain SeisComp – SeedLink system developed by GFZ (German Research Centre on Geosciences in Potsdam) (see <http://geofon.gfz-potsdam.de/geofon/seiscomp/> and <http://www.seiscomp3.org/>) and the largest network using this software is the global GEOFON network consisting of GFZ stations and cooperating stations in other countries (see Fig. 8.5). The stations in the GEOFON network all send their data in real time to the GFZ in Potsdam. Rapid global earthquake information has recently become a major task of the GEOFON Program. E.g. GFZ is a key node of the European Mediterranean Seismological Centre (EMSC) and has the responsibility for rapid global earthquake alerts. The GFZ became also a leading force in earthquake monitoring for tsunami warning using SeisComp in the Mediterranean and the NE Atlantic as well as for the Indian Ocean.

# GSN & FEDERATION OF DIGITAL BROADBAND SEISMIC NETWORKS (FDSN)



**Fig. 8.4** The Global Seismic Network (GSN) and other global broadband stations that are members of the Federation of Digital Broad-Band Seismograph Networks (FDSN) (figure from IRIS home page <http://www.iris.edu>).



**Fig. 8.5** The GEOFON network. Figure from <http://geofon.gfz-potsdam.de/geofon/>

## **8.5 Seismic data acquisition and data processing**

### **8.5.1 Digital versus analog data acquisition**

There exist three primary types of physical seismic networks with respect to the technology of data acquisition: analog, mixed, and digital.

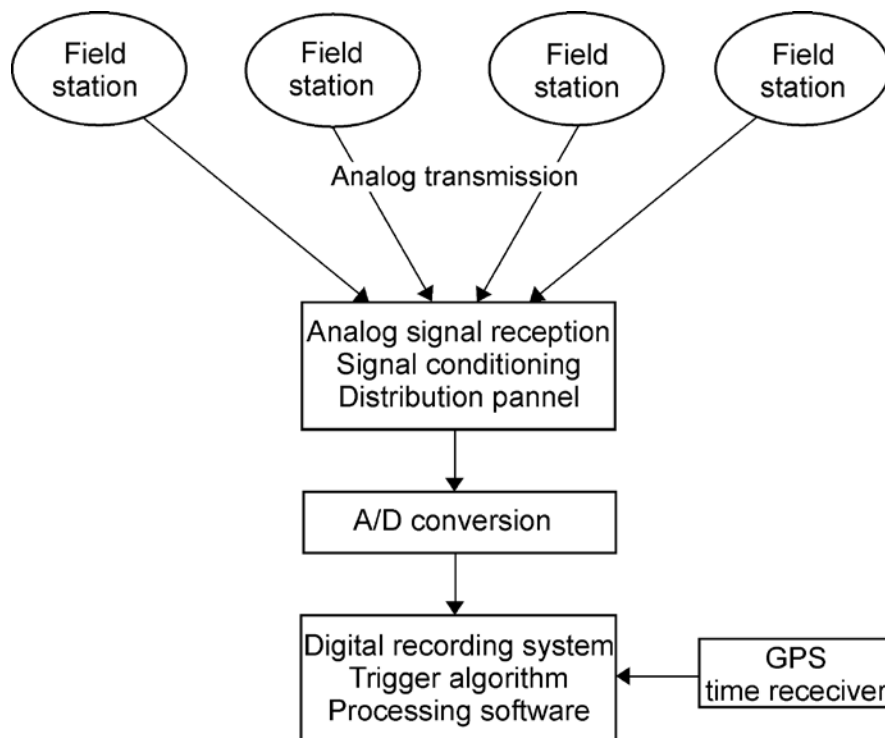
#### **8.5.1.1 Analog seismic systems**

The analog seismic systems include sensors, which are always analog, analog signal conditioning, usually frequency modulated (FM) telemetry through radio (RF) or phone lines, analog demultiplexers, and analog drum or film recorders. Paper or film seismograms are the final result of a completely analog system. Analog systems have serious drawbacks and are no longer built. The low dynamic range and resolution of the acquired data (about 40-45 dB with single and about 60-65 dB with double, low and high-gain data transmission channels) lead to issues of incomplete data. On the one hand, many events have amplitudes that are too low to be resolved on paper or film records, while on the other hand, many records are clipped because their amplitude is too large for undistorted recording. In fact, only a very small portion of the full dynamic range of earthquakes that are of interest to seismologists are actually recorded distortion free on analog systems. Another problem is the incompatibility of paper and film records with computer analysis. This is a very serious drawback today because modern seismic analysis is almost entirely based on computer processing.

#### **8.5.1.2 Mixed analog/digital systems**

Mixed systems, frequently erroneously called digital, have analog sensors, analog signal conditioning, usually FM telemetry, and analog demultiplexers, but digital data acquisition at the central-recording site, digital processing, and digital data archiving.

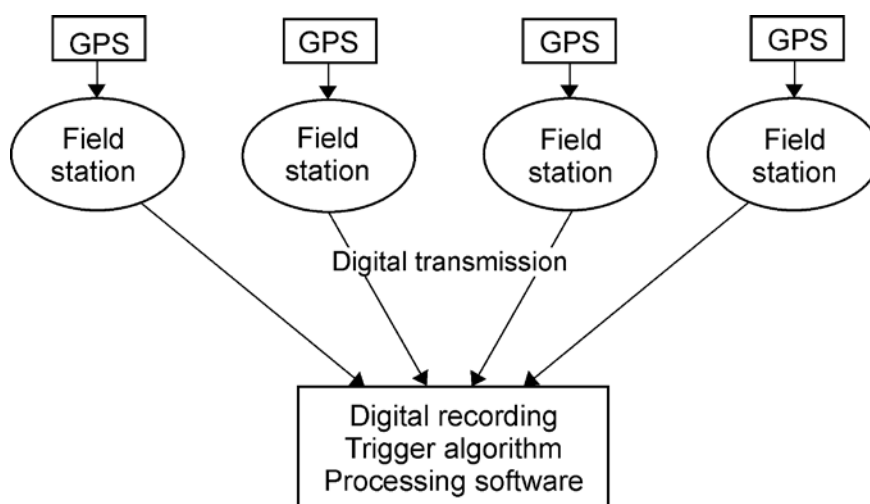
Such systems also have a low dynamic range (usually FM data transmission links are the limiting factor) and therefore, they have the same disadvantage as the analog systems regarding data completeness and quality. However, they can accommodate off-line as well as automatic near-real time computer analysis. One can use most modern analysis methods, except those that require very high-resolution raw data. Such systems are still useful for some applications when the higher dynamic range of a fully digital system is not of prime importance and the purpose of the seismic network is limited to a specific goal. Advantages of these systems include low cost and low power consumption of the field equipment. Fig. 8.6 shows a typical setup. Many systems are still running, but no new ones are built.



**Fig. 8.6** Typical analog-digital network. The analog data is transmitted to the central site over fixed analog communication channels, usually FM modulated radio links or phone lines. At reception, the signals are put into a distribution panel where incoming signals are demodulated. Some filtering may take place before the data are digitized by a PC or similar recording system. Timing is done within the digital system. Today, very few alternatives to GPS exist.

### 8.5.1.3 Digital seismic systems

In digital systems, only the seismometers are analog. All other equipment are digital. The dynamic range and the resolution are much higher than that of analog and mixed type systems. These factors depend mainly, but not only, on the number of bits of the analog-to-digital (A/D) converter. 16- to 24-bit A/D converters are available today, which correspond to dynamic ranges of approximately 90 to 140 dB. In practice, however, the total dynamic range and the resolution of data acquisition is usually less than the number of bits an A/D converter would theoretically allow, since 24-bit converters rarely have a noise level as low as 1 bit. Nearly all A/D converters sold now are of 24 bit type. Fig. 8.7 shows a typical setup of a digital seismic network.



**Fig. 8.7** Typical digital network. The digital data is transmitted to the central site over fixed digital communication channels. At reception, the signals enter the recorder directly. Timing normally takes place at the field stations, although some systems also time stamp the signal on arrival.

Buyers of digital seismic networks sometimes ask for additional paper drum recorders because they wish to continuously monitor incoming signals and/or believe drum recorders will serve as an excellent educational tool. However, there are a number of problems with paper drum recorders in digital systems. One problem revolves around the requirement for additional electronic components, such as digital to analog converters. Being mechanical devices, drum recorders are and will continue to be expensive, often costing more than a multi-channel digital recorder. They require continuous and specialized maintenance and consumables. On the other hand, nearly all modern observatory seismic software packages allow for the continuous, near-real-time observation of the incoming signals and some even simulate the traditional appearance of paper helicorder records. Once a user becomes familiar with a digital system, expensive paper drum recorders soon prove to be of little use and are thus a poor investment.

## 8.5.2 Trigger algorithms and their implementation

### 8.5.2.1 Continuous versus triggered mode of data acquisition

Continuous, digitally-acquired seismic signals by their very nature provide a large amount of data. A reasonably sized, digital, weak-motion seismic network operating in continuous mode will produce a large volume of data. In the past, due to limited storage capability, only the event data was stored. However, now this is no longer a problem and all data can be stored. Recent studies make use of continuous noise recordings, which shows that what previously was considered useless data can now provide valuable results (see Chapter 14). With modern networks it is no longer a question and continuous data must be stored.

The storage problem had frequently led seismic network users to operate their systems on a "triggered" basis (particularly the local and regional seismic networks that require a high frequency of data sampling). Triggered only systems still do continuous, real-time acquisition



and processing of seismic signals, but only store signals associated with seismic events. Such systems do not store continuous time histories of seismic signals, but rather produce "event files". Obviously, systems in triggered mode will lose some weak events and produce a certain number of false triggers. The completeness of data inevitably is impaired because the efficiency of the trigger algorithms currently available is inferior to the pattern recognition ability of a trained seismologist's eye.

The continuous seismic signal recording provides the most complete data, but processing and storage of all data can be difficult and expensive. Continuous recording can be achieved through the use of a ring-buffer system, where a fixed size is allocated to the data and the oldest data is overwritten by the newest. Some systems simply write data files and remove the oldest files when the storage capacity is reached. Most systems now record continuously and run a trigger algorithm only to produce detection lists. However, strong-motion systems typically still operate in triggered mode.

#### **8.5.2.2 Trigger algorithms**

The trigger algorithms used on a seismic recorded or in a central processing system are basically the same.

The amplitude threshold trigger simply searches for any signal amplitude exceeding a preset threshold. A detection is made whenever this threshold is reached. This algorithm is normally used in strong-motion seismic instruments, which are systems that do not require high sensitivity. Consequently, man-made and natural seismic noise will only produce infrequent triggers.

The root-mean-square (RMS) threshold trigger is similar to the amplitude threshold algorithm, but the RMS value of the amplitude in a short time window is used instead of 'instant' signal amplitude. It is less sensitive to spike-like, man-made seismic noise, however it is rarely used in practice.

The ratio of the short-time average to the long-time average (STA/LTA) of the seismic signal is the basis of the most frequently used trigger algorithm in weak-motion seismology. It continuously calculates the average values of the absolute amplitude of the seismic signal in two consecutive moving time windows. The short-time window (STA) is 'sensitive' to seismic events, while the long-time window (LTA) provides information about the temporal amplitude variation of seismic noise at the site. When this ratio exceeds a preset value (usually set between 4 and 8), an event is 'declared'. The STA/LTA trigger algorithm is well suited to cope with slow fluctuations of natural seismic noise. It is less effective in situations where man-made seismic noise of a bursting or spiky nature is present. At sites with high, irregular, man-made seismic noise, the STA/LTA trigger usually does not function well. For more details on STA/LTA algorithm and parameter setting see IS 8.1.

Several more sophisticated trigger algorithms are known from the literature. They are sometimes used in seismic networks but rarely in the seismic data loggers available on the market. In the hands of an expert they can significantly improve the event detections/false triggers ratio, particularly for a given type of seismic event. However, these triggers often require sophisticated parameter adjustments that can prove to be unwieldy and subject to error.

Every triggered seismic system must have an adjustable band-pass filter in front of the trigger algorithm. This is particularly important with BB and VBB seismometers where small earthquake signals are often buried in dominant 0.2-0.3 Hz seismic noise. The adjustable band-pass filter allows the trigger algorithm to be sensitive to the frequency band of one's interest. In this way such events may be resolved and acquired. Some recorders allow several trigger sets to be used simultaneously. This is needed if for example, a BB station has to trigger on microearthquakes, teleseismic P waves and surface waves which each require separate setting of filters, STA and LTA. The GSN Quanterra stations operate in this way.

### **8.5.2.3 Coincidence trigger principle**

In seismic networks with standalone stations, each remote station has its own independent trigger. In such networks data are usually transferred to the central-recording site on request only or it is collected in person. These seismic networks have the lowest effectiveness of triggering and consequently, the smallest detection threshold and/or the highest rate of falsely-triggered records. The completeness of data is modest because not all stations in the network trigger simultaneously for each event. This approach requires a good deal of routine maintenance work in order to "clear" numerous false records from the local data memory if trigger thresholds are set low; if not, the network has a lower detection threshold. Remote stations may encounter 'memory full' situations due to having a limited local memory. Such networks absolutely require the careful selection of station sites with as low as possible man-made seismic noise. If low noise is not assured, an observatory quality network may be so insensitive as to be considered a serious project failure. However, such networks are frequently used as temporal seismic networks. They also function well where high sensitivity is not desired at all, for example, in most strong-motion networks.

Seismic networks that use the coincidence trigger algorithm are much better at detection thresholds and completeness of acquired data. In these systems, data are transmitted continuously from all remote stations to the central-recording site where a complex trigger algorithm discriminates between seismic events and seismic noise. The coincidence trigger takes into account not only signal amplitudes but also correlation in space and time of the activated stations within a given time window (the window allows for wave propagation). The trigger threshold level of such a robust algorithm can be significantly lowered, resulting in a more complete record of small events for the entire network. All stations in the network are recorded for every trigger, which greatly improves completeness of the recorded data.

## **8.6 Seismic data transmission**

### **8.6.1 General considerations**

While data transmission may not seem like an important technical issue for a seismic network, a poorly selected or designed data transmission system is one of the most frequent causes for disappointment and technical failures. The success of a seismic network operation rests largely on the reliability and the quality of data transmission, whether by dedicated lines or by internet.

Another very important but frequently overlooked factor is the cost of data transmission. In fact, these costs may largely determine the budget for a long-term seismic network operation. Many seismic networks all over the world have been forced to change to less expensive modes of transmission after some years of operation. The data transmission costs per year in a network established right after a damaging earthquake may seem completely acceptable at first, but may be viewed as excessive after just a few years of relative seismic quiescence. However, communication is becoming easier and cheaper and this should no longer be an issue.

There are some key technical parameters to consider in designing a data transmission system:

- the required information flow (channel bandwidth for analog links or data transfer rate with digital links);
- the distance to which data must be transmitted;
- the desired reliability (acceptable down-time of the links, that is, the maximum time period per year when the signal-to-noise ratio is lower than required (analog links) or bit error rate is higher than allowed (digital links).
- the communication system which will be used to establish the seismic network (Internet, proprietary WANs (Wide Area Networks), analog public phone network, ISDN, ADSL etc.); and
- the protocols that will be used.

These parameters must fit the available data transmission infrastructure in the country or region, the available network operations budget, and the network's performances goals. Technical considerations, reliability, initial price and operational costs of data transmission links vary widely from country to country. Local conditions in a particular country or region are a very important factor in the selection of an appropriate data transmission system. It is essential to get information about the availability, reliability and cost of different approaches from local communication experts.

### **8.6.2 Types of data transmission links used in seismology**

In seismometry there are several different kinds of data transmission links in use, from simple short-wire lines to satellite links. They differ significantly with respect to data throughput, reliability of operation, maximum distance, robustness against damaging earthquakes, and cost of establishment and operation. A table in IS 8.2 enumerates the most common types, their major advantages and drawbacks, and their potential applications.

Note that strong-motion seismic networks generate far less data than weak-motion networks and therefore, their design might differ significantly. Seismic data transmission links that are fully acceptable for strong-motion data may be inadequate for weak-motion data and data transmission links used in the weak-motion field may be an absolute overkill and too expensive for strong-motion networks.

### **8.6.3 Simplex versus duplex data transmission links**

There are two basic types of digital data transmission links.

Simplex links transmit data only one-way, from remote stations to the center. These links are relatively error prone. Radio interference or fading may corrupt data during transmission and there is no way of recovering corrupted data, unless forward error-correction (FEC) methods are used (see 8.8.6.6). However, the FEC methods are rarely used except for satellite links. They require a significant bandwidth overhead, which is hard to provide using standard, low cost 3.5-kHz bandwidth RF channels. Simplex links usually use the type of error-checking that allows recognition of corrupted data but not its correction. The methods in common use range from a simple parity check or check-sum (CS) error detection to cyclic redundancy check (CRC) methods.

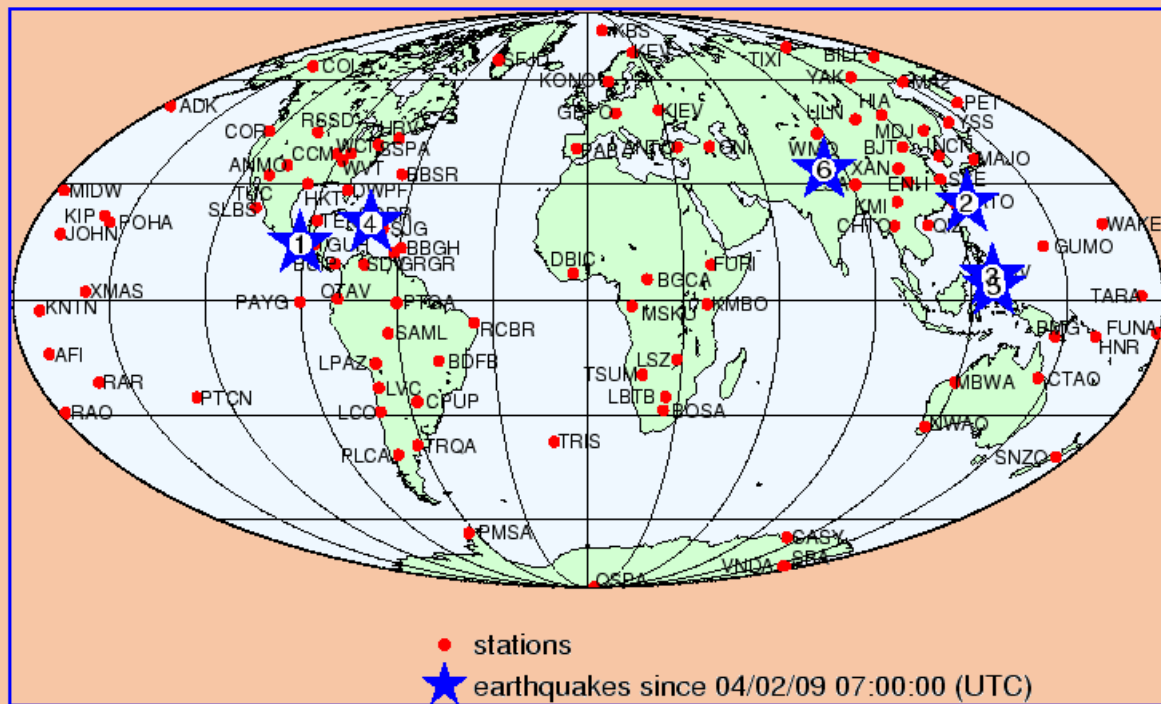
Duplex links allow data flow in both directions – from the remote station to the center and vice versa. Different types of error-checking methods are used, ranging from a simple parity check or CS error detection to CRC error detection. Once an error is detected, the data block is resent repeatedly until it is received correctly. In this way, a very significant increase of reliability of data transmission is achieved. Duplex is the standard with modern equipment and communication. However, duplex radio links nearly double the amount of the equipment and are therefore expensive compared to the simplex links.

Today, no commercial networks using simplex are constructed.

### **8.6.4 Protocols for continuous data transmission**

The majority of new networks and many old networks will now record continuously and this has resulted in de facto standards emerging from major operators. The first protocol was LISS (Live Internet Seismic Server, <http://www.liss.org/>) made by USGS and used on the GSN stations (see Fig. 8.8). The protocol sends 512 byte blocks of MiniSEED data on a socket and application on the receiver side can open a corresponding socket and continuously receive the data. LISS has the drawback that if data is lost, there is no way of requesting data since it is just a one way transmission.

This problem has been solved using the SEEDLink protocol that is implemented within the SEEDLink software (Heinlo and Trabant, 2003). This protocol is the same as LISS with an additional data block containing information about transmission status and request information. SEEDLink can therefore request missing data blocks which also makes it possible to request data back in time. SEEDLink is the most used public standard with retransmission and data from large data centres such as IRIS and ORFEUS can be received from their SEEDLink servers. Other real-time data processing systems such as Earthworm come with data exchange protocols. Another widely used protocol is CD1 developed for the International Monitoring System of the CTBTO. In addition, all major seismic equipment companies have developed their proprietary protocols for real-time data exchange, but they may also support the open standards.



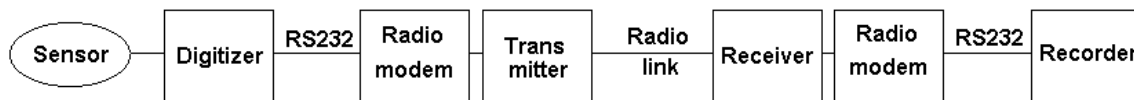
**Fig. 8.8** LISS stations (red dots) April 3, 2009. The map shows seismograph stations which are transmitting their data live, via the Internet, to the Albuquerque Seismological Laboratory ([http://aslwwww.cr.usgs.gov/Seismic\\_Data/telemetry\\_data/map\\_sta\\_eq.shtml](http://aslwwww.cr.usgs.gov/Seismic_Data/telemetry_data/map_sta_eq.shtml)). The live data can be seen interactively by pressing the red dot.

### 8.6.5 Data transmission protocols and some examples of their use

Serial data communication and Ethernet are the most commonly used ways to transmit digital seismic data.

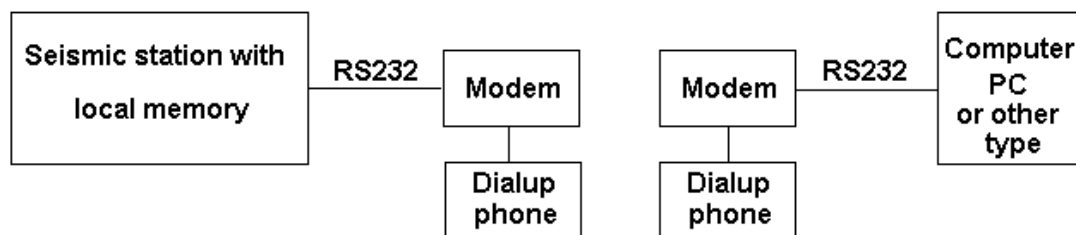
Most seismic digitizers will send out a stream of data in serial format and all computers have hardware and software to communicate with serial data. A serial line requires at least 3 lines: One for sending data, one for receiving data, and ground. If data only is to be sent or received, two lines suffice. The serial lines use either RS-232 protocol or the RS-422 protocol. The former can run on up to 50 m long cables and the latter on cables up to 2 km long. Serial line communications may be used by modems, radio links, fixed telephone lines, cellular phone, and satellite links.

*Example 1:* An example of one-way continuous communication (see Fig. 8.9), a remote station has a digitizer sending out RS-232 data, which enters a radio link to a PC, which reads the data and processes it. The communication is governed by the RS-232 protocols. Data acquisition system is done on the PC.



**Fig. 8.9** One-way communication from a remotely installed digitizer via a digital radio link to a centrally located PC. The radio modem and transmitter /receiver might be one unit.

*Example 2:* Interactive communication with a remote seismic station (see Fig. 8.10). A user calls up a terminal emulator on his PC, connects to a modem with one of the PC's serial lines, dials the phone number of the modem connected to the remote station, and logs into the field station. Once logged in, several options are usually available. One is to browse a log file containing all triggered events in the local memory of the station. Another option is to initiate a download of event data. A very common way to do this over a serial line connection is to list the event file in ASCII form and then set up the terminal emulator at the local PC to capture the data. This is one way of getting data from a standard GSN seismic station. The advantage of this type of communication is that only very simple software is required and it is easy to access to many different seismic stations. This process can also be easily automated (see IS 8.3).



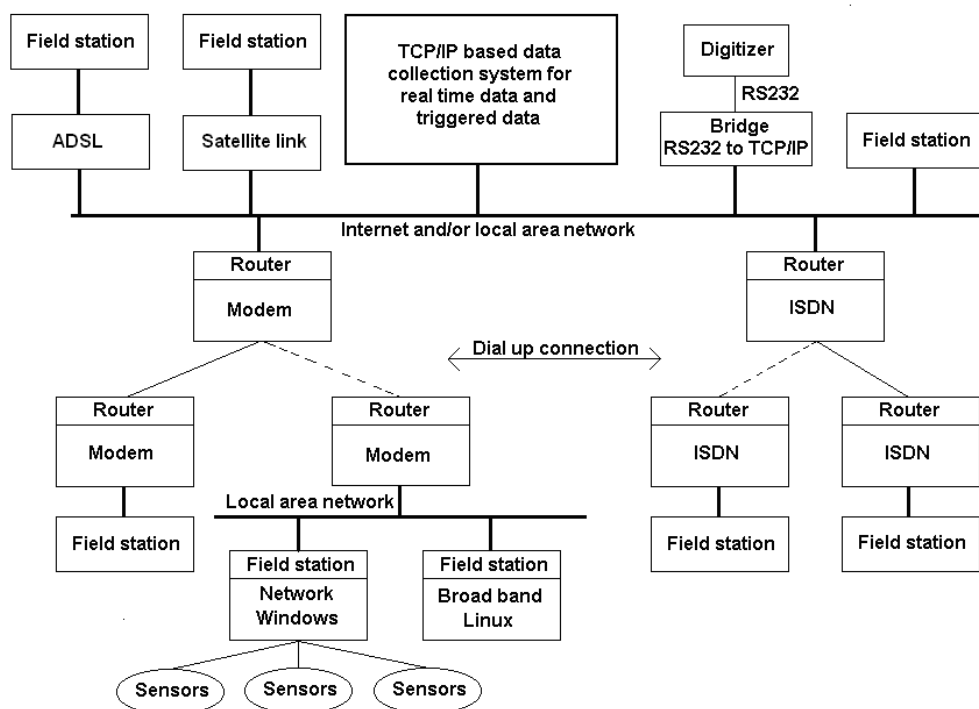
**Fig. 8.10** Manual dial-up to a seismic station for data inspection and/or download. The dialing computer can be of any type as long as a terminal emulator program, such as Hyperterminal in MS Windows, is available.

*Example 3:* Interactive communication with a remote seismic station using proprietary software. The user starts up a manufacturer-supplied program on his local PC. The program handles all the communication to the field station purchased from this manufacturer. The user's connection with the field station will be as if sitting next door. Data download, acquisition parameter settings, system state-of-the-health verification, and diagnostic commands (if applicable) are managed through simple menus, and the event files may be automatically transferred to the user's local PC. The process can be run manually or in an automatic, unattended mode at specified times. With some systems, remote stations can initiate the transfer of triggered seismic events. The advantage with this setup is that communication with a particular remote station is very easy. Unfortunately, most of the software systems in the market work only with one type of seismic station.

In high-speed local area networks, Ethernet most commonly connects computers. This low-level protocol is not what the user sees directly, but rather a high-level communication

protocol working on top of the Ethernet protocol. TCP/IP is the most widely used protocol for file transfer and remote log-in. This is also the protocol used by the Internet although the low-level protocol used between the different Internet nodes might not be Ethernet. TCP/IP can also be used over serial lines and ISDN telephone lines. Today, most seismic recorders are able to communicate via TCP/IP protocol. A remote seismic station, which can be reached by TCP/IP either through Internet, satellite, ADSL, ISDN, or regular phone lines, represents the most general purpose and flexible system available.

Fig. 8.11 shows the most common way of establishing TCP/IP connections to a central data collection system. Dashed lines between routers indicate that the connection is made to one station at a time. Large central routers that can communicate with many ISDN nodes at the same time are also available.



**Fig. 8.11** Different ways of getting a TCP/IP connection to a central data collection system. The thick solid lines indicate permanent Ethernet connections.

Getting seismic data from a GSN station using Internet via a local computer is simple. The user uses the Telnet to login to the station. Once logged in, he can check available seismic data and use the FTP file transfer protocol to copy the data to the local computer. The process is easy to automate. Fig. 8.12 shows the communication links for the GSN network. It is seen that almost all stations have a permanent connection and consequently almost all continuous data is transmitted by the computer network to IRIS. Only going back to 2001, the majority of stations were dial up and data had to be sent to IRIS on tape.





temporary memory and the link's throughput will suffice in case of large, long-duration events.

### **8.6.7 Error-correction methods used with seismic signals**

All digital communications experience errors. In the transmission of seismograms this is particularly fatal since just one bit of error might result in a spike in the data with a value a million times larger than the true seismic signal. Obviously, this could wreak havoc in trigger systems, and one byte missing in an event file might corrupt the whole event file.

One of the principles of error correction is that data is sent in blocks, e.g., 1 s long, and along with the block of data there is some kind of 'check-sum'. If the check-sum does not tally with the received data, a request is sent to retransmit that particular block of data. Obviously this type of error correction requires duplex transmission lines and local data memory at the remote station. If only one-way transmission is available, the errors can not be corrected using the check-sum method but they can be detected and appropriate action taken at the receiving end. However, loss of data is inevitable. Error correction can be utilized at different hardware and software levels and can be of various types.

Proprietary error correction is used in many systems on the market. In these cases, the system operates over dedicated links and all the responsibility for transmission and error correction lies with the system. An example would be a digital radio link to remote stations that uses a manufacture's protocol for error correction. The protocol in this case would be built into the commercial product.

Standardized hardware error correction is another possibility. The hardware unit, where the data enters the digital link and where it comes out, has its own error correction built in. From the user's standpoint, it is assumed that no errors occur between the input and output of this hardware. The most common example of such hardware error correction is a telephone modem that uses internal, industry-standard error correction.

Computer networks use their own error-correction methods. When computers are linked with common computer network protocols like TCP/IP, error correction is built in from computer to computer. This is obviously the best solution, however it requires that the seismic remote stations operate quite sophisticated software. This is now the most common form of low level error correction.

Satellite data transmission links usually use forward error-correction (FEC) methods. FEC works on simplex links and doesn't require any retransmission of data blocks to correct errors. FEC is similar to check-sum error detection. By comparing the transmitted check-sum and that of the received data, corrections can be made. One drawback, however, is an increased data channel bandwidth due to a significant data overhead dedicated to error correction.

One should carefully consider the interplay between the error-correction system built into a seismic system with that of the particular communication equipment to be used.

### **8.6.8 Seismic data transmission and timing**

All digital data acquisition and transmission systems create a certain time delay or latency. This delay depends on the digitizer, the digital protocol used for transmission and the computer receiving the data. For this reason, nearly all digital field stations time stamp the data at the remote station and subsequent delays in the transmission have no effect on timing accuracy. However, there are also digital network designs where the timing takes place centrally. This can be done if the digital data arrives at the central site with a predictable or measurable time delay. The central computer must then time stamp the data when it arrives in real time and later correct it for the known transmission and digitizer delays. One advantage with this system is that only one clock is needed for timing the network. A further advantage is a simpler and less expensive remote station consisting only of a sensor and a digitizer. The disadvantage is that timing accuracy is not as good as with time stamping at the remote sites because time delays are known only with a limited precision and they may also vary in time. Also if the central clock or its synchronization with GPS time signals fails, the whole network fails. Most networks are moving towards time stamping at the station because GPS clock prices are now a small fraction of total digitizer costs.

### **8.6.9 Notes on dial-up phone lines and selection of modems**

Dial-up phone lines are sometimes proposed for seismic data transmission because they are readily available and apparently cheap. However, they have important limitations of which one must be aware. First, continuous seismic data transmission is not possible via dial-up lines. Second, their throughput is, in practice, frequently limited in spite of the high baud-rate capabilities of modern modems.

In practice, dial-up weak-motion networks based on phone lines can not 'digest' earthquake swarms and the numerous aftershocks after strong events. Therefore, they are an appropriate choice for low seismicity regions only. In addition, as they often do not function for several hours after strong events, due to either especially high usage of the public phone system or technical difficulties, they may not be the best choice for networks with the predominant purpose of giving seismic alarms. On the other hand, however, the USGS National Strong-Motion Program's dial-up network of about 200 stations (out of a total of 645 stations) ([http://nsmg.wr.usgs.gov/near\\_real\\_time.html](http://nsmg.wr.usgs.gov/near_real_time.html)) has successfully contributed to local ShakeMaps (<http://quake.usgs.gov/research/strongmotion/effects/shake/about.html>) in California since 1999. The data typically are downloaded, processed, and exported automatically to clients within 3-5 minutes after strong-ground shaking begins.

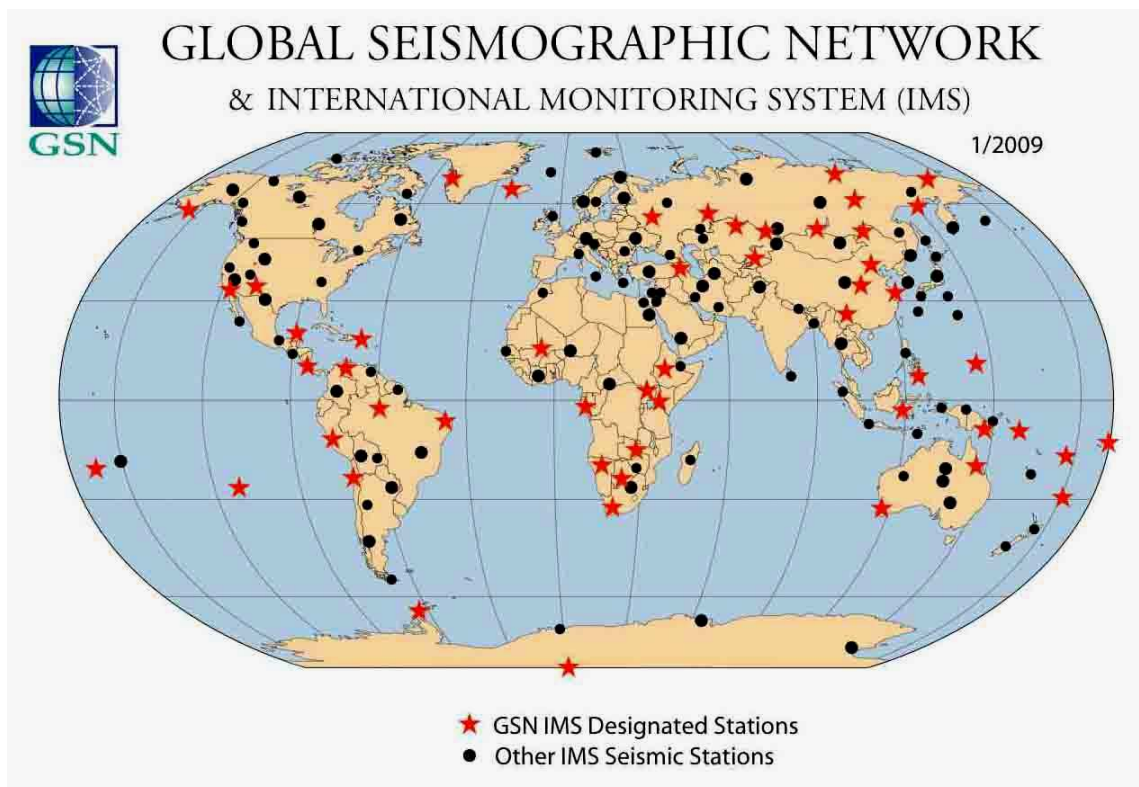
In many countries, public phone networks have specific properties and special 'tricks'. Therefore it is advisable to purchase modems locally. In general, it cannot be recommended to use modems for a new network.

## **8.7 Some network examples**

Along the lines described in the preceding section, some more examples are given of different types of seismic networks in operation, briefing both on their technical solutions and purpose.

### 8.7.1 International Monitoring System (IMS)

In recent years, a new global network, the International Monitoring System (IMS) has been set up aimed at monitoring the Comprehensive Nuclear-Test-Ban Treaty (CTBT) (see [www.ctbto.org](http://www.ctbto.org), Chapter 15 of this Manual, and Barrientos et al., 2001). The IMS is composed of seismic, infrasound, hydroacoustic and radionuclide stations. The seismic network consists of 50 stations designated as “primary”, mostly arrays (see Chapter 9), with real-time data transmission to the IDC of the CTBTO in Vienna, Austria. In addition there are 120 “auxiliary” stations that provide data on request to the IDC. Many of the auxiliary stations are members of the Federation of Digital Broadband Seismograph Networks (FDSN; see Fig. 8.4 and <http://www.fdsn.org>). The IMS network (Fig. 8.13) is currently the largest and most modern physical real-time network in the world. However, when requesting data from auxiliary stations, it works like a virtual network where the real-time network makes the detections and preliminary locations and then requests additional information from remaining stations for improving these preliminary findings.



**Fig. 8.13** Stations in the International Monitoring System (IMS).

### 8.7.2 Southern California Seismic Network (SCSN) (by E. Hauksson)

The authoritative earthquake-monitoring region of the Southern California Seismic Network (SCSN) extends across southern California, from the US-Mexico international border to Coalinga and Owens Valley in central California. The SCSN also reports on earthquakes in northern Baja California, Mexico, because these could possibly also cause damage in the United States. The region is home to almost 20 million inhabitants, including two of the ten

largest cities in the United States (Los Angeles and San Diego) and the two largest harbors (Los Angeles and Long Beach) in the Nation.

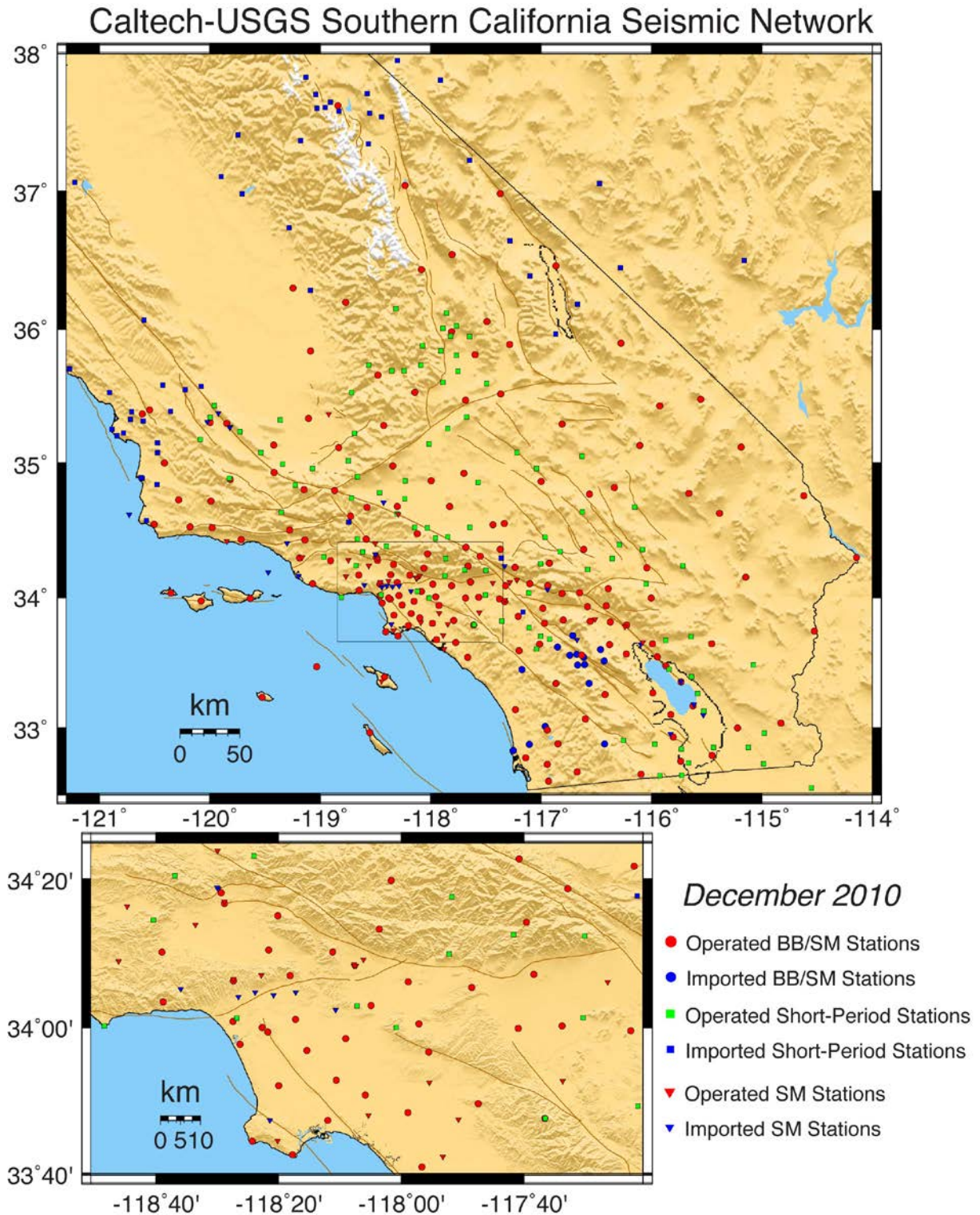
The SCSN operates about 300 seismic stations and receives real-time waveform data from an additional 125 stations operated by partner networks (Fig. 8.14). The SCSN seismic stations can be divided into three groups. First, the most modern group of 167 SCSN stations consist of broadband sensors, strong motion sensors, and high resolution dataloggers for digitization and data communications. The data from these stations are used for real-time determination of earthquake locations, magnitudes, moment tensors, and amplitudes for ShakeMap. The second group consists of about 35 stations that are equipped with only strong motion sensors and a datalogger. These stations provide amplitudes for ShakeMap and occasional arrival-time picks. The third group consists of 125 short-period stations that use legacy analog, frequency-modulated (FM) audible tone technology and Earthworm digitizers for data acquisition. Although, the amplitude data have limited dynamic range, these stations provide arrival-time picks for improved location and depth determination as well as coda duration for determining magnitudes of small earthquakes.

The data flow path from the remote seismic stations includes the on-site equipment, the data communications path, and a data acquisition computer at the central recording site. The last mile data communications to each site may use copper wire, spread-spectrum radios, or optical fiber. The main long-haul data communications include digital phone lines (frame relay or T1 lines) and digital microwave circuits. The waveform data are acquired by front-end processors that broadcast the data on a private local network. Many computers that are connected to this network, capture and process the waveforms or capture the data for archiving. For instance, a processing thread of Earthworm modules carries out the real-time picking of arrival-times, association, and hypocenter determination. A different thread harvests amplitudes and determines local magnitudes. Similarly, an earthquake-early-warning processing thread captures the waveform data to facilitate development of algorithms to estimate the size of the event quickly.

The Southern California Earthquake Data Center (SCEDC) archives and distributes all SCSN data and products. As the SCSN processes the data for detected earthquakes, outside researchers can also access all the data in the database, within minutes or as soon as the data are entered into the archive. The SCSN products include the southern California earthquake catalog, event gathers (triggers) of waveforms for nearly all events back to 1981 and for some events extending back to 1976, continuous broadband waveforms from 1999 to present, continuous short-period waveforms since 2008, and continuous strong motion waveforms since 2010. The SCSN and the SCEDC have maintained and published a catalog of earthquakes above magnitude 3.25 since 1932 and above magnitude 1.8 since 1980 with consistent magnitudes over the whole time (Hutton, *et al.* 2010). See also: <http://www.scsn.org>; <http://www.data.scec.org>.

A complementary information about the California Integrated Seismic Network (CISN) is contained in IS 8.4.

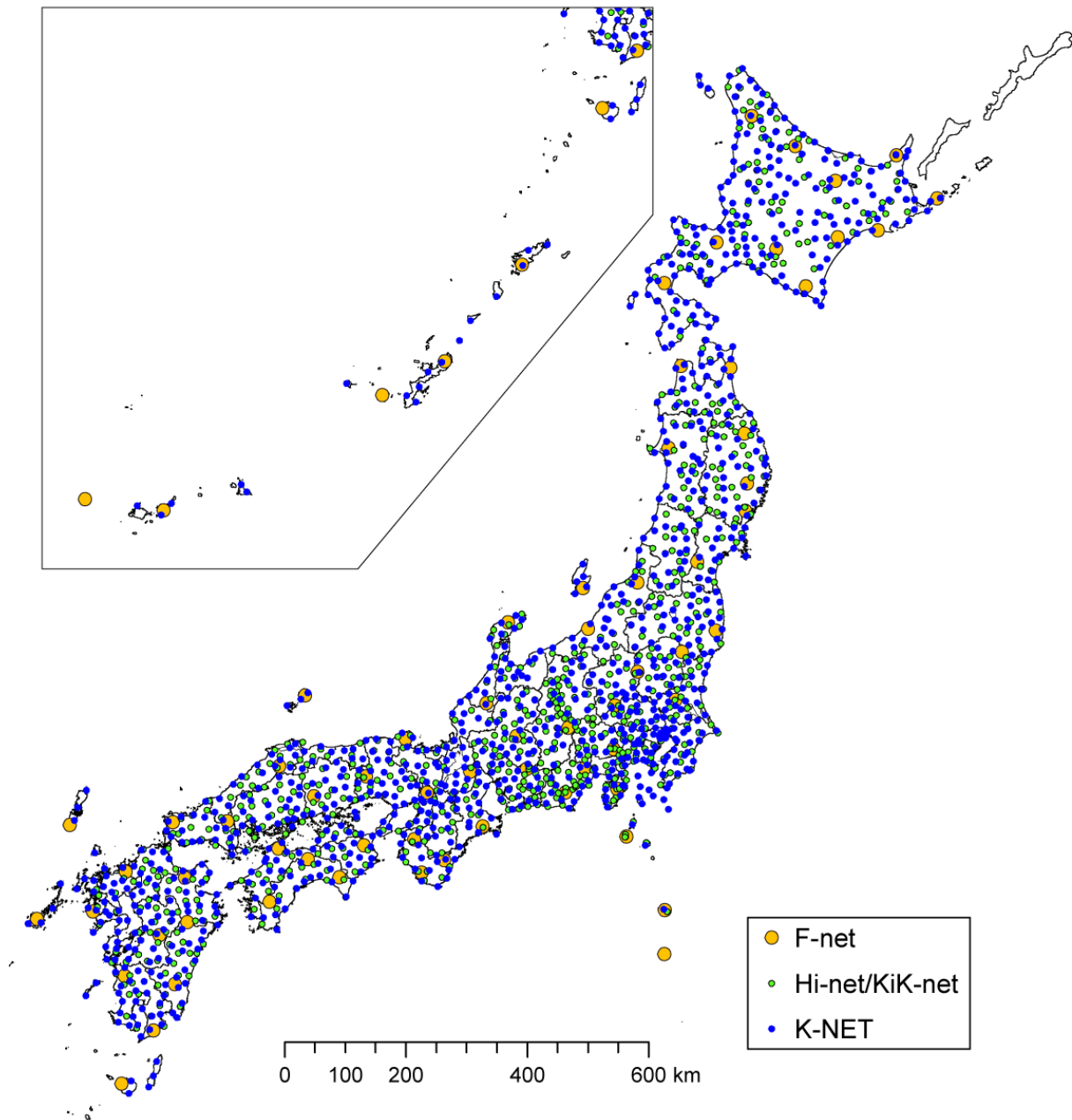




**Fig. 8.14** The Southern California Seismic Network (SCSN) in 2010. Both SCSN operated (shown with red and green symbols) and partner network stations (shown with blue symbols) recorded by the SCSN are shown. BB- broadband seismometer, SM- strong motion instrument.

### 8.7.3 Japanese seismic networks (Hi-net, F-net, and K-NET/KiK-net), operated by the NIED (by Shin Aoi)

Four seismic networks are presently operated by the National Research Institute for Earth Science and Disaster Prevention (NIED) in Japan (<http://www.bosai.go.jp/e/>) (Fig. 8.15).



**Fig. 8.15** Japanese Seismic Networks (Hi-net, F-net, and K-NET/KiK-net; status in 2010).

The first one is a high sensitivity seismograph network, named Hi-net, which comprises about 800 stations. At each Hi-net station a short-period, velocity-type seismograph is installed at the bottom of a borehole with a typical depth of 100 to 200 m (the deepest borehole reaches

3500 m depth). The second network is a broadband seismograph network, named F-net, which comprises about 70 stations. At each F-net station a broadband seismograph (STS-1 or 2) and a strong-motion sensor having a velocity-proportional response, are installed at the end of a horizontal tunnel of about 50 m length. Both Hi-net and F-net data are continuously transmitted to NIED via a TCP/IP network. The third network is a strong-motion seismograph network, named K-NET, which comprises more than 1000 stations. At each K-NET station an accelerometer is installed on the ground surface. The event-triggered data are automatically sent from each K-NET station immediately after the triggering. Another strong-motion seismograph network, named KiK-net, has its sensors installed in the same boreholes used by the Hi-net stations. Each KiK-net observation point is equipped with two accelerometers, one at the ground surface and one at the bottom of the borehole. Data collection procedure for the KiK-net is almost the same as that for the K-NET.

All data from these networks are freely available through the Internet. The Hi-net data are sent to the Japan Meteorological Agency (JMA) and used by the “Earthquake Early Warning” system, operated by JMA. Seismic intensities recorded by the K-NET, which is part of the nationwide seismic intensity network, are widely broadcast by television and radio stations. Many results, such as ground-motion prediction equations and underground structure models, have been obtained using a huge amount of data and related information from the above mentioned networks and largely contribute to a reliable seismic hazard evaluation. For more information, please see the webpages of the

- J-SHIS (Japan Seismic Hazard Information Station; <http://www.j-shis.bosai.go.jp/?lang=en>);
- NIED (<http://www.bosai.go.jp/e/>);
- Hi-net (High sensitivity seismograph network (<http://www.hinet.bosai.go.jp/>);
- F-net (Full range broadband seismograph network (<http://www.fnet.bosai.go.jp/top.php?LANG=en>);
- K-NET (Kyoshin strong-motion network (<http://www.k-net.bosai.go.jp/>);
- KiK-net (Kiban Kyoshin strong-motion network of vertical arrays (<http://www.kik.bosai.go.jp/>).

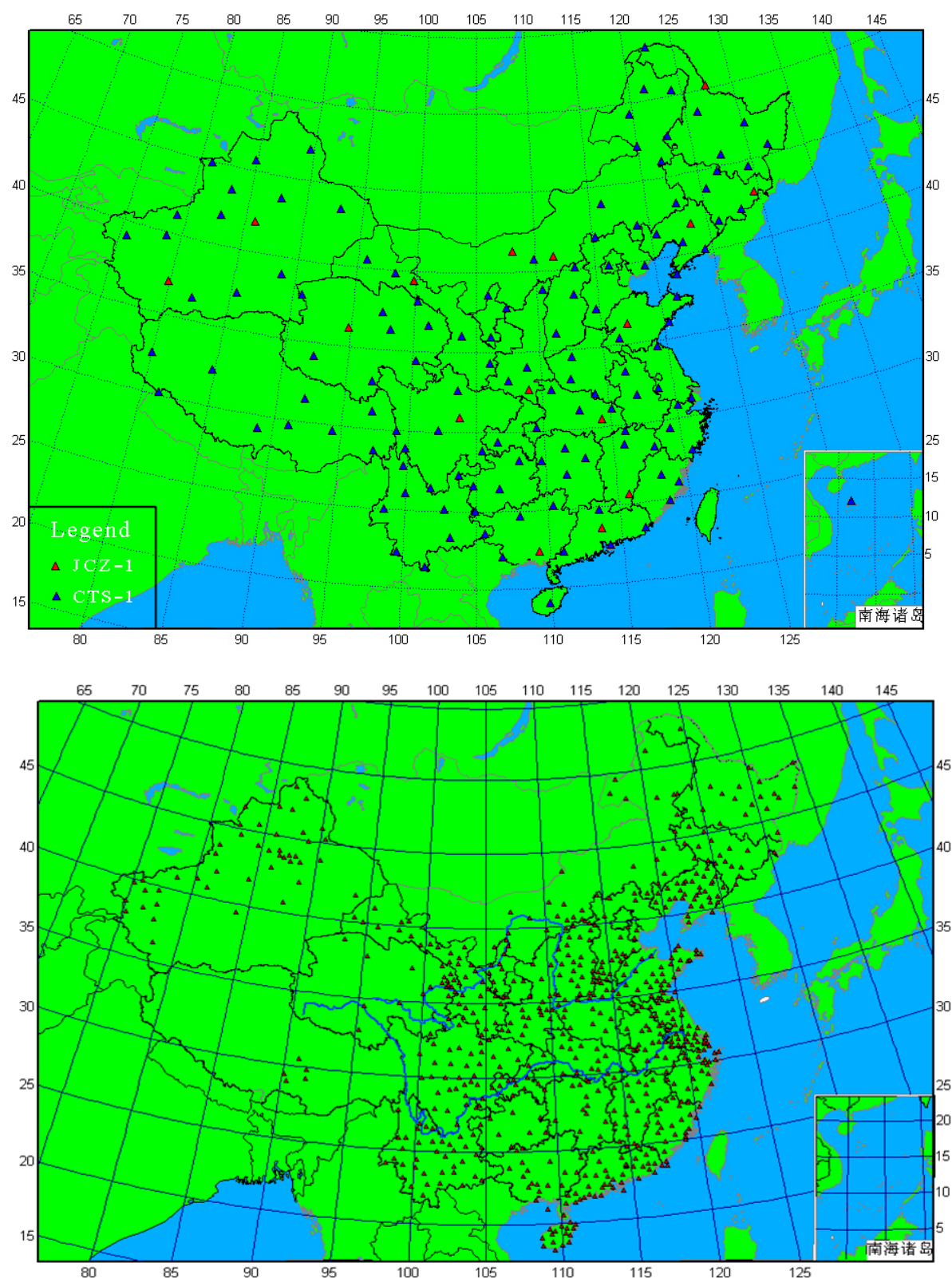
#### **8.7.4 China seismic networks** (by Liu Ruifeng and P. Bormann)

Since 2008 China has the following seismic networks (status as of December 2010):

- National Seismic Network with 145 stations, complemented by 2 small seismic arrays with 9 stations each;
- 31 Regional Seismic Networks with 792 stations in total;
- 6 local networks with 33 stations in total for volcano monitoring;
- 800 mobile seismic stations (600 for research projects, 200 for fast response deployments after strong earthquakes).

The National Seismic Network is an earthquake monitoring network with uniformly distributed seismic stations all over China. The average distance between seismic stations is about 250 km, except in the Qinghai-Tibet Plateau. Of the 145 stations of the National Seismic Network, 16 stations are equipped with Chinese made three-component very broadband seismographs of type JCZ-1, which have a flat velocity response from 360 s to 50 Hz and a flat acceleration response from 360 s to DC. At the other 129 stations, Chinese broadband seismometers of the type CTS-1 have been deployed, which have a flat velocity response from 120 s to 50 Hz. Detailed instrument parameters are given in the Data Sheet DS

5.1. The locations of these stations have been plotted in the top panel of Fig. 8.16 (top), with different symbols for these different types of instruments.



**Fig. 8.16 Top:** The China National Seismic Network and **Bottom:** the Chinese regional seismic networks.



Based on data from both national and regional seismic stations, CENC is capable of detecting  $M_L 2.5$  earthquakes in most areas on the Chinese mainland. In most of North China, northeastern China, central China, some parts of northwestern China and eastern coastal areas of China, events are detected down to  $M_L 2.0$  and in the key areas for earthquake surveillance and protection as well as in heavily populated major cities the detection threshold is as small as  $M_L 1.5$ . The combined use of GSN and CENC data greatly increased the speed of information release and location accuracy of earthquakes in China's border areas or outside of China.

CENC produces fast catalogues in several steps. A first fully automatic catalogue is available after 2 minutes, the primary catalogue (including interactive measurements) for Beijing region within 8 min, for other regions after 12 min, and the final catalog for Beijing in 10 min, for other regions in 20 min. Besides this, CENC compiles all catalog data, also of the regional network centers. These data are available for general use via <http://data.earthquake.cn> and further information is provided on <http://www.ceic.ac.cn>. The quick production of specialized complementary data such as fault mechanism and moment tensor solutions, reconstruction of rupture processes, the production of shake maps and so on lies in the responsibility of the Geophysical Institute of the CEA in Beijing, which also serves as a back-up center for CENC.

### 8.7.5 German Regional Seismic Network (GRSN) (by K. Stammer)

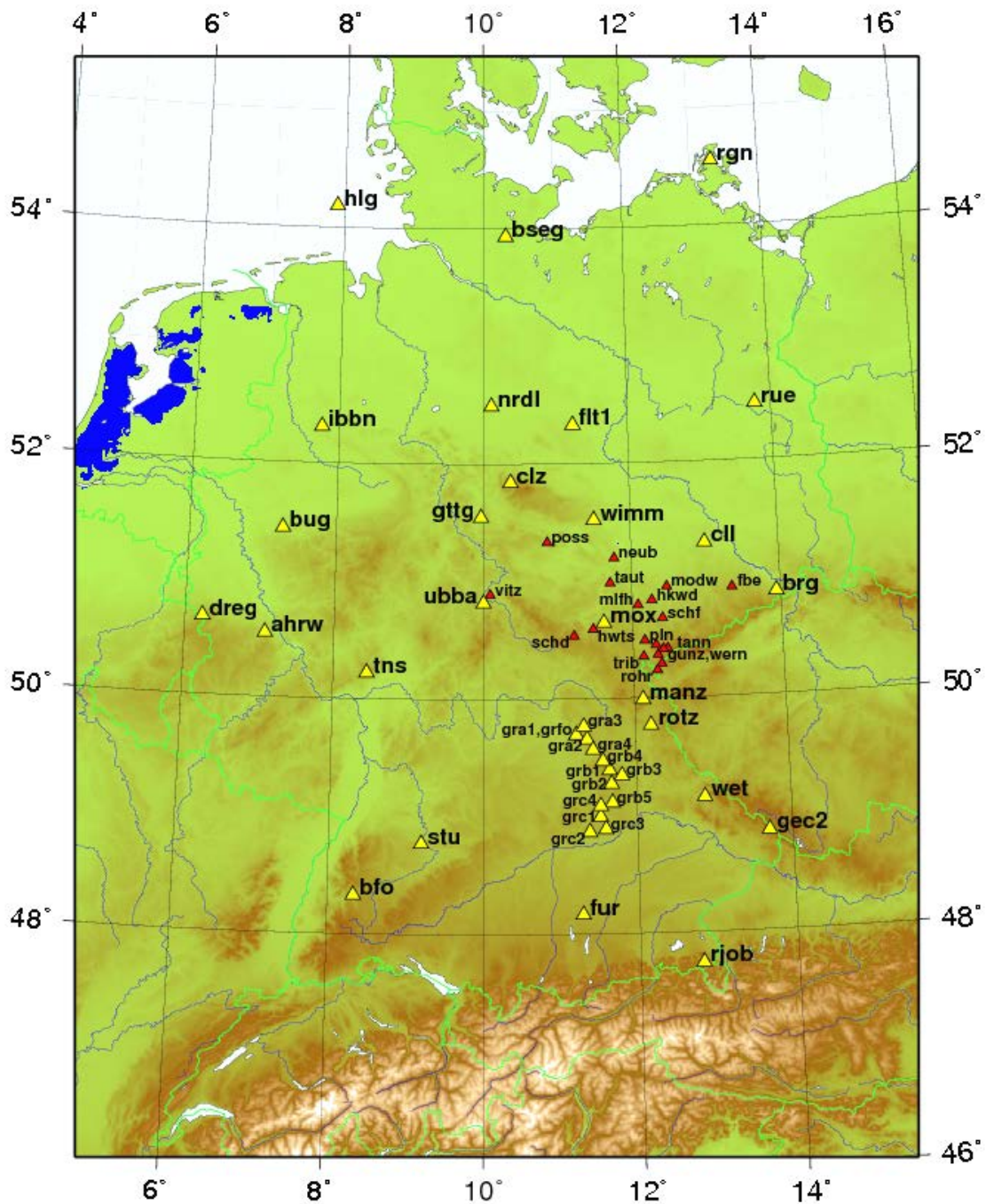
The German Regional Seismic Network (GRSN) currently consists of 27 stations and the Graefenberg array (GRF), consisting of 13 stations. They are the seismological broadband backbone of Germany. These stations have been operated continuously since 1976 (GRF), and 1991 (first stations of the GRSN), respectively. All stations are equipped with broadband sensors, in most cases with Streckeisen STS-2 sensors which have a flat, velocity-proportional response characteristic in the frequency range 8.33 mHz to 40 Hz. The GRF array was upgraded from STS-1 to STS-2 sensors in the year 2006. These GRF stations are operated by the Federal Institute for Geosciences and Natural Resources (BGR) whereas the GRSN sites are operated in collaboration of the universities in Germany, the Geoforschungszentrum Potsdam (GFZ) and the BGR.

The stations installed aim at monitoring and collecting high-quality data from regional and global seismic events as well as at recording and locating events with  $M_L > 2$  on German territory. All stations of GRF and GRSN are continuously recording, have a network connection via DSL or cell phone network and transmit their data in near-real-time to the Seismological Observatory of the BGR in Hanover using the SeedLink data protocol (except station GTTG which currently is using CD1 data protocol).

Since the local networks of the German states Saxonia and Thuringia also openly offer their continuous data via the Seismological Observatory of the BGR, these data are available there as well as those from the GRF and GRSN. These local networks operate currently 21 stations in total, mostly equipped with mid-period instruments Lennartz LE3D5s (flat response down to 0.2 Hz), only a few stations have broadband instruments like Streckeisen STS-2 or G ralp CMG 40T.

Fig. 8.17 shows the station sites of the Gr fenberg Array (GRF), the German Regional Seismic Network (GRSN) (yellow triangles) and of the associated local network stations (red

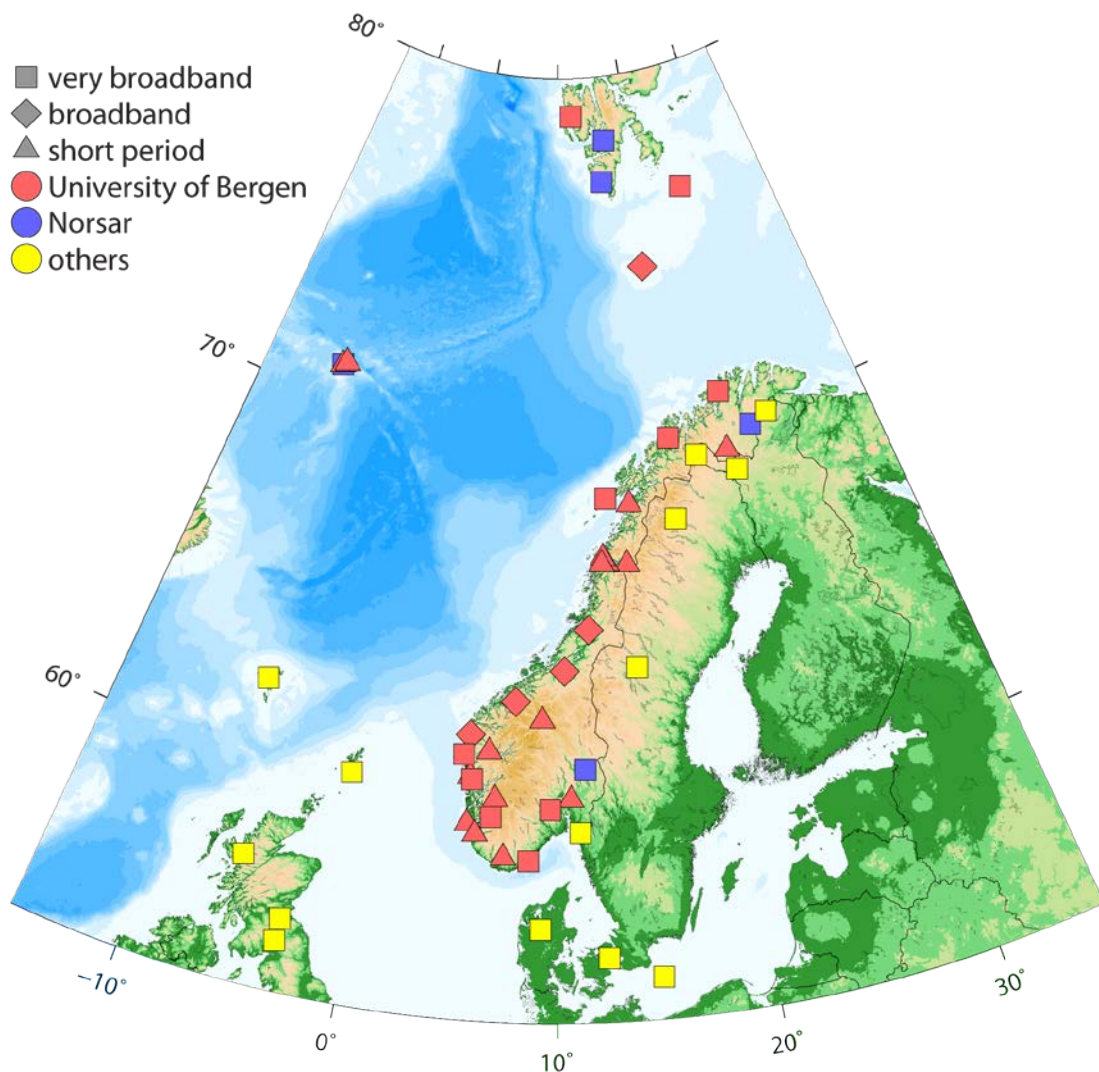
triangles). More details on the stations of GRF and GRSN are given on the websites <http://www.seismologie.bgr.de> and <http://www.szgrf.bgr.de>, which also provide information on available earthquake lists, bulletins, waveform data, station information, long-period and short-period dayplots of station recordings, the SHM analysis software manual as well as links to international and German seismic services and centers.



**Fig. 8.17** Map of the station sites of the Gräfenberg Array (GRF) and the German Regional Seismic Network (GRSN) (yellow triangles) and associated local network stations (red triangles). Figure from [http://www.szgrf.bgr.de/station\\_map.html](http://www.szgrf.bgr.de/station_map.html).

### 8.7.6 Norwegian National Seismic Network

The network is a typical internet based real-time network operated by a combination of different systems. The network consists of a total of about 50 stations (Fig. 8.19), where some of the stations belong to networks in other countries. The three stations on Jan Mayen are the last ones still connected in an analog sub-network.. Stations are equipped with either short-period or broadband seismometer, digitizer, recording computer and communications equipment, The majority of the data is continuously collected in the NNSN system from field stations with SeedLink. The acquisition at the field stations is done with the SEISLOG software (Utheim et al, 2001) where both triggers and continuous data are produced. The central network triggering is done with EarthWorm. Triggers from EarthWorm and SEISLOG are merged and moved to the final SEISAN data base using SEISNET. Likewise the continuous data is collected from all systems, homogenized and moved to SEISAN using SEISNET.



**Fig. 8.19** Norwegian National Seismic Network (NNSN) and associated stations. Red and blue NNSN stations are operated by the University of Bergen (UiB) and NORSAR respectively and yellow stations are contributions from other networks.

## 8.8 Seismic shelters

### 8.8.1 Purpose of seismic shelters and lightning protection

Civil engineering structures at seismic stations assure a good mechanical contact between seismic sensors and non-weathered, solid bedrock. They protect equipment from temperature, humidity, dust, dirt, lightning, and small animals. The shelter should also provide a good, low-resistance electric ground for sensitive electronic equipment and lightning protection, as well as easy and safe access for equipment maintenance and servicing. The well-engineered seismic shelter structure must also minimize distortion of seismic signals due to structure-soil interaction and man-made and wind generated seismic noise.

Seismic sensors require a stable thermal environment for operation, particularly BB and VBB sensors. With passive sensors, mass position may change too much and with active sensors, temperature changes result in an output voltage drift, which can not be resolved easily from low-frequency seismic signals. This can greatly reduce the signal-to-noise ratio at low frequencies or even clip the sensor completely. Also, many active sensors require mass centering if temperature slips below a few °C or the temporal temperature gradient is too large. Less than 0.5°C peak-to-peak temperature changes in a few days should be assured for good results when using broadband sensors. This is not a trivial requirement for a seismic shelter. Extremely demanding (usually non-vault type) VBB shelters can assure even better temperature stability. Peak-to-peak temperature changes as small as  $\sim 0.03^{\circ}\text{C}$  in two months (Uhrhammer et al., 1998) are reported for the very best shelters. Passive SP seismometers and accelerometers are much less demanding than BB and VBB seismometers with respect to the thermal stability of sensor environment. Many will work well in an environment with many degrees of temperature fluctuation.

Two vital, however often overlooked issues with potentially fatal consequences, if neglected, are lightning protection and grounding system. Lightning is the most frequent cause of seismic equipment failures. One needs to research the best lightning protection for each particular situation (lightning threat varies dramatically with station latitude, topography, and local climate) and then invest in its purchase, installation and maintenance. Several seismic networks have lost half or more of their equipment less than two years after installation because network operators simply neglected adequate lightning protection measures. A good, low-impedance grounding system keeps instrument noise low, allowing proper grounding and shielding of equipment and cables. It is a prerequisite for a good lightning protection system and is also absolutely required for an interference free VHF or UHF RF telemetry.

In some areas a light fence may be required around the vault to minimize man- and animal-made seismic noise and to protect stations against vandalism. The area covered by the fence may range from 5 x 5 m to 100 x 100 m, depending on several factors, e.g.: what kind of activity goes on around the site; the population density in the vicinity; the ground quality; natural seismic noise levels; and the depth of the vault. Note that fencing often represents a significant portion of the site preparation costs.

Inadequate site preparation and seismometer placement can easily wipe out all the benefits of expensive, high-sensitivity, high-dynamic range seismic equipment. For example, thermal and wind effects on a shallow seismic vault located on unconsolidated alluvial deposits instead of

bedrock can make broadband recording useless. It is pointless to invest money in expensive seismic equipment only to have its benefits wasted because of improper site conditions.

### **8.8.2 Types of seismic shelters**

The three main types of seismic shelters are:

- surface vaults which are the least expensive and by far the most frequently used, however they suffer the greatest level of natural and man-made seismic noise (see 7.4.2);
- deep vaults placed in abandoned tunnels, old mines or natural caves which are usually the best locations with respect to the price/seismic-noise-performance ratio, however, they may not be available and sometimes require extensive cabling, which can increase their cost (see 7.4.3);
- borehole seismic stations with depths from 10 to 2000 m are the best choice from the perspective of seismic noise. Improvement of the signal-to-noise ratio of up to 30 dB in ground velocity power density at about 0.01 Hz can be obtained by a 100-m deep hole. For high frequencies above 1 Hz the greatest gains in noise level reduction are realized within the first 100 m of hole depth. Wind-generated high frequency noise can be attenuated as well, however a complete shielding from it is possible only with a very deep borehole (Young et al. 1996). Boreholes are expensive. They may cost from US\$ 5,000 to US\$ 200,000 for the borehole itself, plus the cost of borehole type sensors, which are significantly more expensive than regular surface sensors. Boreholes are used principally in regions covered entirely by alluvial deposits where sites with good bedrock outcroppings are not available; or for the most demanding research work requiring low tilt-noise in horizontal component BB and VBB installations (see 7.4.5).

Shallow boreholes with a depth from a few meters to 15 m are sometimes used instead of surface vaults for pure economic reasons. A 15-m deep surface vault in a difficult terrain may cost more than a shallow borehole of the same depth. Seismic noise improvement in such shallow boreholes is negligible.

In terms of network cost, it might be cheaper to increase seismic station density to achieve a desired detection level rather than install a few borehole systems

### **8.8.3 Civil engineering works at vault seismic stations**

Today, seismic stations are most often in the ground vault form. The massive, solid concrete "seismic piers", traditionally found in old seismic observatories, are no longer built. Above-ground buildings or shelters are not desired at all. In fact, above-ground structures are far less suitable than underground vaults because of potential structure-soil interaction problems as well as wind generated seismic noise caused by the above-surface structural elements. (Bycroft, 1978; Luco et al., 1990). Also, sufficient thermal stability of the environment is much easier to achieve in an underground vault. If small buildings of any kind already exist at the selected location, make sure the seismometer vault is placed far enough away to minimize wind-generated noise as recommended already in the old Manual of Seismological



Observatory Practice (Willmore, 1979) (see IS 7.3). The structure of the vault should be light and above-ground parts kept to a minimum, creating as little wind resistance as possible.

Surface seismic vaults usually measure between 1 and 2 m in diameter, depending on their depth, the amount of installed equipment and the desired ease of maintenance. They are from 1 to 10 m deep, depending on the depth, quality, and weathering of bedrock at the site. Round or rectangular cross sections are equally suitable. Examples of their design are given in Figs. 7.39 and 7.40.

## **8.9 Establishing and running a new seismic network**

### **8.9.1 Planning and feasibility study**

#### **8.9.1.1 Goal setting**

The very first step toward establishing a new seismic network is understanding and setting the network's goals. These goals can differ significantly (see Tab. 8.1 in 8.3.5). The same applies to the seismic system requirements. Also, just as each country has unique seismicity, seismotectonics and geological formations, so every seismological project has unique contextual combinations that one must consider in order to find the optimal system design for that project.

Several issues must be addressed:

- the user's interests in ranked order: local seismology (epicentral distances < 150 km), regional seismology (epicentral distances between 150 and 2,000 km), and/or global seismology (epicentral distances > 2,000 km);
- the main purpose of setting up a network is usually either to monitor a region's general seismicity or to perform special studies (monitoring of special seismotectonic features, of important civil engineering structures, of engineering and/or nuclear explosions, of man-induced seismicity, etc.);
- the relative importance to the project's alarm function for civil defense purposes: Is the seismological research aimed at the long-term mitigation of the country's seismic risk or at the scientific research of the Earth's deep structure.

Many countries that have little or no seismic equipment should initially consider buying a system to monitor the region's general seismicity. They should expect the new system to help mitigate the region's seismic risk over a long period of time. Nevertheless, even for a project of such a well-defined scope, several questions must still be answered, including the country's needs as well as its financial, personnel, and infrastructure capabilities:

- how big is the region to be monitored?
- what is the seismicity level in the region?
- what is the institution's existing level of seismometry knowledge, and what are its resources for improving that knowledge?
- what is available in terms of communication infrastructure?

- how much money is available to establish the system?
- how many staff are available, per year, to operate and maintain the system, and to support research work using the system's data?

Having realistically quantified the above facts, one can then begin shopping for a seismic system that meets those criteria. There is always a trade-off between desires and reality. This procedure ensures that the new network will perform successfully in the existing environment, if carried out realistically.

If there are few or even no seismology experts available in the country, it is advisable to get help from consultants in the international academic world who are independent of commercial interests. In this early phase, focus on your country's specific socioeconomic needs and seismic awareness, and do not worry too much about specific equipment. Wait until the later phases of network design to contact sales and system engineers from seismic equipment manufacturers for help in defining the technical details of your system.

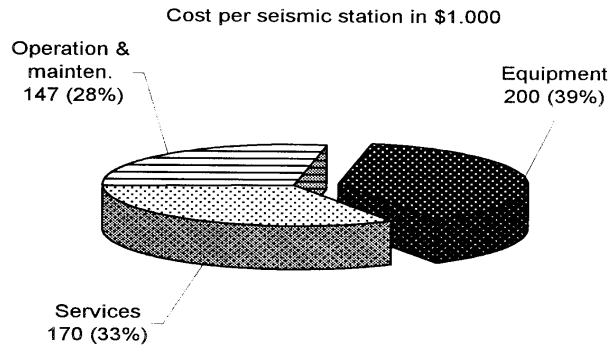
### **8.9.1.2 Financial reality**

Often, newcomers to seismology do not know how to allocate their finances to obtain the optimal seismic network design. Too often they spend the majority of their network funds purely on purchasing equipment, even though an identically important expenditure is required for proper operation of this complex equipment. To make sure one has correctly prepared for the purchase of seismic network equipment, one's budget must include money for the following:

- a feasibility study that examines potential network layouts, site selection, and potential seismic systems;
- preparation of remote stations;
- installation of the central-recording site;
- purchase of the network equipment;
- cost of installation, training, maintenance, and long-term support;
- cost of salaries and training for the new scientific and technical personnel usually required;
- network operation costs, including personnel, data transmission, data processing hardware and software, printing, backup storage, consumables, and spare parts;
- network servicing and maintenance cost.

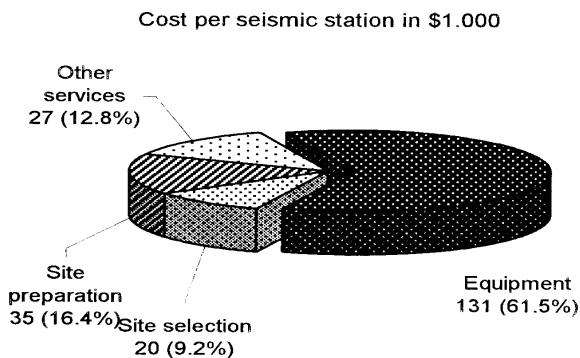
The five figures on the following pages show examples of funding proportion among several different established seismic network projects. The numbers in the figures show the amounts allocated to different tasks (normalized per single station), both in thousands of US dollars and as a percentage of the project's total cost.

Fig. 8.20 shows an approximate cost distribution (per station) for establishing and operating the global seismic network (GSN) during five years, according to the IRIS plan 1990-1996. The IRIS consortium is composed of universities in the USA and other countries with a research program in seismology. Not only did this network use the most demanding and expensive equipment available, expensive site preparation and worldwide maintenance were often required which increased the cost per station.



**Fig. 8.20** Cost distribution of establishment and 5-year operation of a global seismic network (GSN) station. Number in ( ) is percentage of the project's total cost.

Fig. 8.21 shows details of the IRIS GSN system's establishment costs (excluding all operations costs; again, costs are averaged per station). Surface vault seismic stations are considered only. IRIS constructed many of the sites of GSN network as deep, expensive borehole installations. Even if they are not taken into account in this figure, IRIS still allocated substantial funds for the vaults and to tasks other than equipment buying.

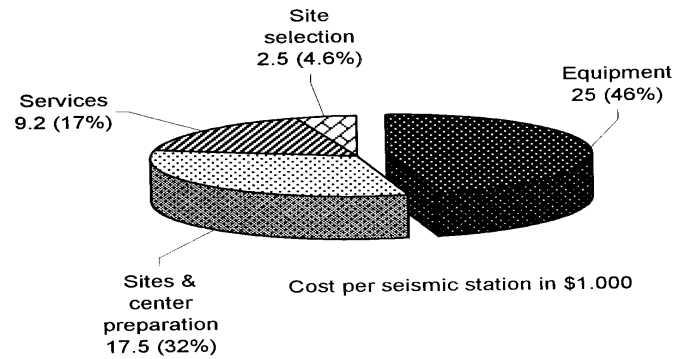


**Fig. 8.21** Cost distribution of establishment of IRIS GSN surface vault seismic stations. In this and the following figures the number in ( ) is percentage of the project's total cost.

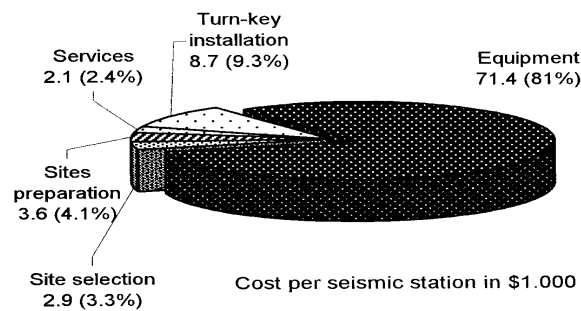
Fig. 8.22 shows a distribution of the finances which a developing country spent to establish a reasonably large seismic network, using analog RF telemetry. The country's significant investment in services (21.6%) paid for training at the factory and during installation, as well as one year of the manufacturer's full-time engineer support. These expenditures were critical for the successful start-up and operation of this network.

Fig. 8.23 shows a negative example of cost distribution, for a small, yet technologically demanding seismic network. Note the small amount invested in tasks other than equipment-purchases, particularly the site preparation works; 4.1% is surely not sufficient, making it difficult to believe that these sites could provide ample working conditions for such demanding sensors as very broadband (VBB) STS1 and STS2 seismometers. The relatively high amounts spent for services (9.3% for installation) came mostly because the purchasers desired a turnkey type of system. With no experiences in seismometry, the chances of efficiently using the installed equipment seem small.



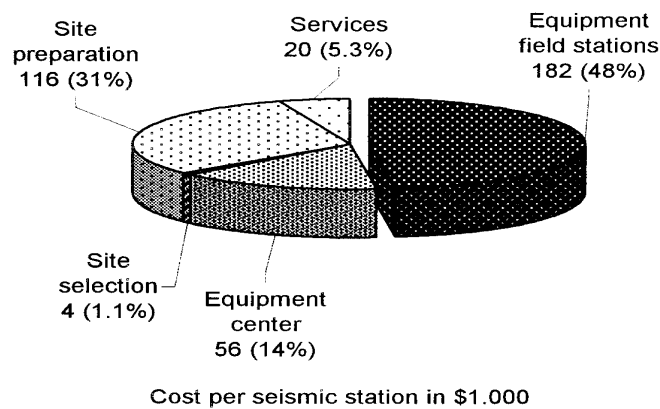


**Fig. 8.22** Cost distribution of a relatively large national seismic network with 20 SP seismic stations, strong-motion instrumentation, and analog FM telemetry.



**Fig. 8.23** Cost distribution of a small, technologically high-end seismic network with an inappropriate allocation of funds.

Fig. 8.24 shows another example of a national seismic network installed in a large country and using high-end technology and duplex, digital telemetry system. But again, despite the network's size, the most modern equipment, and the central-recording equipment for two centers, the country only invested about 60 % of its total project funds in the equipment. The other half of the money was spent on follow-up services, including a great deal of training and two years of full-time engineer support provided by the equipment's manufacturer.



**Fig. 8.24** Cost distribution of a very large national, high-end technology, duplex-digital RF and phone-line telemetry seismic network with two central-recording centers.

The funding distributions shown in Figs. 8.20 through 24 are approximate and for illustration purpose only. Generally, the prices of seismic equipment are somewhat lower today. Actual conditions (including the type of network, the level of existing local technical knowledge, local labor prices, and the type of seismic site preparation required) will change from country to country, thus significantly influencing distribution of the funds. Regardless, the main message of these figures stays the same: one should not spend almost all the allocated funds on equipment. Despite deviations and the differences in absolute cost, these figures seem to indicate that the percentages of the total cost for each task remain nearly the same from network to network. As a rule, one should allocate at least one third of the money for a feasibility study, for establishing the proper working conditions, and for gaining the seismic expertise necessary to exploit the purchased equipment.

### **8.9.1.3 Basic system engineering parameters**

Once the goals are clear and the funds properly allocated, one has to clarify the entire project's interrelated seismological and technological aspects. Attention should be paid to:

- the size and the layout of the proposed seismic network (this should affect the choice of the type of transmission links for transmission of seismic data from the remote stations to the center);
- the seismicity level to be monitored - in other words, the amount of data one will deal with (this should affect data transmission equipment, central processing site's real-time and offline capabilities, whether the system will need continuous or triggered data recording capabilities, if and what type of trigger algorithm it will use, the type of data archive system; this should also affect the partitioning between weak-motion and strong-motion equipment);
- how accurate and where one wants the network's central-recording site to be located (this will affect the number of stations and the network's layout);
- the importance of the new system having alarm capabilities for civil defense purposes and the desired alarm response time (this should influence which data transmission links will be chosen, as well as how much real-time processing power will be needed at the central-recording site);
- the amount of technical reliability one expects from the system (this should affect the choice of data transmission links, how much hardware system redundancy one can afford for mission critical applications, like auto-duplicating disk drives, tandem computers, etc., as well as decision between 'office-grade' and industrial-grade computers);
- the desired robustness of the system in terms of functioning throughout damaging earthquakes (this should influence the selection of data transmission links, of power backup utilities for the remote stations and the central-recording site, and last but not least, of seismic vulnerability of the building that houses the central processing site).

After reasonably assessing these aspects and making a decision for each unique situation, one can then create a rough system design and begin selecting equipment that best matches these goals. Obviously, certain tradeoffs will need to be made.

#### **8.9.1.4 Determining the layout of a seismic network**

Determining a layout for one's seismic network requires two steps: 1) determining the total number of stations required and their approximate locations, and 2) determining the final station locations. Since the first stage closely relates to the goals of the network and available funds, the network operator should delineate how many stations he requires and can afford to set up, and where approximately they should be located. Since the second stage typically requires knowledge of seismometry, seismo-geology, data transmission technology (if applicable), and seismic equipment capabilities and limitations, the operator, if not having the experience, may want to consult with others having the experience.

#### **8.9.1.5 Number of stations in a seismic network**

The number of seismic stations should be based on the goals of the network, the size of the network, and, of course, on the available funding. For space reasons we will not go into details on the minimum number of stations that are technically required for a given seismological goal, but following there is a short overview.

For determination of an event location (based on phase readings), the theoretical minimum is four independent measurements, such as three P-arrival times and one S-arrival time. However, remember that such results, due to their uncertainty, usually have little value. For a more accurate determination of location, six stations acquiring good records of an event should provide scientifically credible evidence of an event's location, and ten to fifteen stations with good azimuthal coverage all around the source (see criteria outlined in Table 1 and 2 plus Figure 1 of IS 8.5) and acquiring good quality records of the event should provide an acceptable basis for more sophisticated studies of the earthquake's source properties. IS 7.4 describes a program which permits detectability and earthquake location accuracy modeling of smaller size national networks. Examples of theoretical network optimization with respect to event location and other parameter determinations are presented and critically discussed in IS 8.7.

Larger countries or regions will require a greater number of stations, unless, of course, their interest is only in the strongest earthquakes. Yet, for networks covering a large region, large epicentral distances often prevent the triggering of distant stations, or the earthquake signals get buried in the seismic noise. Thus the total information available for a given event, unless it is a strong one, typically comes from only a portion of the total network. How one can assess under such varying conditions, with respect to precision-located "ground truth" GT5 criteria (e.g., Bondár and McLaughlin, 2009), the quality of an operational authoritative location scheme applied in the real-time event information service of the European Mediterranean Seismological Center (EMSC) is described in IS 8.5. How ground truth reference events, suitable for network performance and location quality evaluation, can be identified and collected by way of own high precision network locations is described in IS 8.6. And Murphy et al. (2005) present an example of the calibration of stations of the International Monitoring System (IMS) aimed at improving seismic event location.

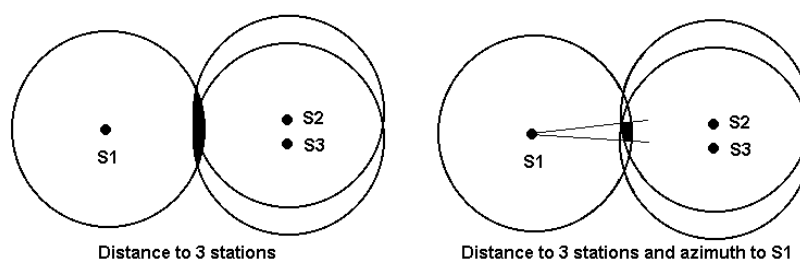
Note that seismic researchers do not care much about the total number of stations in a network; what counts is the number of stations in the network that adequately record a given event ('adequately record' means that they triggered data acquisition, the records are not saturated and that the records have a high signal-to-noise ratio). Moreover, waveform analysis of digital, high dynamic range, three-component records leads to good results already with fewer stations. In principle, even a single well calibrated 3-component station can determine, via multi-phase phase identification and polarization analysis, the epicenter, source depth (if below the crust), origin time and magnitude of earthquakes with reasonable accuracy using global 1-D models and correcting for systematic biases (if any), that may depend on azimuth and distance or source region (see Chapter 11, section 11.2.6.2. and EX 11.2). For more details on location procedures and related problems, both for networks and single stations, see IS 11.1.

#### 8.9.1.6 Layout of a new seismic network

Although the spatial distribution of the stations in a seismic network is very important for the network's capabilities of event determination, due to limited space we will only give a few, brief recommendations. For seismic arrays and their special location procedures and performance see Chapter 9.

On a map, subdivide the region to be monitored into a series of reasonably irregular triangles having approximately equal areas. Avoid very narrow, long triangles. Avoid thinking in rigid patterns, such as locating the stations into perfect triangles, circles or straight lines, because such rigidity may result in "blind spots" - that is regions with poor event location determination. The corners of these triangles are the approximate points where one will try to locate seismic stations. Take into account any existing seismic stations in neighboring countries or regions as well. If there are none, push some of the seismic stations as close as possible to the borders of the region being monitored.

The geometry of the network will determine the accuracy of location in different directions, and a reasonably regular grid will give most uniform location accuracy. The worst configuration is a network with stations that are aligned (see Fig. 8.25 as an example).



**Fig. 8.25** Network geometry of aligned stations. The figure to the left shows three stations (S1, S2 and S3) almost aligned in the x-direction (left - right). The event has been located by using the distances to the three stations, and the shaded area in the middle gives an indication of the area within which the epicenter can be found. The figure to the right shows the same situation except that an azimuth determination has been made with three-component records at station S1. This limits the y-direction within which the epicenter can be located and thus reduces the epicenter error.

It is advisable to have realistic expectations concerning the earthquake depth determination based on phase readings. Previous studies (e.g., Francis et al., 1978; Uhrhammer, 1980; and McLaren and Frohlich, 1985) have shown that the accuracy of focal depths for shocks occurring in the vicinity of a seismic network is primarily a function of the geometry of the network, the number of the P- and S-phase arrivals read, and the adequacy of the assumed velocity model. Depths are generally more accurate for earthquakes where the distance from the epicenter to the closest station is less than the calculated focal depth for events located within the network or on its periphery. The accuracy of focal depths usually increases as the number of picked S-phase arrivals increases; however, systematic S-phase timing errors (due to mistaken identification of a converted phase as S) or "bad" S picks can degrade the focal depth estimation accuracy by several kilometers even when the azimuthal coverage is good (Gomberg et al., 1990). Estimate the depth of the shallowest events for which good depth control is desired then make sure that the average distance between stations in the seismic network does not exceed twice that depth. The latter is admittedly a tough requirement, especially in the large regions and in the regions where the events are typically shallow! Only a few small countries and practically none of the larger countries can afford such a dense network.

Yet, one can still temporarily afford to make the network denser in places. Buy a few portable seismic stations and then temporarily install them in any sub region of particular interest at the time. For example, such temporary networks are regularly established to perform aftershock studies in the epicentral region immediately after a strong event. At least for a time, this will drastically increase the seismic network's density in the region of interest, allowing the determination of much better locations, depths, and focal mechanisms. Such studies can be done with low-cost portable instruments since the main purpose is to get more phase readings.

Have realistic expectations also about the system's earthquake epicenter determinations. For events outside the seismic network, expect large errors in determining epicenters. Generally, do not expect reliable determination of events, unless the "seismic gap" (the largest of all angles among the lines connecting a potential epicenter with all the stations in the network that recorded the event) is less than 180 degrees. Thus, to increase the accuracy of epicenter determinations, especially for the events outside the seismic network, one needs to include data in the analysis from seismic stations in neighboring countries, as well as from any other available national or international sources. It is common practice to share near real-time data across the borders and integrate the data into a virtual seismic network. Acquiring this wider database is usually necessary for determining reliable event locations on the border or outside any seismic network.

### **8.9.2 Site selection**

The matter of seismic site selection is too often not given sufficient depth of study and attention in spite of the fact that a weak-motion seismic network can only have a high detection threshold if the sites have satisfactory noise levels, no matter how technologically advanced and expensive equipment is. If seismic noise at the sites is high, all or a part of the benefits of modern equipment with large dynamic range are lost. If an excessive burst or spike-type, man-made seismic noise is present, high trigger thresholds and therefore poor event detectability will result. If stations are situated on soft ground, the VBB or even the BB

recording can be useless and SP signals may be unrepresentative due to local ground effects. If the network layout is inappropriate, some event locations may be inaccurate or even impossible. For good results, many factors at the sites must be taken into consideration. A professional site selection procedure is therefore essential for success of any seismic network. There can also be specific reasons (e.g., political) why a station may have to be deployed in a particular place, but also then it may be possible to select the “better” site.

Generally, it is best to begin the process of site selection by choosing two to three times as many potential sites as one actually plans to use. Then each site is studied to see which sites meet as many of the criteria as possible. Gradually, one will eliminate the poorest sites and get down to the number of sites required plus two or three. By comparing the results of computer modeling of a few of the most likely network layouts (see IS 7.4) one will be able to make an informed decision about the best network.

Note that one should not rely too much on algorithms designed to optimize seismic network configuration (e.g., Kijko, 1977; Rabinovitz and Steinberg, 1990). This is because the theoretical optimum configuration can hardly ever be realized nor their predicted theoretical potential information gain be exploited under real conditions. Stations often can not be installed at the recommended locations due to factors such as inaccessibility, poor ground conditions, proximity of strong noise sources, lack of required power, or unavailable communication link.

On the other hand, these programs may be of help selecting the best of a few realistic network configurations (e.g., Trnkoczy and Živčić, 1992; Hardt and Scherbaum, 1994; Steinberg et al., 1995; Bartal et al., 2000). For an existing network, they could help decide how best to improve the network by adding new stations or which stations, if removed, would cause least harm to the network. Keep in mind, however, that the best configuration for locating earthquakes may not be optimal for source mechanism determinations, tomographic studies or other tasks (Hardt and Scherbaum, 1994). These various aspects of network optimization and performance assessment are richly illustrated and commented in more detail in a PPT tutorial, annexed as IS 8.7.

Here we will present only the basic steps of the site selection procedure. More details on this issue can be found in Chapter 7, sections 7.1 and 7.2.

The site selection procedure encompasses office and field studies. Off-site, “office” studies are relatively inexpensive and therefore the first ones to be performed. From an office, one can study maps and contact local authorities to gather information about potential sites. The first step is defining the geographical region of interest. The next step is to gather and examine existing geological faults, seismotectonic features, and all available information about seismicity in the area. If the main goal of the new network is monitoring general seismicity in an entire country, this stage is, of course, simpler. Then prepare a simplified map of regional seismo-geologic conditions showing the quality of bedrock. The rule is: the higher the acoustic impedance (acoustic impedance is the product of the density and the velocity) of the bedrock, the less the seismic noise and the higher the maximum possible gain of a seismic station. Next, study the topographical aspects of the possible locations. Moderately changing topography is desired. To study man-made and natural seismic noise sources in the region, one should evaluate road traffic, railway traffic, heavy industry, mining and quarry activities, agricultural development of the region, and any other sources of man-made seismic noise around the potential sites, along with the natural sources like oceans and lakes, rivers,

waterfalls, animals, etc. (see IS 7.3). Much of the information we need can be found on maps, geographic information systems or obtained by asking questions of local authorities.

Real-time communication is a requirement in most modern seismic networks and will add some constraints to site selection. Communication is typically based on Internet technology for radio, satellite, mobile phone and fixed line links. Radio and satellite links are the most flexible as they can be deployed almost anywhere. Mobile phone services and coverage vary strongly, but may be an attractive option. Fixed lines such as ADSL are generally inexpensive and provide high data rates, but are not available anywhere. If the new network is a radio frequency (RF) telemetry system, one has to correlate RF data transmission requirements with seismological requirements. Topographic profiling of RF paths based on topographical maps is performed. The next section "VHF, UHF and SS radio-link data transmission study" explains why this is highly recommended.

If one plans to use main power, the availability of main power lines and the distances to which new lines would have to be laid must be checked. The alternative is batteries, preferably charged by solar panels. It is also very important to research land ownership, animal habitats, land protection and future land use plans for the potential sites. It makes no sense to undertake extensive studies if one will be unable to use certain sites because of property ownership issues, endangered or protected animal species issues, or if it appears that future development will make the site unsuitable for seismic stations.

The climate at the sites also influences site selection and preparation. Temperatures, wind, precipitation, insulation data (for solar-panel powered stations), lightning threat, etc. may all influence site selection.

Field studies are the next step in the site selection process. Expect to make several visits to each site. A seismologist familiar with seismic noise measurements, a seismo-geologist, and a communications expert (if we are considering a telemetry network) should all visit each site. They should verify the ease of access to the site, search for local man-made seismic noise sources, which may not be apparent from maps, perform seismic noise measurements (of at least a few days duration), study the local seismo-geological conditions at the site, investigate the local RF data transmission conditions (if applicable), and on site verify power and phone line availability.

Local geology should be studied to determine its complexity and variations as well as seismic coupling between local seismic noise sources and the potential station site. To the extent possible, uniform local geology is preferred for seismic stations. The degree of weathering that local rocks have undergone is another important parameter, although it can give an unreliable estimate of the required depth of the seismic vault. The ideal approach for high-quality site selection is to make a shallow profile at each potential site to make sure the vault will reach hard bedrock. If this is too costly, then expect surprises when you begin digging seismic vaults. Many times it is a matter of almost pure chance what one might run into. Note that in some areas it will not be possible to reach bedrock.

After all these studies one ends up with two or three potential sets of the best suitable seismic stations. The resulting network layouts are then studied for the best network performance possibly by computer modeling. By comparing the results, one will be able to make an informed decision about the final seismic network layout.

### **8.9.3 VHF, UHF and SS radio-link data transmission study**

#### **8.9.3.1 The need for a professional RF network design**

The most frequent technical problems with radio-frequency (RF) telemetry seismic networks originate with inadequately designed data transmission links. Therefore we are discussing this topic separately. For more detailed description see 7.3.

The design of RF telemetry links in a seismic network is a specialized technical matter, therefore guessing and "common sense" approaches usually cause problems or even complete project failure. There are quite a few common misunderstandings and oversimplifications. The amount of data that must be transmitted and the degree of reliability required for successful transmission of seismological data are frequently underestimated. The significance of "open line of sight" between transmitters and receivers as a required and sufficient condition for reliable RF links is misunderstood. Frequently, over-simplified methods of link verification are practiced. However, the real issues in the RF link design and link reliability calculations are: the frequency of operation, Fresnel ellipsoid obstructions by topographic obstacles, the curvature of the Earth, the gradient of air reflectivity in the region, expected fading, potential-wave diffraction and/or reflections, time dispersions of the RF carrier with digital links, degradation of signal strength due to weather effects, etc. All these are specialized technical issues.

To prevent failures, a professional RF survey in planning a new seismic network is strongly recommended. It includes the calculation of RF links based on topographical data and occasional field measurements. A layout design based on a professional RF survey can significantly increase robustness of the radio network. The survey will:

- determine the minimum number of required links and RF repeaters in the network. Note that, in most designs, every RF repeater obviously increases the probability of link-down time and the price of the system;
- determine the minimum number of licensed frequencies required;
- determine the optimal distribution of RF frequencies over the network, which minimizes the probability of RF interference problems;
- chose in a less polluted RF space in the country;
- determine the minimum antennae sizes and mast heights, resulting in potential savings on antenna and antenna-mast cost.

The cost of a professional RF survey represents generally a few percent of the total investment. We believe that the combined benefits of an RF survey are well worth the investment, and are a major step toward the reliable operation of the seismic network.



### **8.9.3.2 Problems with RF interference**

Radio frequency interference caused by other users of VHF or UHF RF space in many countries is quite a common and difficult problem. There are several reasons for that. In some countries, there is confusion and a lack of discipline in matters of RF space: army, police, security authorities, and civil authorities may all operate under different (or no) rules and cause unforeseen interference. In other countries, poor maintenance of high-power communication equipment results in strong, stray radiation from the side lobes of powerful transmitters. This radiation can interfere with seismological radio links. Extensive, unauthorized use of walkie-talkies can also be the cause of problems.

The best, and more or less the only solution is to work closely with local RF experts during the design phase of a seismic network. They are practically the only source of information about true RF space conditions in a country. Note that RF interference problems are generally beyond control of seismic system manufacturers and seismological community. All RF equipment, no matter who manufactures it, are designed to be used in an RF space where everybody strictly obeys the rules. One also has to, as much as possible, avoid other high-power RF space users (see IS 7.2).

### **8.9.3.3 Organization of RF data transmission network design**

An RF layout design is always an integral part of a seismic site selection procedure. Theoretically the seismic system purchaser can perform it if he has adequate knowledge in this field. However, practice shows that this is rarely the case. Even if the RF survey is purchased from an independent company or from a seismic equipment manufacturer as a part of the services, the process still requires involvement of the seismic system buyer. For efficient office and fieldwork, the customer has to prepare beforehand an approximate initial seismic network layout, road and topographic maps, and climatic data. He has to make available knowledgeable staff members and well-informed local people acquainted with local conditions at the sites, who will join the site selection and RF survey field team. He should also assure efficient logistics during the fieldwork.

A detailed list of what to prepare is given in the IS 7.1.

### **8.9.4 Purchasing a seismic system**

There are different ways of buying equipment for a completely new network. If the user has little experience, it might be desirable to get all equipment and installation from one company while if the user has sufficient knowledge or can get help from other institutions, only equipment should be bought and possibly from several different companies in order to get the best price/performance. Buying all from one company is often a more expensive solution. However, compatibility between the different components is essential, but more difficult to achieve when mixing. Some countries/institutions require that equipment (and installation) go through a bidding process. The bidding process may help to carry out a more systematic evaluation and pay off. Some institutions insist that all equipment and services are obtained from a local representative. While there may be good reasons for this, it can increase the cost.

A very important requirement for the new network is specification of how and in which format the data is delivered. Today, most equipment is able to produce data in one of the standard formats used for processing, however the data must also be easily available for integration into one or several of the standard systems used for processing. Just recording all data continuously is not much help if there is no user-friendly software available to access and extract any part of that data. So an integral part of the specification is a requirement that the data must be delivered ready for processing in a given processing system.

#### **8.9.4.1 The bidding process**

While sending out a Request for Proposal and asking for bids on a new seismic system may be a good way to get started, there are a number of important issues one must be aware of when requesting bids or proposals. First, certain technical requirements and business standards must be met in order to be able to compare "apples to apples" when it is time to analyze the system proposals received. Second, in order to find the most suitable system, one needs to invest a fair amount of additional time in research and investigation before sending out the bid specifications. Namely, some very important issues may be hard to define in the Request for Proposal. The proposals can easily give unclear information regarding the following crucial issues:

- reliability of the equipment;
- user friendliness of the system;
- availability of long-term support by the manufacturer including true availability of spare parts in the next years;
- financial stability of the manufacturer.

In the Request for Proposal one should include all relevant basic technical information, so that the manufacturer can put together the corresponding technical solution. However, we recommend that the Request for Proposal does not contain an over-detailed technical description of the desired system. With too many technical details, one can end up limiting one's choices and even disqualifying the most suitable system just because a relatively unimportant technical detail can not be fulfilled.

Avoid buying brand new systems in the market unless you are really assured of excellent support from the manufacturer. Brand new systems frequently have more problems than older more tested systems. Their use will require a high level of knowledge and a really good working relationship with the manufacturer while solving these problems.

Some countries are required by law to accept the lowest bid. Unfortunately, crucial qualities like services, equipment reliability, user friendliness of the system, amount of factory testing, setup and long-term support might be easily lost if one bases the choice solely on the lowest price for all of the stated (but practically never really sufficient) requirements of the bid.

Manufacturers of seismic equipment can offer a turnkey system whereby they will purchase all of the necessary components not made by them. They will include their administrative labor costs for acquiring these components. Do not assume that they will be able to purchase every item at prices lower than you will be able to. Federal, state, and local governments and universities (typical operators of seismic networks) often have secured special pricing from vendors that can be substantially cheaper than what seismic equipment manufacturers can obtain.

#### **8.9.4.2 Selecting a vendor**

When evaluating the proposals, one should assess not only the technical qualities of the system, but also the quality of every manufacturer. It is a common practice that companies invite potential buyers to a promotion visit. This does not necessarily give relevant information on how well the system would perform. If a visit is desired, carefully select the people who will participate in these visits. In addition to a member fully responsible in financial issues, one member of the team should be the individual responsible for future operation of the network. Other members of the team should be those most knowledgeable and experienced in seismology, no matter what their position in the hierarchy or which institution they belong to.

Ask for visits with manufacturer's sales or system engineers. Data sheets themselves seldom give enough technical information about a seismic system. Sales and system engineers can provide all the details of a particular technical solution. Such visits, however, are less appropriate during the early stage of the project when one's goals are not yet specifically set. It is understandable that sales representatives will be biased toward the equipment of the manufacturer they represent. Visiting other networks of similar size in different countries is often better spent money than visiting a company trying to promote their equipment and may give more realistic information of reliability and capabilities of different equipment. Reliability and company service are important. This information will mostly be unavailable from the company offers and only by talking to other users can one find out. It is in itself no guarantee that the company is large and well known, although such a company might provide good long term service.

#### **8.9.4.3 Equipment selection**

As already mentioned, data sheets of seismic equipment alone seldom provide enough information. In addition, it is not easy to compare the data sheets of various manufacturers because each one, to some extent, uses a different system of specifications, measurement units, and definitions of technical parameters. In most cases small differences in specifications have no importance in real life, like e.g. if a digitizer has 22 or 23 bit resolution. The seismic equipment has a hardware and software component, which are almost equally important. The acquisition unit (digitizer possibly integrated) usually has a user interface that allows to change settings and monitor operation, but also to monitor and control the seismometer. The software also provides the functionality for real-time data transmission. When selecting a system it is important to evaluate the overall integration and not just to look at the performance of individual components. So for example, the digitizer with the best dynamic range is of no use in a real-time network if it does not have all functionality for data transmission.

Ideally, we recommend buying one piece of key equipment such as a sensor, a data logger, processing software with demo data or an RF link and testing the product yourself. In the case of large projects with adequate financing, manufacturers will often loan equipment for testing purposes free of charge. While it is ideal to get some firsthand experience before settling on which new system to purchase, this approach requires personnel who are knowledgeable about seismology and instrumentation.

The technical systems must be able to operate in the intended environment. Today most equipment is installed in protected locations like houses and vaults and communication is delivered by third parties (like a phone line), so in many cases there is no strict requirement for temperature and humidity. So paying extra money for a recorder which is able to sustain 2 m of water is wasted money if the unit is installed in a protected environment. Outdoor equipment today is GPS antennas which by default can be expected to be made for outdoor use and communication equipment. For the latter, usually only antennas are installed outdoors. The main indoor problem might be temperature and humidity.

Each technical system, or element in it, properly operates within a certain set of parameters, or "range". One should be familiar with these ranges and know where, within this range, the system will actually operate. If one or more of elements of the system are to operate at the extreme end of their operation range on a regular basis, most probably a different element or system should be selected. Note that there is always a price to pay for operating equipment under extremes. It is always best to have a safety margin in your system and do not expect it to operate continuously, efficiently, and reliably in extreme ranges.

#### **8.9.4.4 The seismic equipment market is small**

The global market for seismic systems and equipment is naturally quite limited. With very few exceptions, instruments are produced in small numbers. Inevitably, this sets a limit to the quantity and thoroughness of testing of the newly developed equipment. This is not a result of a lack of quality or commitment on the part of manufacturers in this field, but a simple, economic reality. Compared to industries with a far broader and more powerful economic base, like computer and electronic companies, seismic equipment moves into the field with relatively little testing, even by the most reputable manufacturers. In general, the equipment arrives with a higher than average number of bugs and technical imperfections that will need to be solved by the manufacturer and the user working in tandem.

Currently, most seismic equipment and technical documentation is less user-friendly and complete than desired. Customers are rarely given comprehensive and easy-to-follow instructions on how to setup and use the system. Equipment and software does not always arrive with the promised features. It may be possible to address some of these issues during acceptance tests.

#### **8.9.4.5 Processing software**

The majority of seismic network manufactures have relatively little experience in seismic signal processing and as a general rule, do not have adequate software. It simply does not pay to develop this kind of software. On the other hand, there are public domain software packages available, which can solve these tasks and these are often offered by the manufactures. However, very little training is offered and a new network operator may end up with an expensive network but very primitive processing tools. Therefore, obtaining adequate processing software and training is an important and integral part of the planning of a new network. Unfortunately this is often not the case and the value of the network can be greatly reduced.

## **8.9.5 System installation**

### **8.9.5.1 Four ways of seismic system installation**

Generally we can define four methods for the installation of a new seismic system.

- 1) The user installs the new system. Only ‘boxes’ are purchased. In this option, the customer is responsible for the proper functioning of the system as a whole and the manufacturer remains responsible for proper functioning of the elements, unless they are improperly used or installed. This approach gives the user great flexibility, but also the main responsibility. It is only an option if qualified staff can be appointed to this task and/or if local or international organizations can participate.
- 2) The manufacturer demonstrates installation on a subsystem (a few stations, a sub-network). The user installs the rest. In this case, the manufacturer and the user, share responsibility for the system functioning. This approach is often successful. However, the customer must have a certain amount of experience with seismic, computer, and communication equipment for this method to work.
- 3) The manufacturer installs the whole system with a full assistance from local technical and seismological staff that will be responsible for running, maintaining, and servicing the network in the future. Responsibility for making sure the system functions lies with the manufacturer. The main benefit of this approach for the users is that they learn an enormous amount during the hands-on installation and associated problem solving time. This is probably the most efficient method of training. The user should not expect savings and potential shortening of the installation time but rather some additional time and effort will be required from manufacturer. In our experience, this is the best way of installing a seismic network in a country where little or no experience with seismic equipment exists.
- 4) The manufacturer has the complete responsibility for installing a turnkey system and making sure it functions adequately without any assistance from the customer. In this case, the network will no doubt be successfully installed, but local staff members will not learn about its operation nor how to solve potential future problems. In general, this is not a desirable solution.

Two technical details relating to system installation should also be mentioned here. In the case that the system buyer will install the system or its parts, do not select the 'standard length' cables sometimes offered by seismic system manufacturers. The 'standard' cables rarely work well in the field. They are, according to Murphy's laws, always too short or too long. Do not loop or coil extra cable length because that will increase the threat of lightning damage, unnecessarily increase system noise, and in the end, you will be paying for the “extra” cable. Rather, ask for bulk cables with separate connectors or cables of a reasonable length margin and one-side mounted connectors only. During installation in the field they can then be cut to precisely the desired length. Note, however, that reliable, high quality soldering of connectors requires experience. Inexperienced technicians have little chance of performing the job correctly and poorly installed connectors are among the most frequent causes of problems at a seismic station.

### **8.9.5.2 Organization of civil engineering works**

Whatever construction work is needed to prepare the sites, it is usually arranged and paid for by the customer of the new network rather than the manufacturer of the seismic equipment. Very large national projects may be an exception to this rule. Site construction will require a great deal of preparation and involvement by the system buyer. There are generally a number of good design alternatives from which to choose and we suggest hiring a local civil engineering contractor to design the best solution for a particular system and specific circumstances in the country. A seismo-geologist and a civil engineer should supervise the construction work. Their main responsibility is assuring that the enclosure is watertight and that the sensors have a good contact with solid bedrock. The system's manufacturer can usually provide sketches and suggestions for the procedure and may also supervise the work, but usually does not provide true structural engineering drawings for seismic shelters. Working in tandem with a local civil engineer is usually a better choice because the engineer will be familiar with all local circumstances that are unknown to the manufacturer of the seismic equipment. Local builders know best what materials and construction methods are available and workable in a particular country. Do not "over-engineer" the project; it is usually not necessary to have a big civil engineering firm design every detail, oversee all seismic site preparation, and then build the site.

### **8.9.5.3 Summary of recommendations for setting up a new network**

- To ensure long term maintenance and get lowest possible operational cost, the user should build as much as possible of the network himself
- Study and visit other networks to learn the requirements
- Decide on communication
- Decide on type of network based on communication possibilities
- Decide on integration of all components into one system
- Make sure processing and data storage systems are in place (including training) before network installation is finished

#### ***Real time network***

- If a real time network there are two possibilities to ensure real time completeness of data(retransmission):
  - (a) Use public domain SeedLink with standard computers and digitizers. Several digitizers can be used. Requires some computer skills but is the cheapest and most flexible option.
  - (b) Use a company system where digitizer, communication protocol and receiving software is a company standard. Only equipment from that company can then be used. Simple and reliable but more expensive and less flexible than a public domain system.
- Triggering: All real time system should have a trigger. It is recommended to use one of the public domain systems such as EarthWorm and Seiscomp.

#### ***Networks with no communication***

This type of networks are mostly temporary but could also be permanent

- Chose a recorder with a standard interface and data recording format. Chose between
  - (a) A black box with built in software and a simple way to tap out continuous data

(e.g. memory stick). Usually a reliable solution and it often represent a robust mechanical solution

(b) A PC with a digitizer and a public domain software. Requires more computer skills, protected housing but can be cheaper and more flexible.

- Make sure software is available for handling and integration of the continuous data into the processing software data base.

## **8.9.6 Running a seismic network**

### **8.9.6.1 Tuning of seismic networks**

Before a seismic network can function with its full capacity, it must be tuned to local seismo-geological and system conditions. Tuning is especially important for networks that run in triggered mode. Unfortunately, many operators are not aware of the importance of fine-tuning.

The local and regional Earth's structure, the seismic network dimensions and layout, regional seismicity, seismic noise levels at station sites, seismic signal attenuation in the region, all play a role in these adjustments. One will not be able to correctly tune the system's recording and processing parameters until one has acquired sufficient experience with natural and man-made seismic noise and earthquake signals at all the sites in the network and until one fully understands the parameters that have to be tuned. Therefore, tuning a network takes normally months of systematic work. Because of the long time required to accomplish this task, the system's manufacturer simply can not do it. Only the network operator can correctly tune the network. Moreover, since seismic noise conditions at the sites may change with time, new stations may be added, the goals of the network may change, etc., re-tuning of the network will probably be required from time to time. In reality, tuning a seismic network is an ongoing task, which can not be done 'once and for all.'

Actual tuning procedures are manifold. Besides appropriate gain setting and accurate quantification at individual stations, the most important tuning is that of the trigger parameters:

- trigger threshold values;
- detrigger threshold values;
- trigger time windows' duration and other parameters;
- weights of individual stations in coincident trigger algorithm;
- grouping of stations into sub-regions for a coincidence-trigger algorithm;

Detailed discussion of individual parameters is beyond the scope of this text. Note that not all enumerated parameters exist in every seismic network and that some adjustments may be missing from this list. A thorough description and parameter adjustment procedure for the short-time-average/long-time-average (STA/LTA) seismic trigger algorithm is given in the annexed IS 8.1 on "Understanding and parameter setting of STA/LTA trigger algorithms". Further guidelines for other network tuning procedures may be added later as complementary Information Sheets.

The following are some of the offline seismic analysis software issues that must be studied and prepared for efficient routine observatory work, and parameters that have to be adjusted for correct analysis of seismic records:

- files containing information about data acquisition parameters (data acquisition configuration file(s));
- files containing data about geometrical configuration of seismic stations (network configuration file(s));
- parameter files containing sensor calibration data;
- Earth model parameters of event location program(s) (layer thickness, seismic-wave velocity, seismic station weights, epicentral distance weighing function, and similar parameters depending on the program used);
- automatic phase-picker parameters;
- magnitude scale parameters;
- preparation of different macros and forms for routine, everyday analysis of seismic signals.

Some parameters, e.g., for the Earth model, are often insufficiently known at the time of network installation and require long-term seismological research work, which results in gradual refinement of the model and increasingly better event locations.

No manufacturer can optimally pre-adjust all these parameters to the specific local conditions. Seismic networks usually come with a set of default values for all these parameters (factory pre-selected values based on 'world averages'). These values may work sufficiently well for the beginning of network operations, however, optimum seismic network performances requires reconsidering most of them.

#### **8.9.6.2 Organizing routine operation tasks**

Keeping one's network failure-free and in perfect working order while waiting to record earthquakes year after year requires hard and responsible work and a lot of discipline. Well-defined personal responsibility with respect to altering network operation parameters and strict obedience to the established procedures is an absolute must.

This goal is generally not simple to achieve. Seismic observatory staff will have to operate in a highly professional and reliable manner with:

- clearly defined personal responsibility for each task associated with the routine operation of the network and for other everyday analysis and archiving activities;
- regular maintenance of hardware and software;
- continuous verification of all tasks and hardware operations;
- maintenance of precise records of all relevant activities that effect data parameters, availability, continuity, and quality, such as changes to network operational parameters, processing procedures, data archiving, equipment maintenance and repair.

Regular processing of seismic data requires that all details of how data is processed and stored is well planned and that personnel are adequately trained.



Network recording parameters should be changed only if there is an important and well thought through reason. Because any change to the recording parameters will affect the network's ability to detect earthquakes, these changes should be avoided as much as possible. From the point of view of monitoring seismicity, ideally, there should be no changes for years after the network is fully adjusted. Nevertheless, those changes that are inevitably required from time to time should be kept to a minimum and carefully documented and archived.

Careful and continuous documentation of network operation parameters in a logbook, log file, or in the seismic database itself, is essential. This operational information should contain all information about data acquisition parameters and their changes, a documentation of all station calibrations, a precise record of each station's downtime, descriptions of technical problems and solutions, and descriptions of maintenance and service work. The exact times of parameter changes must be thoroughly recorded. This time-dependent information must become an integral part of the seismic data archive because without it the data can not be properly interpreted.

Usually a seismic network team is divided into a seismological and a technical group. This is fine as it relates to every day network operation activities and responsibilities. However, as much as possible, the basic technical as well as basic seismological knowledge should be 'evenly' distributed among the members of both groups. This favorably influences the general quality of the work of a seismic observatory. It also helps very much in many of critical situations, such as following a severe, unexpected technical problem, following a large earthquake, during the rapid deployment of portable stations following a main shock, or when any other situation dramatically increases the amount of work for a limited period of time.

The technical group must accept that no matter how modern and sophisticated the seismic network is that they operate; their customers are the seismologists. Therefore the seismologists must define the goals of seismic network operation and its working parameters. Frequently personal frictions may appear if this issue is not clearly defined by the management.

Many seismological observatories in high seismic risk regions must have people on duty at the central-recording site 24 hours per day. This may be a more or less explicit government requirement in order to be able to quickly notify public and civil defense authorities in the case of a strong, potentially damaging earthquake. No matter how understandable such desire may be, however, this working regime is really feasible only in a very large seismological institution. Only they have enough seismologists capable of quickly and competently interpreting seismic data. Even a fully-automated central recording and processing facility requires verification and confirmation of automatically determined earthquake parameters by trained personnel. The interpretation of automatically determined earthquake parameters in terms of the expected intensities in a given region and the probability of potential fatalities and damage is still a matter of experience and is not yet a matter of automatic calculations.

In practice, the around-the-clock human presence at the observatory is often achieved using all of the available, but mostly untrained, personnel in order to formally fulfill higher authorities' requirements. Of course, the actual value of such a 'solution' is questionable. If the alarms are of primary importance for a new network, one should consider using a system of a message on a mobile phone that will automatically alarm the institution's seismologists and give a first rough location and magnitude in the event of a strong earthquake. The seismologist should in turn be able to access the database remotely, if not living very closely

to the observatory, and interactively reassess and refine the first quick parameter estimates if required.

### **8.9.6.3 System maintenance**

Maintaining a seismic network's hardware and software is a continuous activity that inevitably requires well-trained personnel. Nowadays, many vital operational parameters and equipment health at seismic stations can be remotely monitored. Such parameters are for example: backup battery voltage, presence of charging voltage, potential software and communication problems, absolute time keeping, remote station vault and/or equipment temperature, potential water intrusion, etc. These utilities significantly reduce the need for field service work and therefore lower the cost of network operation. However, regular visits to the stations are still necessary, though far less frequently than in the past. Yet, once per year seems a minimum.

Note that it is a mistake to simply put off visits to remote seismic stations until something goes wrong. Periodic visual checks of cables and equipment, of potential corrosion problems on equipment and grounding and lightning system, and for intrusion of water and small animals are important. Batteries, burned lightning protection elements, and desiccants must be changed regularly, and cleaning the vaults and solar panels will also help to eliminate technical problems before they occur.

When something does go wrong, the technical staff must be certain that they can respond immediately with the right personnel, action, and spare parts. One should always maintain a good stockpile of the most common spare parts and have a well-trained technician with a pager on duty around the clock. Having technical personnel, in addition to seismologists, on call 24 hours a day for potential action is a good practice in the observatory seismology business, however this is rarely possible and not really necessary with most networks.

Operators of large networks may not have the manpower or budget resources to visit all of their stations annually. The major differences in maintenance procedures for small networks versus large regional or national networks are response time to site outages, site sensor-calibrations, and preventive maintenance (PM) visits. A large, dense seismic network lessens the need for 100% uptime for all sites; maintenance visits for site outages can be scheduled with PM visits in an area, something that a small, local network of 10 to 20 sites can not afford. This eliminates the need for immediate technician response and a 'beeper' for field repairs. For example: The U.S. Geological Survey's Northern California Seismic Network (NCSN), a large, dense regional network (352 analog and 93 digital stations), visits their telephone telemetered sites every 20 months and solar-powered sites every 4 years for site electronic equipment exchanges. These maintenance intervals are possible due to the robustness and reliability of their electronic amplifier/telemetry packages and associated equipments.

Be aware that batteries require special attention. If the lightning damages are the most frequent source of technical failures during normal operation conditions of a network, then battery failures will be the number one reason for failures during main power failures and unusually high-periods of seismicity. It should be noted that the output voltage alone of a battery provides little information about its overall health and capacity. Many types of batteries may still have adequate output voltage while at the same time their charge capacity is

reduced to a small fraction of its original strength. Batteries in this condition will not do the job in case of a long-duration power failure, as may occur after a damaging earthquake.

Ideally, all of the batteries in the seismic system should be laboratory tested once a year for their remaining charge capacity. The batteries should be fully discharged, then fully charged, and again discharged in a controlled manner and their true charge capacity determined. Once the measured charge capacity is less than 60% - 70% of their nominal capacity, they should be replaced with new ones. Relying solely on measurements of battery voltage will certainly lead to technical failures in the long run. The most important moment in the lifetime of the seismic network may happen only once a decade or less. One certainly does not want to miss it because of old batteries with insufficient charge capacity!

However, large networks may again not be able to laboratory test each battery once per year. The NCSN exchanges batteries using an operational window system for battery life (based upon the quality and the replacement cost of the batteries used, and their long-term experience with battery lifetimes) rather than with annual testing and rejuvenation. Their operational window for solar-panel batteries is 4 years (Tom Burdette, personal communication, 2002).

Non-chargeable batteries, particularly the lithium type, should be replaced regularly, in accordance with the manufacturer's instructions, regardless of their output voltage at the moment of lifetime expiration.

#### **8.9.6.4 Sensor calibration**

Seismological observatories should calibrate all of the sensors in their seismic system regularly - ideally, once a year. Strictly speaking, only the seismic signals recorded between two successive sensor calibrations that show no significant change in the sensor frequency response function are completely reliable. Sensor and sensor calibration issues are also different for a dense network equipped with modern sensors. Modern sensors are very robust, and many broadband sensors have automatic self-leveling, self-correcting features that eliminate the need for annual calibrations. In addition, site electronics can be installed to provide regular, telemetered sensor tests for response and operation. These features, along with a dense network sensor configuration allow for sensors to be replaced and recalibrated on a regular schedule. For NCSN, the short-period sensors are replaced at 10-year intervals, unless a sensor fails beforehand. NCSN short-period sites have built-in calibrators that perform daily mass releases to test sensor operation and response (Tom Burdette, personal communication, 2002).

Seismic sensor calibration requires knowledge that often is not available locally. In digital seismology, the sensor transfer function representation in form of poles and zeros is most commonly used. Both issues are discussed in detail in Chapter 5 and the annexed Exercises and Program Descriptions. A comprehensive description of basics is also given in Scherbaum (1996, 2001, and 2007). A description of a popular seismometer calibration program UNICAL is given in Plešinger et al. (1995).

#### 8.9.6.5 Archiving seismic data

After several decades, or even years, of operating a seismic network, the scientific and financial value of the recorded data is extremely high. Therefore, full attention must be paid to data archiving and a failsafe backup of the data. Seismology is a typical non-experimental science and lost or corrupted seismic data can never be regenerated. It is therefore an absolute must to provide a complete and reliable backup archive. The backups should be kept in a different physical location, no matter whether they are on paper, tape, disk, CD or other memory medium. Today, most networks record continuously giving very large data sets and one carefully needs to consider the backup options. Whenever possible, one copy (or the originals) should be stored in fire-resistant cabinets or safes. It is important to note that microfiche, film, and computer media require more protection than paper records. Paper records can withstand temperatures to 177°C (350°F), but computer media is damaged beyond use by temperatures above 52°C (125°F) and 80% humidity.

When one first sets up a seismic network, one needs to think thoroughly about organizing the data that is recorded in light of the fact that eventually the network will have many, many years of accumulated records. Often, this crucial aspect of seismic system organization is overlooked or left to on-the-spot decisions by whoever is in charge of the initial network operation. This may work fine for a while, but eventually everybody will run into serious problems if the archiving system chosen is inappropriate. It is necessary to carefully think through the archiving organization at the outset and to keep the long-term future in mind. At the same time archiving systems will be outdated in 5-10 years, and one has to be prepared to change to new systems regularly.

In a small, weak-motion network in a region of low seismicity that generates only a small number of events records each year, or in a small or medium size strong motion network, one can probably get by with a directory tree organization for the data archive, particularly if only event data is stored. However, it is now more common also for small networks to record continuously. Nevertheless, filename coding of events must be thoroughly thought out to avoid confusion and/or file name duplications. File names also should reflect complete date and time of each event. Larger networks in moderate to high seismicity regions might require a better-organized, relational database for archiving purposes. One should carefully consider the various options used by other seismological observatories and those available on the market before the network starts recording data. It is very painful to change the data coding or archiving method after several years of network operation, once thousands upon thousands of records are already stored. Today, the most common format for data storage is MiniSEED.

Very powerful professional databases may not be the most suitable choice for smaller networks, primarily due to their high initial and annual maintenance cost, and secondly, due to too many expensive build-in utilities which will never be used in seismology. However, they are the only choice for larger data centers with 1,000s of real-time channels. Special databases which have been developed by the seismological community for the needs of seismology, thoroughly tested in several existing applications, and accepted by many, seem to be the best choice at the moment.

If possible, it is advisable to keep the raw seismic data as produced by the digitizer (raw event files, or sequences of continuous data) in the archive along with the full documentation about the recording conditions (data acquisition parameters and accompanying information).

Processing and seismic analysis methods will change and evolve as time passes, so it can be useful to go back.

#### **8.9.6.6 Dissemination of seismic data**

International cooperation in the dissemination of seismic data is another prerequisite for the high-quality operation of any new seismic network. Broad-minded data sharing is the best way for a less experienced institution to get feedback about the quality of its own work and is also a widely accepted international obligation. Data formats for parameter and waveform data exchange are dealt with in Chapter 10.

The Internet is nowadays the most common way to disseminate earthquake information and waveform data through email, web-pages, etc. Other forms of communication such as fax, pager, mobile phone are also used for earthquake alert purposes. Some of the currently most relevant and often used Internet addresses of global, regional and national seismological data centers can also be found and directly linked via <http://www.szgrf.brg.de> and <http://www.seismo.ethz.ch/seismosurf/seismobig.html>.

Everyone can greatly improve their own work by observing and comparing their phase readings, event locations, magnitude determinations and source mechanism results with the results of others published in national or international seismological bulletins. Any seismic study should also include as much seismic information as possible from the neighboring regions and countries. Not only one's own data, but also all available pertinent data from others should be used in seismic research work. Disseminating one's own data will, in turn, facilitate easy and fast accessibility of data from others. It's very important to establish a generous data sharing relationship with other seismological institutions.

The U.S. Geological Survey National Earthquake Information Center (<http://neic.usgs.gov>) compiles data contributed from networks located around the globe in order to determine, as rapidly and as accurately as possible, the location and size of all destructive earthquakes that occur worldwide. This information is disseminated immediately to concerned national and international agencies, scientists, and the general public. The NEIC collects and provides to scientists and to the public an extensive seismic database that serves as a solid foundation for scientific research, principally through the operation of modern digital national and global seismograph networks and through cooperative international agreements. Similar services are provided by a number of agencies around the world.

Data from the NEIC is transferred to the International Seismological Centre (ISC) (<http://www.isc.ac.uk/>) for final bulletin creation about two years behind real time. The International Seismological Centre is a non-governmental organization charged with the final collection, analysis and publication of standard earthquake information from all over the world. Earthquake readings are received from almost 3,000 seismograph stations representing every part of the globe. The Center's main task is to re-determine earthquake locations making use of all available information, and to search for new earthquakes, previously unidentified by individual agencies.

Besides these global data centers for parametric data, there are many national, regional or global centers that provide access to seismological waveform data. Examples of global data centers are IRIS, GEOFON and GEOSCOPE, Data is also available from regional networks

such as MEDNET or nationally operating networks, e.g. GRF/GRSN (see 8.7.5), ICC etc. Suitable starting links are provided, e.g., from the web sites of the US Advanced National Seismic System (<http://www.anss.org/>) and of the Observatories and Research Facilities for European Seismology (ORFEUS) (<http://orfeus.knmi.nl>).

Traditionally, seismic observatories of national seismic networks or larger regional networks regularly publish preliminary seismological bulletins (weekly, or monthly), final seismological bulletins (yearly), and earthquake catalogs of the country or region (yearly, but with a few years delay so that the data from all other external sources can be included in the analysis). These catalogs are one of the bases for earthquake hazard assessment and for risk mitigation studies.

## Acknowledgments

The authors and the Editor are very thankful to John Lahr and Kent Fogleman of the USGS for their careful English prove-readings and complementary remarks that have helped to improve the original draft. Thanks go Egill Hauksson for contributing a new text and figure on the California Integrated Network, to Shin Aoi for providing an updated text and figure on the Japanese Seismic Networks, to Klaus Stammer for updating text and figure for the German seismic networks that transmit their data for joint processing at the BGR in Hannover, and to Liu Ruifeng for making available the new Fig. 8.16 and an extended power point presentation which allowed us to produce the summary text in section 8.7.4, which has been authorized by Liu Ruifeng.

## Recommended overview readings

Barrientos et al. (2001)  
Havskov and Alguacil (2006)  
Hardt and Scherbaum (1994)  
Hutt et al. (2002) (give details for list of references!)  
Lee and Steward (1981)  
Rabinovitz and Steinberg (1990)  
Uhrhammer et al. (1998)  
Willmore (Ed.) (1979).

## References

- Barrientos, S., Haslinger, F., and IMS Seismic Section (2001). Seismological monitoring of the Comprehensive Nuclear-Test-Ban Treaty. *Carl Hanser Verlag*, München, Kerntechnik, **66**, 3, 82-89.
- Bartal, Y., Somer, Z., Leonhard, G., Steinberg, D. M., and Horin, Y. B. (2000). Optimal seismic networks in Israel in the context of the Comprehensive Test Ban Treaty. *Bull. Seism. Soc. Am.*, **90**, 1, p 151-165.
- Bondár I. and McLaughlin, K. L. (2009). A new ground truth data set for seismic studies. *Seismol. Res. Lett.*, **80**(3), 465-472

- Bycroft, G. N. (1978). The effect of soil-structure interaction on seismometer readings, *Bull. Seism. Soc. Am.*, **68**, 823.
- Fleming, K., Picozzi, M., Milkereit, C., Kühnlenz, F., Lichtblau, B., Fischer, J., Zulfikar, C., Özel, O., and the SAFER and EDIM working groups (2009). The self-organizing seismic early warning information network (SOSEWIN). *Seismol. Res. Lett.*, **80**(5), 755-771; doi: 10.1785/gssrl.80.5.755,
- Francis, T. J. G., Porter, I. T., and Lilwall, R. C. (1978). Microearthquakes near the eastern end of St. Paul's Fracture Zone. *Geophys. J. R. astr. Soc.*, **53**, 201-217.
- Gomberg, J. S., Shedlock, K. M., and Roecker, W. W. (1990). The effect of S-wave arrival times on the accuracy of hypocenter estimation. *Bull. Seism. Soc. Am.*, **80**, 6, 1605-1628.
- Hardt, M., and Scherbaum, F. (1994). The design of optimum networks for aftershock recordings. *Geophys. J. Int.*, **117**, 716-726-
- Havskov, J., and Alguacil, G. (2006). Instrumentation in earthquake seismology. *Springer*, Berlin, 2<sup>nd</sup> edition, 358 pp. plus CD.
- Hutt, C. R., Bolton, H. F., and Holcomb, L.G. (2002). US contribution to digital global seismograph networks. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam*, 319-332.
- Hutton, L. K., Woessner, J., and Hauksson, E. (2010). Seventy-seven years (1932 – 2009) of earthquake monitoring in southern California., *Bull. Seismol. Soc. Am.*, **100** (2); 423-446; doi: 10.1785/0120090130.
- Kijko, A. (1977). An algorithm for the optimum distribution of a regional seismic network - I. *Pageoph*, **115**, 999-1009.
- Kradolfer, U. (1996). AutoDRM - The first five years. *Seism. Res. Lett.*, **67**, 4, 30-33.
- Lee, W. H. K., and Steward, S. W. (1981). Principles and applications of micro-earthquake networks, *Advances in Geophysics*, Supplement 2, *Academic Press*, New York.
- Luco, J. E., Anderson, J. G., and Georgevich, M. (1990). Soil-structure interaction effects on strong motion accelerograms recorded on instrument shelters. *Earth. Eng. Struct. Dyn.*, **19**, p 119.
- McLaren, J. P., and Frohlich, C. (1985). Model calculations of regional network locations for earthquakes in subduction zones. *Bull. Seism. Soc. Am.*, **75**, 2, 397-413.
- Murphy, J. R., Rodi, W., Johnson, M., Sultanov, D. D., Bennett, T. J., Toksöz, M. N., Ovtchinnikov, V., Barker, V. W., Reiter, D. T., Rosca, A. C., and Shchukin, Y. (2005). Calibration of International Monitoring System (IMS) stations in Central and Eastern Asia for improved seismic event location. *Bulletin of the Seismological Society of America*, **95**, 1535-1560.
- Ottewöller, L., and Havskov, J. (1999). SeisNet: A general purpose virtual seismic network. *Seism. Res. Lett.*, **70**, 5, 522-528.
- Plešinger, A. (1998). Determination of seismograph system transfer functions by inversion of transient and steady-state calibration responses. *Studia geoph. et geod.*, **42**, 472-499.
- Rabinowitz, N., and Steinberg, D. M. (1990). Optimal configuration of a seismographic network: A statistical approach. *Bull. Seism. Soc. Am.*, **80**, 1, 187-196.
- Scherbaum, F. (1996). Of poles and zeros; Fundamentals of Digital Seismometry, *Kluwer Academic Publishers*, Boston, 256 pp.

- Scherbaum, F. (2001). Of poles and zeros: Fundamentals of digital seismology. Modern Approaches in Geophysics, *Kluwer Academic Publishers*, 2<sup>nd</sup> edition, 265 pp. (including an CD-ROM by Schmidtke, E., and Scherbaum, F. with examples written in Java).
- Scherbaum, F. (2007). Of poles and zeros; Fundamentals of Digital Seismometry. 3<sup>rd</sup> revised edition, *Springer Verlag*, Berlin und Heidelberg.
- SEED (2007). SEED reference manual. Standard for the exchange of earthquake data, SEED format version 2.4. International Federation of Digital Seismograph Networks Incorporated Research Institutions for Seismology (IRIS), USGS. [www.iris.edu](http://www.iris.edu)
- Steinberg, D. M., Rabinowitz, N., Shimshoni, Y., and Mizrachi, D. (1995). Configuring a seismograph network for optimal monitoring of fault lines and multiple sources. *Bull. Seism. Soc. Am.*, **85**, 6, 1847-1857.
- Trnkoczy, A., and Živčić, M. (1992). Design of Local Seismic Network for Nuclear Power Plant Krsko, *Cahiers du Centre Europeen de Geodynamique et de Seismologie*, Luxembourg, **5**, 31-41.
- Uhrhammer, R. A. (1980). Analysis of small seismographic station networks. *Bull. Seism. Soc. Am.*, **70**, 1369-1379.
- Uhrhammer, R. A., Karavas, W., and Romanovicz, B. (1998). Broadband Seismic Station Installation Guidelines, *Seism. Res. Lett.*, **69**, 15-26.
- Utheim, T., Havskov, J., and Natvik, Y. (2001). Seislog data acquisition systems. *Seism. Res. Lett.*, 77-79.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.
- Young, Ch. J., Chael, E. P., Withers, M. W., and Aster, R. C. (1996). A comparison of the high- frequency (>1Hz) surface and subsurface noise environment at three sites in the United States. *Bull. Seism. Soc. Am.*, **86**, 5, 1516-1528.



| Topic   | Understanding and parameter setting of STA/LTA trigger algorithm                                                             |
|---------|------------------------------------------------------------------------------------------------------------------------------|
| Author  | Amadej Trnkoczy (formerly Kinemetrics SA);<br>E-mail: <a href="mailto:amadej.trnkoczy@siol.net">amadej.trnkoczy@siol.net</a> |
| Version | September 1999, amended 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_8.1                                                                |

|                                                                                       | Page      |
|---------------------------------------------------------------------------------------|-----------|
| <b>1 Introduction</b>                                                                 | <b>1</b>  |
| <b>2 Purpose</b>                                                                      | <b>3</b>  |
| <b>3 How it works - basics</b>                                                        | <b>5</b>  |
| <b>4 How to adjust STA/LTA trigger parameter</b>                                      |           |
| 4.1 General considerations                                                            | 5         |
| 4.2 Seismometers and/or accelerometers?                                               | 5         |
| 4.3 One- and three-component seismic stations                                         | 6         |
| 4.4 Sensitivity of seismic sensors                                                    | 6         |
| 4.5 Frequency range of seismic sensors                                                | 7         |
| <b>5 How to adjust associated parameters for proper triggering and data recording</b> | <b>10</b> |
| 5.1 Selection of trigger filters                                                      |           |
| 5.2 Selection of post-event time (PET) parameter                                      |           |
| 5.3 Selection of pre-event time (PEM) parameter                                       |           |
| 5.4 Selection of voting scheme parameters                                             |           |
| <b>6 Practical recommendations for finding optimal trigger parameters</b>             |           |
| <b>References</b>                                                                     |           |

## 1 Introduction

By introducing digital seismic data acquisition, long-term continuous recording and archiving of seismic signals has become a demanding technical problem. A seismic network or even a single seismic station operating continuously at high sampling frequency produces an enormous amount of data, which is often difficult to store (and analyze) locally or even at the recording center of a network. This situation has forced seismologists to invent triggered seismic data acquisition. In a triggered mode, a seismic station or a seismic network still process all incoming seismic signals in real time (or in near-real-time) but incoming data is not stored continuously and permanently. Processing software - a trigger algorithm - serves for the detection of typical seismic signals (earthquakes, controlled source seismic signals, underground nuclear explosion signals, etc.) in the constantly present seismic noise signal. Once an assumed seismic event is detected, recording and storing of all incoming signals starts. It stops after trigger algorithm 'declares' the end of the seismic signal.

Automatic trigger algorithms are relatively ineffective when compared to a seismologist's pattern recognition ability during reading of seismograms, which is based on years of experience and on the enormous capability of the human brain. There are few exceptions, where the most complex detectors, mostly dedicated to a given type of seismic signals, approach to human ability. In all practical cases, automatic trigger loose some data on one side and generate falsely triggered records, which are not seismic signals, on the other. Small

amplitude seismic signals are often not resolved from seismic noise and are therefore lost for ever, and, if the trigger algorithm is set sensitively, false triggers are recorded due to irregularities and occasionally excessive amplitude of seismic noise. False triggers burden off-line data analysis later and unnecessarily occupy data memory of a seismic recording system. As a result, any triggered mode data acquisition impairs the completeness of the recorded seismic data and produces some additional work to delete false records.

Several trigger algorithms are presently known and used - from a very simple amplitude threshold type to the sophisticated pattern recognition, adaptive methods and neural network based approaches. They are based on the amplitude, the envelope, or the power of the signal(s) in time domain, or on the frequency or sequency domain content of seismic signal. Among the more sophisticated ones, Allan's (1978; 1982) and Murdock and Hutt's (1983) trigger algorithms are probably the most commonly known. Many of these algorithms function in association with the seismic phase time picking task. Seismic array detection algorithms fall into a special field of research, which will not be discussed here. For more advanced algorithms see, e.g., Joswig (1990; 1993; 1995). However, in practice, only relatively simple trigger algorithms have been really broadly accepted. and can be found in seismic data recorders in the market and in most network's real time processing packages.

The simplest trigger algorithm is the amplitude threshold trigger. It simply detects any amplitude of seismic signal exceeding a pre-set threshold. The recording starts whenever this threshold is reached. This algorithm is rarely used in weak-motion seismology but it is a standard in strong motion seismic instruments, that is in systems where high sensitivity is mostly not an issue, and where consequently man-made and natural seismic noise amplitudes are much smaller than the signals which are supposed to trigger the instrument.

The root-mean-square (RMS) threshold trigger is similar to the amplitude threshold algorithm, except that the RMS values of the amplitude in a short time window are used instead of 'instant' signal amplitude. It is less sensitive to spike-like man-made seismic noise, however it is rarely used in practice.

Today, the 'short-time-average through long-time-average trigger' (STA/LTA) is the most broadly used algorithm in weak-motion seismology. It continuously calculates the average values of the absolute amplitude of a seismic signal in two consecutive moving-time windows. The short time window (STA) is sensitive to seismic events while the long time window (LTA) provides information about the temporal amplitude of seismic noise at the site. When the ratio of both exceeds a pre-set value, an event is 'declared' and data starts being recorded in a file.

Several more sophisticated trigger algorithms are known from literature (e.g., Joswig 1990; 1993; 1995) but they are rarely used in the seismic data loggers currently in the market. Only some of them are employed in the network's real time software packages available. When in the hands of an expert, they can improve the events/false-triggers ratio significantly, particularly for a given type of seismic events. However, the sophisticated adjustments of operational parameters to actual signals and seismic noise conditions at each seismic site that these triggers require, has proven unwieldy and subject to error in practice. This is probably the main reason why the STA/LTA trigger algorithm still remains the most popular.

Successful capturing of seismic events depends on proper settings of the trigger parameters. To help with this task, this Information Sheet explains the STA/LTA trigger functioning and gives general instructions on selecting its parameters. Technical instructions on setting the trigger parameters depend on particular hardware and software and are not given here. Refer to the corresponding manuals for details.

## 2 Purpose

The short-time-average/long-time-average STA/LTA trigger is usually used in weak-motion applications that try to record as many seismic events as possible. These are the applications where the STA/LTA algorithm is most useful. It is nearly a standard trigger algorithm in portable seismic recorders, as well as in many real time processing software packages of the weak-motion seismic networks. However, it may also be useful in many strong motion applications, except when interest is limited to the strongest earthquakes.

The (STA/LTA) trigger significantly improves the recording of weak earthquakes in comparison with amplitude threshold trigger algorithms. At the same time it decreases the number of false records triggered by natural and man-made seismic noise. To some extent it also allows discrimination among different types of earthquakes.

The STA/LTA trigger parameter settings are always a tradeoff among several seismological and instrumental considerations. The goal of searching for optimal parameter settings is the highest possible seismic station sensitivity for a given type of seismic signal (which may also include the target 'all earthquakes') at a still tolerable number of false triggers.

The STA/LTA trigger is most beneficial at seismically quiet sites where natural seismic noise (marine noise) is the dominant type of seismic noise. It is also effective in case of changes of 'continuous' man-made seismic noise. Such changes, for example, occur due to day/night variation of human activity nearby or in urban areas. The STA/LTA algorithm is less effective in the presence of irregular, high amplitude man-made seismic noise which is often of burst and/or spike type.

## 3 How it works - basics

The STA/LTA algorithm continuously keeps track of the always-present changes in the seismic noise amplitude at the station site and automatically adjusts the seismic station's sensitivity to the actual seismic noise level. As a result, a significantly higher sensitivity of the system during seismically quiet periods is achieved and an excessive number of falsely triggered records is prevented, or at least mitigated, during seismically noisy periods. Calculations are repeatedly performed in real time. This process is usually taking place independently in all seismic channels of a seismic recorder or of a seismic network.

The STA/LTA algorithm processes filter seismic signals (see section 5.1 'Selection of trigger filters' in this Information Sheet) in two moving time windows – a short-time average window (STA) and a long-time average window (LTA). The STA measures the 'instant' amplitude of the seismic signal and watches for earthquakes. The LTA takes care of the current average seismic noise amplitude.

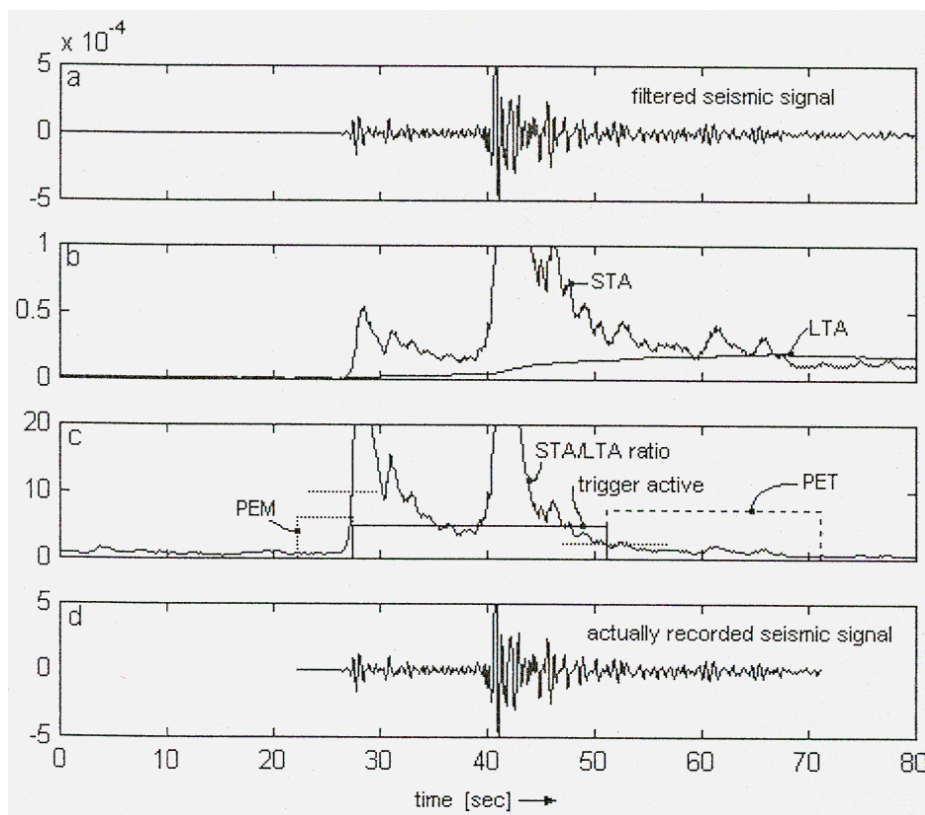
First, the absolute amplitude of each data sample of an incoming signal is calculated. Next, the average of absolute amplitudes in both windows is calculated. In a further step, a ratio of both values — STA/LTA ratio—is calculated. This ratio is continuously compared to a user selected threshold value - STA/LTA trigger threshold level. If the ratio exceeds this threshold, a channel trigger is declared. A channel trigger does not necessarily mean that a multi-channel data logger or a network actually starts to record seismic signals. All seismic networks and most seismic recorders have a 'trigger voting' mechanism built in that defines how many and which channels have to be in a triggered state before the instrument or the network actually starts to record data (see section 5.4 below - 'Selection of voting scheme parameters'). To

simplify the explanation, we shall observe only one signal channel. We will assume that a channel trigger is equivalent to a network or a recorder trigger.

After the seismic signal gradually terminates, the channel detriggers. This happens when the current STA/LTA ratio falls below another user-selected parameter - STA/LTA dettrigger threshold level. Obviously, the STA/LTA dettrigger threshold level should be lower (or rarely equal) than the STA/LTA trigger threshold level.

In addition to the data acquired during the 'trigger active' time, seismic networks and seismic recorders add a certain amount of seismic data to the event file before triggering – pre-event-time (PEM) data. After the trigger active state terminates, they also add post-event-time (PET) data.

For better understanding, Figure 1 shows a typical local event and the trigger variables (simplified) during STA/LTA triggering.



**Figure 1** Function and variables of STA/LTA trigger calculations (see text for explanations).

Graph a) shows an incoming continuous seismic signal (filtered); graph b) shows an averaged absolute signal in the STA and LTA windows, respectively, as they move in time toward the right side of the graph; and graph c) shows the ratio of both. In addition, the trigger active state (solid line rectangle), the post-event time (PET), and the pre-event time (PEM) (dotted line rectangles) are shown. In this example, the trigger threshold level parameter was set to 10 and the dettrigger threshold level to 2 (two short horizontal dotted lines). One can see that the trigger became active when the STA/LTA ratio value exceeded 10. It was deactivated when the STA/LTA ratio value fell below 2. On graph d) the actually recorded data file is shown. It includes all event phases of significance and a portion of the seismic noise at the beginning.

In reality, the STA/LTA triggers are usually slightly more complicated, however, the details are not essential for the understanding and proper setting of trigger parameters.

## 4 How to adjust STA/LTA trigger parameters

To set the basic STA/LTA trigger algorithm parameters one has to select the following:

- STA window duration
- LTA window duration
- STA/LTA trigger threshold level
- STA/LTA detrigger threshold level.

However, optimal triggering of a seismic recorder or a seismic network does not depend only on these parameters. There are usually four additional associated parameters which, only if well tuned with the trigger parameters, guarantee optimal data recording. These parameters are:

- trigger filters
- pre-event time (PEM)
- post-event time (PET)
- trigger voting scheme.

Although not directly related to the STA/LTA trigger algorithm, these additional parameters are also be discussed below in order to provide a complete information.

The STA/LTA trigger parameter and associated parameters' settings depend on the goal of the application, on the seismic noise condition at the site, on the properties of seismic signals at a given location, and on the type of sensor used. All these issues vary broadly among applications and among seismic sites. Obviously, there is no general, single rule on setting them. Each application and every seismic site requires some study, since only practical experience enables the determination of really optimal trigger settings.

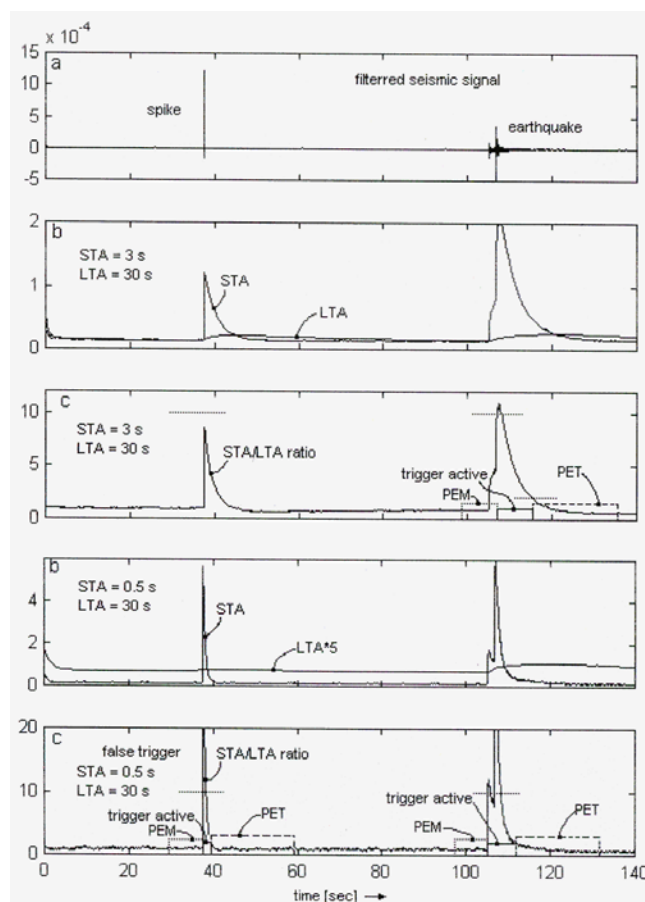
Note that seismic recorders and network software packages come with a set of default (factory set) trigger and trigger associated parameter values. They are rarely optimal and must therefore be adjusted to become efficient in a particular application. For best results, changing these parameters and gradually finding the best settings is a process which requires a certain amount of effort and time.

### 4.1 Selection of short-time average window (STA) duration

Short-time average window measures the 'instant' value of a seismic signal or its envelope. Generally, STA duration must be longer than a few periods of a typically expected seismic signal. If the STA is too short, the averaging of the seismic signal will not function properly. The STA is no longer a measure of the average signal (signal envelope) but becomes influenced by individual periods of the seismic signal. On the other hand, STA duration must be shorter than the shortest events we expect to capture.

To some extent the STA functions as a signal filter. The shorter the duration selected, the higher the trigger's sensitivity to short lasting local earthquakes compared to long lasting and lower frequency distant earthquakes. The longer the STA duration selected, the less sensitive it is for short local earthquakes. Therefore, by changing the STA duration one can, to some extent, prioritize capturing of distant or local events.

The STA duration is also important with respect to false triggers. By decreasing the duration of the STA window, triggering gets more sensitive to spike-type man-made seismic noise, and vice versa. Although such noise is usually of instrumental nature, it can also be seismic. At the sites highly polluted with spike-type noise, one will be frequently forced to make the STA duration significantly longer than these spikes, if false triggers are too numerous. Unfortunately, this will also decrease the sensitivity of the recording to very local events of short duration. Figure 2 explains the effect of STA duration on local events and spike-type noise.



**Figure 2** Influence of STA duration on trigger sensitivity to short local events and 'spiky' noise in seismic signals.

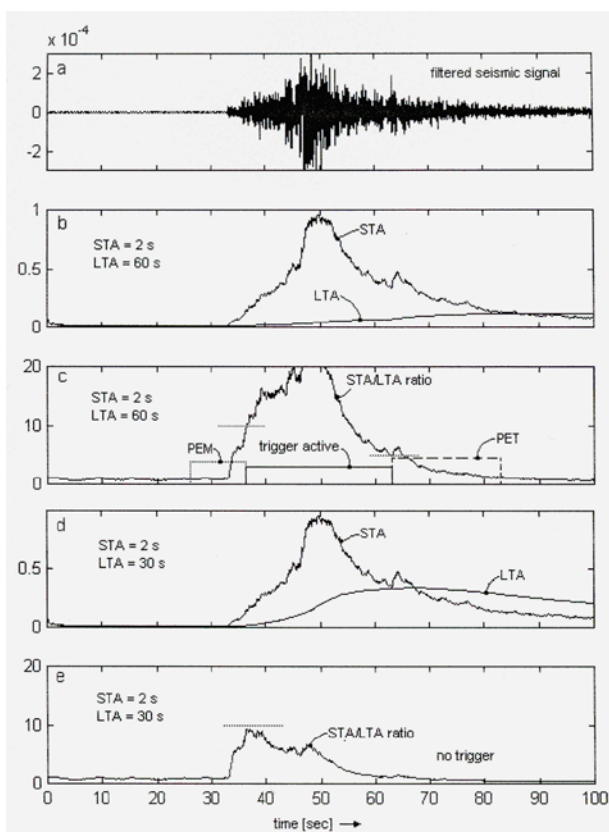
On graph a) a signal with an instrumental spike on the left and with a short, very local earthquake on the right side is shown. Graphs b) and c) show STA, LTA, STA/LTA ratio, and trigger active states along with PEM and PET. The STA/LTA trigger threshold was set to 10 and dettrigger threshold to 2. One can see that when using a relatively long STA of 3 sec, the earthquake did trigger the system, but only barely. However, a much bigger amplitude (but shorter) instrumental spike did not trigger it. The STA/LTA ratio did not exceed the STA/LTA threshold and there was no falsely triggered record due to the spike. The lower two

graphs show the same variables but for a shorter STA of 0.5 sec. The spike clearly triggered the system and caused a false record. Of course, the earthquake triggered the system as well.

For regional events, a typical value of STA duration is between 1 and 2 sec. For local earthquakes shorter values around 0.5 to 0.3 s are commonly used in practice.

#### 4.2 Selection of long-time average window (LTA) duration

The LTA window measures average amplitude seismic noise. It should last longer than a few 'periods' of typically irregular seismic noise fluctuations. By changing the LTA window duration, one can make the recording more or less sensitive to regional events in the 'Pn'-wave range from about 200 to 1500 km epicentral distance. These events typically have the low-amplitude emergent Pn- waves as the first onset. A short LTA duration allows the LTA value more or less to adjust to the slowly increasing amplitude of emergent seismic waves. Thus the STA/LTA ratio remains low in spite of increasing STA (nominator and denominator of the ratio increase). This effectively diminishes trigger sensitivity to such events. In the opposite case, using a long LTA window duration, trigger sensitivity to the emergent earthquakes is increased because the LTA value is not so rapidly influenced by the emergent seismic signal, allowing Sg/Lg waves to trigger the recording. Figure 3 explains the described situation.



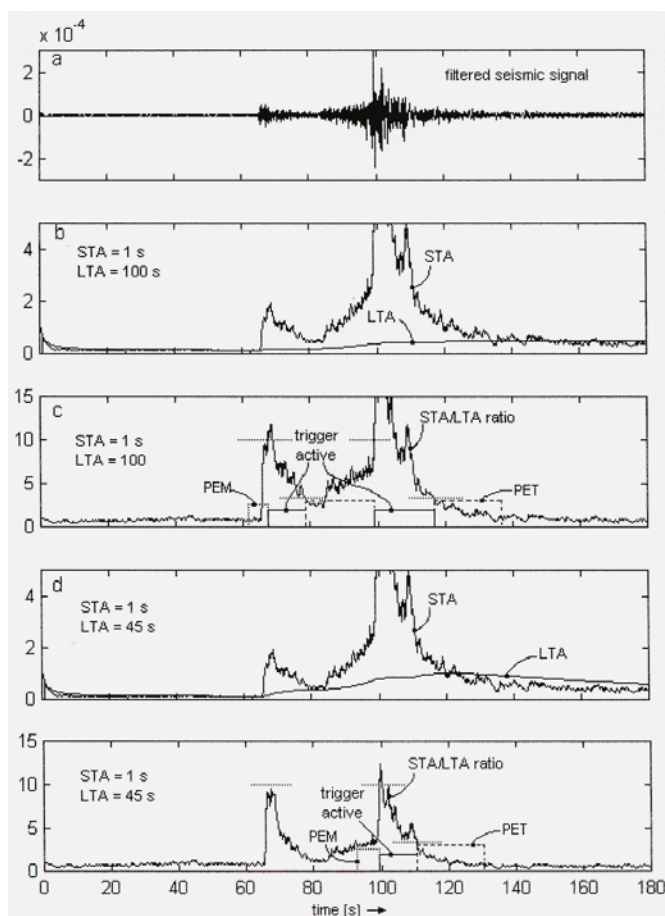
**Figure 3** Influence of LTA duration on trigger algorithm sensitivity to earthquakes with emergent seismic signals.

In graph a) an event with emergent P waves is shown. Graphs b) and c) show the time course of trigger parameters for a relatively long LTA of 60 sec. The LTA does not change fast,

allowing the STA/LTA ratio to exceed the STA/LTA trigger threshold (short horizontal dotted line) and a normal record results. Graphs d) and e) show the same situation with a shorter LTA of 30 s. The LTA value increases much faster during the initial phase of the event, thus decreasing the STA/LTA ratio value which does not exceed the STA/LTA trigger threshold. No triggering occurs and the event is missed.

Similarly, efficient triggering of recording of events with weak P waves compared to S waves requires a longer LTA for two reasons. First, if P waves do not trigger, they 'contaminate' true information about seismic noise prior to the event measured by LTA, since their amplitude exceeds the amplitude of seismic noise before the event. This results in diminished trigger sensitivity at the moment when S waves arrive. This 'contamination' is decreased if a longer LTA duration is selected. Second, longer LTA makes the trigger more sensitive to P waves as well, if they are not strictly of impact type.

Figure 4 represents such a case. Graph a) shows a typical event with significantly bigger later phase waves than P waves. Graphs b) and c) show trigger parameters for a long LTA of 100 s. P wave packet as well as S wave packet trigger the recorder. Appropriate PEM and PET assure that the event is recorded as a whole in a single file with all its phases and a portion of seismic noise before them. Graphs d) and e) show the same situation but for a shorter LTA of 45 sec. One can see that the P waves did not trigger at all, while the S waves barely triggered. The STA/LTA ratio hardly exceeds the STA/LTA trigger threshold. As the result, the recorded data file is much too short. P waves and information about seismic noise before them are missing in this record. A slightly smaller event would not trigger at all.

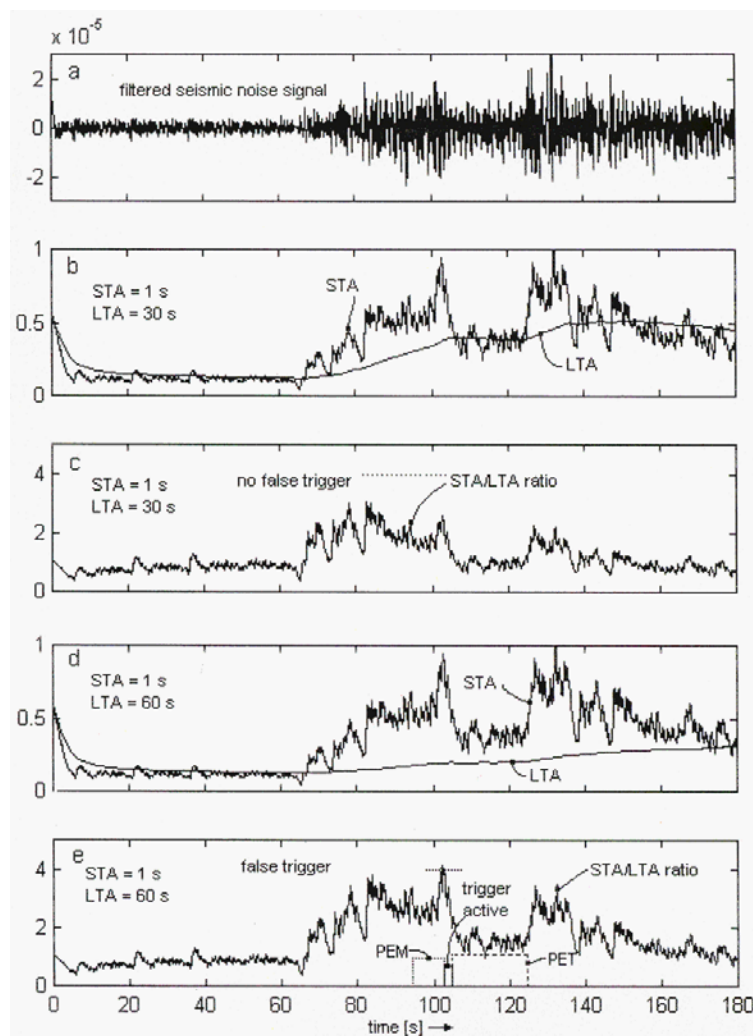


**Figure 4** Influence of LTA duration on trigger algorithm sensitivity to earthquakes containing weak P waves.



On the other hand, a short LTA will successfully accommodate recorder sensitivity to gradual changes of 'continuous' man-made seismic noise. Such 'transition' of man-made seismic noise from low to high is typical for night-to-day transition of human activity in urban areas. Sometimes, using a short LTA can mitigate false triggers due to traffic. Examples of such cases could be a single heavy vehicle approaching and passing close to the seismic station on a local road, or trains on a nearby railway. A short LTA can 'accommodate' itself fast enough to such emerging disturbances and prevent false triggers.

Figure 5 shows an example of the LTA response to increased seismic noise. Graph a) shows seismic noise, which gradually increased in the middle of the record. Note that the change of its amplitude is not sudden but lasts about 20 to 30 sec. Graphs b) and c) show the situation at a short LTA of 30 sec. One can see that the LTA value more or less keeps track of the increased noise amplitude. The STA/LTA ratio remains well below the STA/LTA trigger threshold and there is no false trigger in spite of significantly increased seismic noise at the site. Graphs d) and e) show the situation with a longer LTA of 60 s. In this case, the LTA does not change so rapidly, allowing a higher STA/LTA ratio during noise increase. As the result, a false trigger occurs and a false record is generated which unnecessarily occupies data memory.



**Figure 5** Influence of LTA duration on false triggering when seismic noise conditions change.

Natural seismic noise (marine noise) can change its amplitude by a factor exceeding the value of twenty. However, these changes are slow. Significant changes can occur only during a few hours period, or at worst, in several tens of minutes. Therefore even the longest LTA duration is short enough to allow LTA to accommodate completely to marine noise amplitude variations.

The LTA duration of 60 seconds is a common initial value. A shorter LTA duration is needed to exclude emergent regional events from triggering, if desired, or if quickly changing man-made noise is typical for the site. A longer LTA can be used for distant regional events with very long S-P times and potentially emergent P waves.

### 4.3 Frozen versus continuously updated LTA during events

Calculations of the LTA value during an event, that is after a channel trigger is declared, can be performed in the first approximation in two different ways.

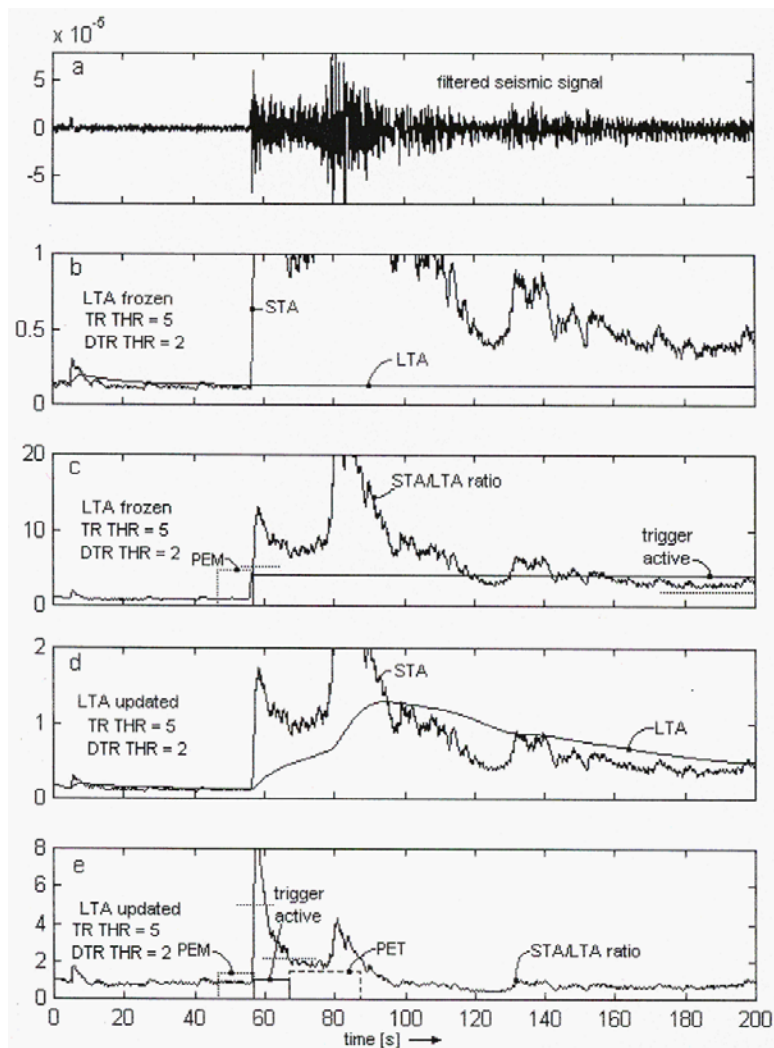
Either the LTA value is continuously updated and calculated during the event as usual, or the LTA value is kept frozen at the moment when channel trigger is declared. In this case the LTA is not allowed to change (increase) during an event at all. Most of seismic recorders available in the market have both frozen or continuously updated LTA user-selectable options. However, each approach has its good and bad points.

The 'frozen' LTA window (the word 'clamped' is also used in literature) can force the unit into a permanently triggered state in case of a sudden increase of man-made seismic noise at the site. The situation is illustrated in Figure 6.

Graph a) shows an earthquake during which seismic noise increases and remains high even after the termination of the event. Such a situation can happen if, for example, a machinery is switched on in the vicinity of the recorder. In such a case, a completely frozen LTA (graph b) would never again allow the STA/LTA ratio to fall below the STA/LTA det trigger threshold level (graph c) and a continuous record would result. The result is that the seismic recorder's memory soon gets full and blocks further data recording.

A continuously updated LTA (the word 'unclamped' is also used in literature), on the other hand, frequently terminates records too early. Graphs d) and e) of Figure 6 explain this situation. Very often records with truncated coda waves result because the LTA increases rapidly if the beginning portion of a large earthquake signal is included in its calculation. Thus the STA/LTA ratio decreases too rapidly and terminates recording prematurely. Coda waves of the event are then lost, as shown in the Figure 6. This undesired result could be even much more distracting for records of regional events with longer duration.

Some seismic recorders work with a special calculation of LTA. The LTA value is, to the first approximation, 'frozen' after a trigger. However, this 'freezing' is not made complete. Some 'bleeding' of event signal into the LTA calculation is allowed. Such an algorithm tries to solve both problems: it does not cause endlessly triggered records in the case of a rapid permanent increase of seismic noise and, at the same time, it does not cut coda waves too early.



**Figure 6** Potential problems with two conventional ways of calculating the LTA: an endless record with a completely frozen LTA and cut coda waves with updated LTA calculations.

#### 4.4 Selection of STA/LTA trigger threshold level

The STA/LTA trigger threshold level to the greatest extent determines which events will be recorded and which will not. The higher value one sets, the more earthquakes will not be recorded, but the fewer false-triggers will result. The lower the STA/LTA trigger threshold level is selected, the more sensitive the seismic station will be and the more events will be recorded. However, more frequent false triggers also will occupy data memory and burden the analyst. An optimal STA/LTA trigger threshold level depends on seismic noise conditions at the site and on one's tolerance to falsely triggered records. Not only the amplitude but also the type of seismic noise influence the setting of the optimal STA/LTA trigger threshold level. A statistically stationary seismic noise (with less irregular fluctuations) allows a lower STA/LTA trigger threshold level; completely irregular behavior of seismic noise demands higher values.

Note that some false triggers and some missed earthquakes are an inevitable reality whenever recording seismic signals in an event-triggered mode. Only a continuous seismic recording, if affordable, completely solves the problem of false triggers and incompleteness of seismic data.

It is a dangerous trap to select a very high STA/LTA trigger threshold level and a high channel gain simultaneously. Many recorders in the market allow this setting without any warning messages. This situation is particularly dangerous in extremely noisy environments, where, due to too many false triggers, the instruments are usually set to record only the strongest events.

Suppose one has set the STA/LTA trigger threshold level to 20. Suppose also that one has set the gain of the channel in such a way that it has about 150 mV of average seismic noise signal at the input of the recorder and the input full scale voltage of the channel is  $\pm 2.5$  V. Obviously, this setting would require a  $0.15 \text{ V} \times 20 = 3 \text{ V}$  signal amplitude to trigger the channel. Since its maximum input amplitude is limited to 2.5 V, it can never trigger, no matter how strong an earthquake occurs. Note that this error is not so obvious, especially in low seismicity regions with rare events. One can operate an instrument for a very long time without records and forever wait for a first recorded earthquake.

With certain products in the market, this potential danger of an erroneous setting is solved in the following way: whenever one uses the STA/LTA algorithm, an additional threshold trigger algorithm remains active in the 'background'. Because of it, the channel triggers whenever its input amplitude exceeds 50% of channel input voltage range, for example, in no relation to the STA/LTA trigger setting. In this way, the strongest and therefore the most important events are still recorded, no matter how carelessly the STA/LTA trigger algorithm parameters are set.

An initial setting for the STA/LTA trigger threshold level of 4 is common for an average quiet seismic site. Much lower values can be used only at the very best station sites with no man-made seismic noise. Higher values about 8 and above are required at less favorable sites with significant man-made seismic noise. In strong-motion applications, higher values are more common due to the usually noisier seismic environment and generally smaller interest in weak events.

#### **4.5 Selection of STA/LTA dettrigger threshold level**

The STA/LTA dettrigger threshold level determines the termination of data recording (along with the PET parameter – for more information see 5.3 below on “Selection of post-event time (PET) parameter”).

The STA/LTA dettrigger threshold level determines how well the coda waves of recorded earthquakes will be captured in data records. To include as much of the coda waves as possible, a low value is required. If one uses coda duration for magnitude determinations, such setting is obvious. However, a too low STA/LTA dettrigger threshold level is occasionally dangerous. It may cause very long or even endless records, for example, if a sudden increase in seismic noise does not allow the STA/LTA ratio to fall below the STA/LTA dettrigger threshold level. On the other hand, if one is not interested in coda waves, a higher value of STA/LTA dettrigger threshold level enables significant savings in data memory and/or data transmission time. Note that coda waves of distant earthquakes can be very long.

In general, the noisier the seismic site, the higher the value of the STA/LTA dettrigger threshold level should be used to prevent too long or continuous records. This danger is high only at sites heavily polluted by man-made seismic noise.

A typical initial value of the STA/LTA det trigger threshold level is 2 to 3 for seismically quiet sites and weak motion applications. For noisier sites higher values must be set. For strong-motion applications, where coda waves are not of the highest importance, higher values are frequently used.

## **5 How to adjust associated parameters for proper triggering and data recording**

### **5.1 Selection of trigger filters**

Nearly all seismic recorders and networks have adjustable band-pass trigger filters. They continuously filter the incoming seismic signals prior to the trigger algorithm calculations. Selection of these filters is important for a proper functioning of the STA/LTA trigger algorithm (as well as for amplitude threshold trigger algorithm). The purpose of these filters is three-fold:

- they remove DC component from incoming seismic signals, namely, all active seismic sensors have some DC offset voltage at the output which, if too high, deteriorates the STA/LTA ratio calculation. The calculation of the absolute value of the signal becomes meaningless if the DC component is higher than the seismic noise amplitude. This results in malfunction of the STA/LTA trigger algorithm and drastic reduction of trigger sensitivity for weak seismic events;
- their frequency band-pass can prioritize frequencies corresponding to the dominant frequencies of seismic events one wants to record; and
- their stop-band can attenuate dominant frequencies of the most distracting seismic noise at a given site.

The trigger filter pass-band should generally accommodate the frequencies of the maximum energy of expected seismic events. At the same time it should have a band-pass that does not coincide with peak frequency components of typical seismic noise at the site. If this is possible, a significant improvement of the event-trigger/false-trigger ratio results. Obviously, one can understand that if the peak amplitudes of seismic noise and the dominant frequencies of the events of most interest coincide, the trigger filter becomes inefficient.

One should not forget that the frequency response function of the seismic sensor used with a recorder or in a network channel also modifies the frequency content of events and noise signals at the input of trigger algorithm. Therefore the sensor used is an important factor in the choice of a trigger filter. The type of sensor output - proportional to either ground displacement, velocity or the acceleration - has a similar effect. Sensors with ground acceleration proportional output - accelerometers - emphasize high frequencies. They usually require a filter protection against excessive high-frequency man-made seismic noise. Ordinary seismometers have typically an output proportional to ground velocity, sometimes also to ground displacement and they are less influenced by high-frequency man-made seismic noise.

The adjustment flexibility of high- and low-corners of these filters varies among different products. The same is true for the steepness of the filter flanks. Generally, one does not need very steep (high order) filters and a lot of flexibility, because events, similar to the seismic noise, are highly variable. It is generally impossible to determine very precisely where exactly to set the frequency limits of these filters.

## 5.2 Selection of pre-event time (PEM ) parameter

Ideally, the triggered earthquake records should include all seismic phases of an event and, in addition, a portion of the seismic noise signal prior to it. Selection of an appropriate pre-event time (PEM) assures that the earthquake records are complete. For the majority of weak events, the trigger algorithm usually does not trigger at the beginning of the event but sometimes during the event, when the waves with the maximum amplitude of ground velocity reach the station. This happens very often with the earthquakes that have emergent onset waves, and with most of the weak local and regional events where the S phase amplitudes can be much bigger than the P phase. In practice, triggering on the S waves of the weak local and the regional earthquakes is actually more frequent than triggering on the P waves. But for seismological reasons, the P onset waves, plus some seismic noise prior to them, should be included in the record. A proper PEM should take care of this.

Technically this is solved in the following way. In seismic recorders and in a network's central recording computer, a portion of seismic signal prior to the instrument trigger time is temporarily stored in a pre-event ring buffer (abbreviation PEM denotes 'pre-event-memory') and added to the data recorded.

PEM must surmount the following periods of time:

- the desired record duration of seismic noise prior to the event;
- the maximal expected S-P time of earthquake records; and
- time needed to calculate the STA/LTA ratio, which, in the worst case, equals one STA window duration.

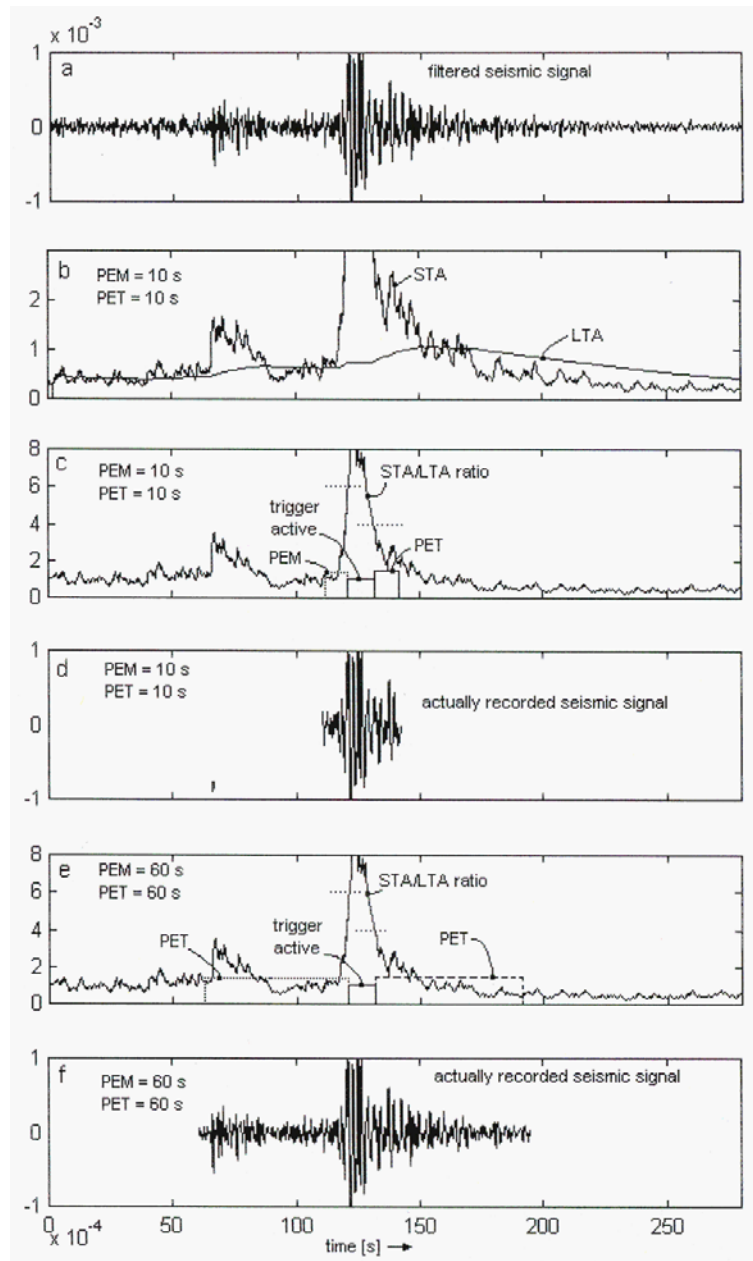
Add these three time periods and the result is the appropriate PEM value.

The effect of a too short PEM is shown in Figure 7. Graph a) shows an event approximately 400 km away from the station with weak P waves partly buried in the seismic noise. On graph b) the STA and the LTA values are shown. Graph c) shows the STA/LTA ratio and the trigger and det trigger thresholds (short horizontal dotted lines). The trigger threshold is set to 6.

One can see that the channel triggers on the S waves. However, a PEM of 10 seconds is much too short to catch the P waves. Graph d) shows the actually 'recorded' event. It starts much too late and contains no seismic noise record. Graphs e) and f) show the same event but with a properly set PEM parameter. Seismic noise as well as the P waves are properly recorded.

The maximum expected S-P time depends on the maximal distance of relevant earthquakes from the station and on seismic wave velocity in the region. For practice and for local and regional events, accurate enough results can be gained by dividing the maximum station-to-epicenter distance of interest by 5 (distance in miles) or by 8 (distance in km) to get the required maximum S-P time in seconds.

The application dictates the choice of the desired pre-event noise record duration. At least a few seconds are usually required. Note that if one wants to study spectral properties of weak events, seismic noise spectra are usually required to calculate signal-to-noise ratio as a function of frequency. This, however, requires a significant length of noise records depending on the lowest frequency of interest. The PEM must be set accordingly.



**Figure 7** Proper and improper setting of the pre-event time (PEM) and the post-event time (PET)

As an example, let us calculate a required PEM parameter value for a temporary local seismic network with 50 km aperture, where 0.5 sec is set for the STA duration. Suppose that no coincidence trigger exists and all stations run independent trigger algorithms. The operator of the network is interested in the seismicity 200 km around the center of the network. He would also like to have a 10 sec long record of seismic noise before the P waves. We need 0.5 sec for STA calculation, 10 sec for seismic noise, and  $\approx (200 \text{ km} + 50/2 \text{ km})/8 \approx 28 \text{ sec}$  to cover the maximum expected S-P time. Note that the most distant station from the epicenter in the network still has to record P waves — that is why we added one half of the network aperture to the maximum epicentral distance of interest. The PEM should therefore be set to  $0.5 + 10 + 28 \sim 40 \text{ sec}$ . Obviously, smaller networks and shorter ranges of interest require shorter PEM and vice versa.

### 5.3 Selection of post-event time (PET) parameter

The post-event time parameter assures complete recording of seismic events after a det trigger. The main purpose of PET is catching the remaining earthquake coda waves that are smaller in amplitude than the STA/LTA det trigger threshold level. Functionally PET is simply a fixed recording time added to the event file after an instrument or a network (not individual channels!) det triggers. It has a similar effect on coda waves as the STA/LTA det trigger threshold level parameter. However, its effect is event-size independent. This makes it a less effective coda wave 'catcher' than a low STA/LTA det trigger threshold level. It is most suitable for local events. Practical values of PET are usually too short to be of any help for large distant earthquakes with very long coda waves. Contrary to a very low STA/LTA det trigger threshold level that may cause re-triggering problems, a long PET is safe in this respect (see 4.5 above on "Selection of STA/LTA det trigger threshold level").

Optimal PET duration depends mostly on the application. If coda waves are important, a long PET should be selected. If coda waves have no significance, use a short PET. Obviously, the short local events require only a short PET, regional and teleseismic events, on the other hand, would require much longer PET.

A reasonable value for local seismology would be 30 sec, and 60 to 90 sec for regional seismology, assuming one wants the coda waves well recorded. To find optimal value, observe coda waves of your records and adjust the PET accordingly.

There are usually no practical instrumental limitations on selection of the longest PET. However, note that very long PETs use up the recorder's data memory easily. So, do not exaggerate, particularly in seismically very active areas or if a high rate of false triggers is accepted.

### 5.4 Selection of voting-scheme parameters

The coincidence trigger algorithm, available either in seismic networks or within a multi-channel stand-alone seismic recorder, or in a group of interconnected seismic recorders, uses voting scheme for triggering. The voting-scheme parameters are actually not directly related to the STA/LTA trigger algorithm. However, inappropriate setting of voting scheme prohibits efficient functioning of overall triggering of a network or a recorder. For that reason we also deal with the voting scheme parameters in this section.

In section 4 'How to adjust STA/LTA trigger parameters', we described how each individual channel would trigger if it were the only one in an instrument or in a network. In the following section we describe how the individual channel triggers are combined to cause the system to trigger in a multi-channel recorder or in a seismic network. We call this 'voting', as a number of votes or weights can be assigned to each seismic channel so that they may cause the system to trigger. Only if the total number of votes exceeds a given pre-set value, does the system actually trigger, a new data file opens, and data acquisition begins.

How this voting system is set up depends on the nature of the signals that one is trying to record and on the seismic noise conditions at sensor sites. The noisy channels, which would frequently falsely trigger, will obviously have less 'votes' or assigned weights than the quiet, 'reliable' channels. One will need some first-hand experience of the conditions at the sites before optimizing this voting scheme. The voting mechanism and the terminology differ to



some extent among products. However, there are usually four basic terms associated with the voting scheme parameters, namely:

- Channel weights (votes)

A channel weight defines the number of votes the channel contributes to the total when it is triggered. If the channel has a good signal/noise ratio, assign it a positive number of votes. The more 'reliable' the channel in terms of the event trigger/false trigger ratio is, the higher the number should be selected. If a channel is noisy and frequently falsely triggers, give it lesser or even zero weight. In case you want a channel to inhibit triggering (a rare case indeed), give it negative weights.

- Trigger weight

This is the total number of weights required to get the seismic recorder or the network to trigger.

- Detrigger weight

The Detrigger weight is a value below which the total trigger weights (sum of all individual weights) must fall in order to cause a recorder or a network to detrigger. The Detrigger weight of 1 usually means that all voting channels must be detriggered before the recorder will detrigger. However, other definitions are also possible.

- External channel trigger weight

This value represents the number of weights one assigns to the 'external trigger channel' source. This parameter is most useful in networks of interconnected stand-alone seismic recorders. In this configuration every triggered recorder 'informs' all other units in the network that it triggered. If one wants to ensure that all recorders in the network trigger when one unit triggers, the external trigger channel should have the same weight as the Trigger weight. If one wants to use a combination of an external trigger with other internal criteria, one should set the weights accordingly.

Understanding of the voting scheme parameters is best gained through examples. The following section gives a few examples of various voting schemes.

- A classic strong-motion seismic recorder set at a free-field site has no interconnected units and normally has a three-component internal FBA accelerometer. One would set all three Channel weights to 1 and also set the Trigger weight to 1. Consequently any channel could trigger the system. At noisier sites a Trigger weight set to 2 would be more appropriate. In the latter case, two channels must be in a triggered mode simultaneously (or within a time period usually named aperture propagation window (APW) time, which is an additional parameter available with seismic networks) for the beginning of data recording.
- A small weak motion seismic network around a mine is designed for monitoring local micro-earthquakes. It consists of 5 surface seismic stations with vertical component short-period seismometers and one three-component down-hole accelerometer. An 8-channel data logger is used at the network's center. One of the surface stations is extremely noisy due to nearby construction works. All others have approximately the same seismic noise amplitudes. One can temporally set a Channel weight 0 to the noisy station to exclude its contribution to triggering and channel weights of 1 to all other surface stations. The down-hole accelerometer is very quiet but

less sensitive than surface stations (accelerometers). Select Channel weight 2 for each of its components. For this network a trigger weight of 3 would be an adequate initial selection. The system triggers either if at least three surface stations trigger, or two components of the accelerometer trigger, or one surface station and one component of the accelerometer trigger. Suppose also that there are frequent blasts in the mine. If one wishes, one can use an External trigger channel weight set to -8 and manually (with a switch) prevent seismic network recording of these blasts (down-hole:  $3 \times 2$  channel votes + surface stations:  $4 \times 1$  channel vote - 8 External votes < 3 Trigger votes).

- Let us suppose that an interconnected strong motion network of two seismic recorders with the internal three-component FBA accelerometers is installed in a building, one in the basement and one on the roof. Initially one can set the Channel weights to 1 for each signal channel, as well as for the External trigger channel. Suppose the Trigger weight is set to 1 as well. As a result each channel of the system can trigger both units in the system.

After a while one discovers that the seismic recorder on the roof triggers the system much too frequently, due to the swaying of the building in the wind. Changing the voting scheme of the roof unit so that Trigger weight is 3, its channels have 1 weight, while the External trigger channel has 3 weights, can compensate for this action. Now, the recorder installed on the roof triggers only if all its three channels trigger simultaneously or if the 'quiet' recorder in the basement triggers. The number of falsely triggered records will be drastically reduced.

- A small regional radio-frequency (RF) FM modulated telemetry seismic network, with a coincidence trigger algorithm at the central recording site, has 7 short period three-component seismic stations. The three stations, #1 west of the center and #2 and #3 east, not far from the center, have a low seismic noise and are connected to the center via three independent reliable RF links. The two stations north to the center, #4 and #5, are linked with the center via a joined RF repeater. The link between this repeater and the center is, unfortunately, frequently influenced by RF interference, resulting in frequent and simultaneous spikes and glitches in all six transmitted seismic signals. Due to unfavorable geology these two stations have a relatively high seismic noise. The two stations south of the center, #6 and #7, are also connected to the center via another common RF repeater. The station #6 is very quiet and the station #7 is influenced by traffic on a nearby new busy freeway. The RF link from this repeater to the center is less RF interference prone.

In such a situation (apart from trying to technically solve the RF link problem with the northern stations and repositioning of the station #7) an appropriate initial voting scheme would be as follows. A Trigger weight set to 7 (to disable otherwise much too frequent false triggering of the system due to RF interference on all 6 channels of the two northern seismic stations) and a Channel weight 1 to all channels of the northern stations (their total Channel weight should not exceed the Trigger weight), a channel weight 3 to all three channels of the station #1 (to allow independent triggering of the system if all three channels of this good station are triggered), a channel weight 2 to all channels of the stations #2 and #3 (to allow triggering of the whole system if at least four channels of these two closely situated stations are activated), a channel weight 2 to all channels of the station #6 (to accentuate its low seismic noise characteristics but to prevent independent triggering of the system due to occasional RF interference), and a weight 0 to all channels of the station #7 (to exclude its partition in triggering due to excessive man-made seismic noise).

These examples should give enough insight into the flexibility of the coincidence triggering options and about some of the ways in which this flexibility can be used for a particular application. Note also that any initial voting scheme can be significantly improved after more experience is gained with seismic noise conditions at the sites.

## 6 Practical recommendations for finding optimal triggering parameters

A systematic approach is required for successful adjustment of the optimal triggering and the associated parameters. First, the goals of the seismic installation must be carefully considered and *a priori* knowledge about seismic noise (if any) at the site(s) must be taken into account. Based on this information, the initial parameters are set. Information about them must be saved for documentation purpose. Start with rather low trigger threshold level settings than with a too high setting. Otherwise one can waste too much time in getting a sufficient number of records for a meaningful analysis required for further adjustment steps.

Then the instrument or the network is left to operate for a given period of time. The required length of operation without changing recording parameters depends strongly on the seismic activity in the region. At least several earthquakes and/or falsely triggered records must be recorded before the first readjustment of parameters is feasible. Judgments based on a single or a few records rarely lead to improvements. Such work simply doesn't arrive at any meaningful adjustment.

Afterwards, all records, including those falsely triggered, must be inspected. The completeness of the event records is checked (seismic noise, the P arrivals, the coda waves), and the causes of the false triggers are analyzed. The ratio of event-records/false-records is calculated and compared to the target level. If the number of false triggers does not reach the accepted level, increase the trigger sensitivity by lowering the STA/LTA trigger threshold level(s). Basically, one will acquire more seismic information for nearly the same price and effort. If the number of false triggers is too high, find the reasons why and try to mitigate them by changing STA/LTA and/or voting scheme parameters. Only if this doesn't help, must one decrease trigger sensitivities.

After the analysis is finished, the parameters are changed according to its findings and the new settings archived for documentation purposes. Again the instrument or the network is left active for a certain period of time, the new records are analyzed, and other changes made if needed. By repeating this process one will gradually find the best parameter setting.

## References

- Allen, R. V. (1978). Automatic earthquake recognition and timing from single traces. *Bull. Seism. Soc. Am.*, **68**, 1521-1532.
- Allen, R. V. (1982). Automatic phase pickers: Their present and future prospects. *Bull. Seism. Soc. Am.*, **72**, 225-242.
- Joswig, M. (1990). Pattern recognition for earthquake detection. *Bull. Seism. Soc. Am.*, **80**, 170-186.
- Joswig, M. (1993). Single-trace detection and array-wide coincidence association of local earthquakes and explosions. *Computers and Geosciences*, **19**, 207-221.

Joswig, M. (1995). Artificial intelligence techniques applied to seismic signal analysis. Conseil de l'Europe, *Cahiers du Centre Européen de Géodynamique et de Séismologie*, **Vol. 9**, Proceedings of the Workshop: Dynamic systems and artificial intelligence applied to data banks in geophysics, 5-15.

Murdock, J. N., and Hutt, C. R. (1983). A new event detector designed for the seismic research observatories. *USGS, Open File Report* **83-7851**.

| Topic          | Communication systems used in seismology                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|----------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Authors</b> | <b>Michael Guenther</b> GFZ German Research Centre for Geosciences, Department Geoengineering Centres, Section Centre for Early Warning, Telegrafenberg, D-14473 Potsdam, Germany;<br>E-mail: <a href="mailto:guenther@gfz-potsdam.de">guenther@gfz-potsdam.de</a><br><b>Angelo Strollo</b> GFZ German Research Centre for Geosciences, Department 2 Physics of the Earth, Section 2.4 Seismology, Telegrafenberg, D-14473 Potsdam, Germany<br>E-mail: <a href="mailto:strollo@gfz-potsdam.de">strollo@gfz-potsdam.de</a> |
| <b>Version</b> | August 2013; DOI: 10.2312/GFZ.NMSOP-2_IS_8.2                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |

|                                                                                                                                                    | Page |
|----------------------------------------------------------------------------------------------------------------------------------------------------|------|
| <b>1 Introduction</b>                                                                                                                              | 1    |
| <b>2 Requirements analysis</b>                                                                                                                     | 2    |
| <b>3 Design of a communication system for a seismic network</b>                                                                                    | 4    |
| <b>4 Overview of current communication technologies</b>                                                                                            | 6    |
| 4.1 Wired communication technologies                                                                                                               | 6    |
| 4.2 Wireless communication technologies                                                                                                            | 8    |
| 4.3 Satellite communication technologies                                                                                                           | 10   |
| <b>5 Conclusions</b>                                                                                                                               | 12   |
| <b>Acknowledgment and comments</b>                                                                                                                 | 14   |
| <b>References</b>                                                                                                                                  | 14   |
| <b>Appendix 1: Case study of a communication infrastructure for a seismic network of 25 stations for tsunami early warning in the Indian Ocean</b> | 15   |
| <b>Appendix 2: List of manufacturers and products</b>                                                                                              | 20   |

## 1 Introduction

With the rapid development of communication technology, especially in the area of satellite communication, new possibilities arise for the design and implementation of seismic networks. A decade ago, the only financially appropriate solution to integrate seismic stations into a network was often the use of terrestrial wireless links, such as radio frequency telemetry systems. These systems, although good value for money, are usually limited by their maximum link distance and their data throughput. As prices for satellite-based communication services have dropped drastically, these services have become a widely used alternative for largely distributed seismic networks.

This Information Sheet provides a short overview of a range of communication systems useful in remote areas where regular Internet access or dial-up phone lines are not available.

Some turnkey solutions for seismic networks are commercially available and most already integrate a communication system (e.g. Libra VSAT provided by Nanometrics) to avoid extensive analysis and design work by the customer. This has advantages and disadvantages (see 8.8.4). The communication system is well optimized for seismic data transmission for the manufacturers turnkey solution, but usually lacks flexibility.

This Information Sheet aims to support the requirements analysis, the design phase and the selection process when upgrades of existing seismic networks or new installations are planned without using a turnkey solution. Communication technologies and their relevant characteristics are briefly compared with emphasis on practical aspects.

Finally, Appendix 1 provides a quick network design schema using the information provided in this sheet and Appendix 2 a list of manufacturers and products the reader may refer to for further information.

## 2 Requirements analysis

Generally the selection process is a trade-off between requirements driven by the application of the seismic data, characteristics of available communication systems and most importantly financial constraints of the project. Cost aspects are usually in focus; however, it is of very high importance to thoroughly understand the requirements of the application for the seismic network before making a good choice. When designing a seismic network the relevant points to be considered with respect to the communication system are the following:

- Are seismic data needed in near real-time or downloaded at defined intervals?
  - Type of communication link needed (permanent online and streaming or dial-up and file-based transmission)
- Is the network a temporary or permanent deployment?
  - Equipment transportation and installation efforts/ costs
- What are the requirements regarding resolution and sampling rate of the waveforms?
  - Data rate limits and operational costs
- How wide is the geographical distribution of the stations?
  - Link distance limits and coverage area
- How remote are the stations?
  - Communication systems availability at site
  - Travel and transportation costs
  - Power supply investments for communication system
  - Back-up communication system for troubleshooting
- What are the availability constraints of the network (especially during natural disasters)?
  - Shared vs. dedicated links
  - Terrestrial vs. satellite based systems
  - Back-up communication system
- Are there data security requirements?
  - Encryption and authentication support
- Last but not least: what is the project budget?
  - Hardware costs
  - Operational costs
  - Transportation costs

- o Power supply investments
- o Travel and maintenance costs

Nowadays, with the increasing demand for high-quality seismic data for both, pure scientific detailed studies as well as early warning and rapid response networks, the communication system is becoming a key feature of a seismic station. In particular, different applications in seismology will impose different requirements on the communication infrastructure. A simplified list of the main seismological applications and their peculiarities, with respect to the points listed in the previous paragraph, is:

1. **Early warning systems** – permanently installed with real-time data delivery to send out timely warnings to the population as a national service (e.g. for earthquakes, tsunami, volcanic eruptions, landslides, etc.). The stations are generally located in very remote areas. Latency of the communication system technology may become an issue for on-site early warning (50 km leads to approximately 6 seconds time between the first P phase and the following destructive S phase). For other natural hazards, generally the warning time is not so short and latency of the communication technique used is not an issue.
2. **Rapid response, earthquake monitoring networks** – the aim of this network type is to have quick parameterization of an earthquake (location, magnitude, shake map) and fast earthquake parameters dissemination. Generally the earthquake information should be disseminated within few minutes from the origin time. Depending on the dimension of the target seismogenic area (local, regional or global), the size of the network and the dissemination time frame may vary.
3. **Emergency deployments** – after disastrous events, seismologists generally increase the number of stations with dense temporary deployments near the epicentre to have a better characterization of the rupture that generated the earthquake, as well as to investigate site effects. In the past, this task was generally done offline. Nowadays, with better integration among rescue teams and seismologists dedicated to emergency deployments, the real-time communication may be useful to follow the aftershock sequence evolution and even predict its evolution. Moreover, real-time state of health monitoring of buildings may be done especially for critical infrastructures (i.e. hospitals, bridges, tunnels, pipelines, etc.)
4. **Building monitoring** – besides the state of health monitoring done in emergency deployments, in some countries exposed to a high earthquake hazard, it is becoming a standard practice to continuously monitor the state of health of critical infrastructures in real-time and follow its state of health during large earthquakes. Moreover, engineers also perform offline data acquisition on buildings in order to better understand their behaviour.
5. **Seismic arrays and micro-arrays** – these are generally small seismic networks installed to have a detailed investigation of small scale phenomena. They may be used to investigate small earthquake swarms, induced seismicity, seismic noise, etc. In general, they do not require continuous real-time links, but for some applications (expected induced seismicity) the network operator may demand the real-time acquisition and processing.

Table 1 provides a schematic overview of the main seismological applications discussed above with respect to their communication systems requirements.

**Table 1** Comparative matrix for applications based on seismic networks. Please note that for large events, the rupture area may go beyond the indicated geographic spread of the network.

| Application Type                | Scale    | Geographic Spread | Sampling/<br>Data Rates                 | Link Type                                  | Required Robustness | Remoteness of Stations                         |
|---------------------------------|----------|-------------------|-----------------------------------------|--------------------------------------------|---------------------|------------------------------------------------|
| Early Warning                   | Local    | <100 km           | High for local and low for global scale | Continuous link for real-time data streams | Very high           | Remote                                         |
|                                 | Regional | <1000 km          |                                         |                                            |                     |                                                |
|                                 | Global   | Global            |                                         |                                            |                     |                                                |
| Rapid Response                  | Local    | <100 km           | High for local and low for global scale | Continuous link for real-time data streams | High                | Remote                                         |
|                                 | Regional | <1000 km          |                                         |                                            |                     |                                                |
|                                 | Global   | Global            |                                         |                                            |                     |                                                |
| Emergency Deployments           | Local    | <500 km           | Very high/High                          | Offline, dial-up or real-time streaming    | High if real-time   | Depend on EQ location (in general urban areas) |
| Building Monitoring             | Local    | <10 km            | Very high                               | Offline, dial-up or real-time streaming    | High if real-time   | Urban areas                                    |
| Seismic Arrays and Micro-Arrays | Local    | ~100 km           | Very high/high                          | Offline or real-time streaming             | Low                 | Depend on the target area                      |

### 3 Design of a communication system for a seismic network

All applications discussed in the previous paragraph have one thing in common: there is a central storage facility to collect and process the data of all stations. This facility is usually located at a data centre or is simply a central computer installed remotely in the field. It is important to be clear about the data flow of the network in order to select a suitable topology to accommodate the general requirements.

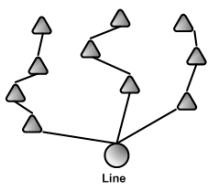
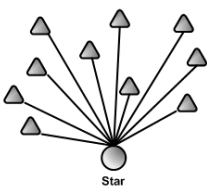
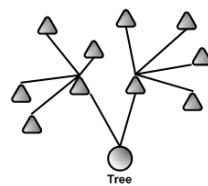
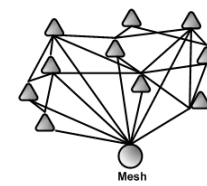
Each network topology has advantages and disadvantages (Table 2). These should be mapped against the requirements of the seismological application in order to optimize data flows. Often a combination of different topologies provides the optimal result, but hybrid topologies are more complex to troubleshoot and maintain. This can lead to higher maintenance costs. To design the network, a topology map should be drafted, indicating link distance and, if not uniform, the amount of data transmitted on each link. The data rate estimation must consider the number of sensors and components at each station, applied sampling rates, and resolution of the digitizer. Data acquisition protocols, such as SeedLink, may improve the efficiency in data transmission, depending on the compression algorithm used. (see 8.6.6). The draft should also indicate if the link needs to be available permanently or if a dial-up link is sufficient.

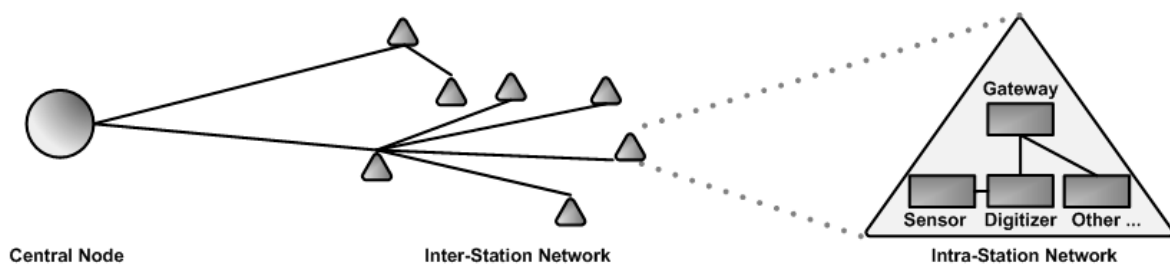
In addition to the inter-station network, each single station also has an internal- or intra-station network to connect digitizers with other local hardware (i.e., a local computer as data-buffer), as well as to the communication gateway on site (Figure 1). The gateway is the element of the



network connecting intra-station and inter-station networks. The intra-station network is usually not very complex. Sensors should be installed in separate compartments from the rest of the equipment, but generally they are in close proximity to each other (see 7.4). However, borehole installations or sensor installations in caves or mines can lead to challenging requirements. Depending on the individual conditions, this can require special solutions for the intra-station communication system. This Information Sheet also describes technical alternatives that can cover larger distances between the equipment on-site in case it is needed.

**Table 2** Network topologies: nodes represent stations and lines the communication links. For data rates we assumed uniform station setup.

| Topology              | Line                                                                              | Star                                                                              | Tree                                                                               | Mesh                                                                                       |
|-----------------------|-----------------------------------------------------------------------------------|-----------------------------------------------------------------------------------|------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------|
| Map                   |  |  |  |         |
| Maximum Link Distance | Short                                                                             | Large                                                                             | Short                                                                              | Large                                                                                      |
| Data Rates            | Data rate differs on each link                                                    | Equal data rates on each link                                                     | Data rates can differ                                                              | Data rates can differ                                                                      |
| Efficiency            | Data exchanged between stations has to travel through other nodes                 | Data exchanged between stations has to travel through central node                | Data exchanged between stations has to travel through other nodes                  | Direct data exchange between stations possible                                             |
| Robustness            | Weak, a link outage can effect more than one node                                 | Robust, a link outage only effects one node                                       | Weak, a link outage can affect more than one node                                  | Very robust, a link outage will not affect availability (using dynamic routing algorithms) |



**Figure 1** Schematic representation of an intra-station network, which connects the equipment at one seismic station, and the inter-station network combining all stations to one seismic network.

Independent of the implementation of the physical network links, the logical network design of the overall network should be planned very carefully. Addressing schemes can be based on public or private IP address ranges (Rekhter et al., 1996). Public IP addresses are assigned by

the Internet Corporation for Assigned Names and Numbers (ICANN, <http://www.icann.org/>) and are guaranteed to be globally unique to enable routing of data packets via the Internet. Usually public IP ranges are rented from Internet service providers (ISP), which implies an additional monthly service charge. If stations do not need Internet access and data is exchanged with the central node only, the use of private IP ranges can reduce operational costs and increase security, as well as reliability.

In a common scenario, each station runs its own private IP network providing addresses for the equipment installed on site. The intra-station networks are then combined into a larger IP network that can be easily routed at the data centre. Therefore IP addressing schemes should be discussed and designed in close cooperation with the IT- or network department of the institution that hosts the central node before the installation campaign starts. It will be difficult to remotely change intra-station IP configurations again after the network installation has been completed. Another aspect to consider is that sensor networks usually change and grow over time, be it with additional equipment added at stations or additional stations added to the network. The developed IP schema should also address this and ensure the scalability of the system.

## 4 Overview of current communication technologies

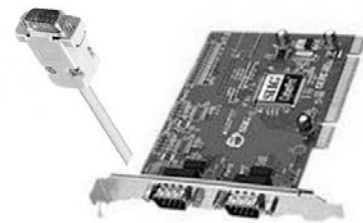
### 4.1 Wired communication technologies

Wired solutions are usually cheap, robust, and provide sufficient throughput capabilities to transport seismic data. However, they have strong limitations regarding the distance to be covered (cable laying effort). Generally, they only provide a good solution for the intra-station network connecting the components on site. Most seismological instruments are already equipped with communication interfaces, such as Ethernet cards or serial ports. The following list also indicates serial and Ethernet based solutions, if the distance at a station exceeds the standard limits of these technologies.

The industry standard ANSI EIA defines different specifications of communication links based on serial ports.

EIA-232 or RS-232 (Texas Instruments, RS-232, <http://www.datasheetarchive.com/ANSI%2FTIA%2FEIA-232-F-datasheet.html>) is the classic serial connection. It provides data rates up to 19200 Baud using a null modem cable with a maximum length of 15 m. Longer transmission lines can be used but will decrease the data rate (e.g., 300 m, 4800 Baud).

Equipment, such as computers, digitizers, or communication terminals often provide integrated RS-232 interfaces already (Figure 2). Configuration of the link is straight forward and easy by configuring baud rate, data bits, parity, stop bits, and flow control on both sides of the link.



**Figure 2** Example of a serial connector and a serial PCI card.

If distances or data rate requirements are too high to use RS-232, another option is the use of EIA-485 (RS-485) cards (Texas Instruments, RS-485, <http://www.datasheetarchive.com/TIA-485-A-datasheet.html>). This specification provides a cable length of 1200 m and 12 Mbit/s throughput, however, this maximum transfer rate is achieved only with cable lengths up to 15 m.

If RS-232 interfaces are already integrated into the equipment and only the distance to be covered is too large, RS-232 to RS-485 adaptors provide a valuable alternative (Figure 3). However, this will not increase the data rate provided by RS-232.



**Figure 3** Example of an RS-232 to RS-485 adaptor.

The Institute of Electrical and Electronic Engineers (IEEE) developed the standard IEEE 802.3, which defines different specifications of Ethernet based communication links (IEEE Standards Association, 802.3, <http://standards.ieee.org/about/get/802/802.3.html>).

The most commonly used specification nowadays is Fast Ethernet (100Base-T) providing up to 100 Mbit/s transmission rate over shielded or unshielded twisted pair copper cable (STP/UTP). Equipment to be connected needs a network interface card (NIC), which is often already integrated (Figure 4). An Ethernet switch then connects the different devices into one local network. The maximum cable length (segment size) on 100Base-T should not exceed 100 m between two devices. For the installation of the RJ45 connectors a crimping tool is needed. Link configuration is easy by configuring IP number, netmask, and gateway IP.



**Figure 4** Example of an RJ45 connector and a PCI based network interface card.

A 100Base-FX PCI card can be added to the equipment if a PCI slot is available. If the equipment already has integrated 100Base-T interfaces, an external 100Base-FX transceiver can be used to convert from 100Base-T to 100Base-FX (Figure 5). 100Base-FX supports a 100 Mbit/s transmission rate over fiber optic cables with a maximum cable length of 2000 m. In addition to supporting longer distances, fiber optic cables are immune to electrical hazards, such as lightning strikes and ground currents that may occur.



**Figure 5** Example of an external 100Base-FX transceiver and SC-connectors.

The fiber optic connectors are commonly known as "SC" connectors. Only qualified personnel may install these connectors and a range of special tools are needed (Fiber Stripper, Cleaver, Wipes, Kevlar Scissors). It is also possible to purchase pre-fabricated cables and take them to the site.

Another possibility to extend FastEthernet beyond the normal IEEE 802.3 limitation of 100 m is the use of a pair of VDSL Ethernet Extenders (Figure 6). With increasing cable length, data rates will decrease from the 100 Mbit/s provided by 100Base-T, but cables between the terminals can be extended up to some kilometers by using normal STP, UTP or even common RJ11 telephone cable. A cable length of 8000 m will limit the data rate to 128 kbit/s.



**Figure 6** Example of a VDSL Ethernet Extender.

#### 4.2 Wireless communication technologies

Wireless solutions are usually cheap and easy to install. They can be used to link longer distances, more than 1000 m. They provide sufficient throughput capabilities to transport seismic data. However, they are relatively sensitive to obstacles in the wave propagation path and to changing atmospheric conditions that can attenuate the signal. A clear "line of sight" between transmitter and receiver is enough for a robust connection for very short links only. Longer distances are very prone to RF signal attenuation based on rain fade and Fresnel zone obstruction. They require a proper link budget calculation to ensure stability (see 7.3.3 and 8.9.3). Wireless modems/ routers have the disadvantage of increased power consumption compared to wired solutions. Systems usually consume between 5-10 W on each side of the link, which impose additional demands on the power system at the station.

The WLAN standards IEEE 802.11g (2.4 GHz) and IEEE 802.11a (5 GHz) support data transmission over a wireless terrestrial link (IEEE Standards Association, 802.11,

<http://standards.ieee.org/about/get/802/802.11.html>).

Due to its widespread use, the cost of 2.4 GHz equipment is generally lower, but the frequency band is also used by other wireless communication technologies, such as Bluetooth and microwave systems, which can lead to harmful interference in populated areas. Running a wireless LAN link involves no operational costs as frequencies can be used freely. WLAN can cover distances from 50 to 500 m using omnidirectional antennas. Using directional antennas, the coverage can be extended to several km (Figure 7). The performance of some devices, especially in combination with directional antennas and a free Fresnel zone, allows connections over very long distances. In tests, links over up to 70 km have been achieved by the use of directional antennas. The maximum allowed radiated power of WLAN systems (calculated from transmission power and antenna gain) is usually limited by national regulations and must not be exceeded.



**Figure 7** Examples of omnidirectional and directional antennas and a WLAN modem.

The data transfer rate on both specifications, 802.11a and 802.11g, is 54 Mbit/s, although these rates are seldom achieved in practice. With the latest specification 802.11n maximum data rates can be increased up to 600 Mbit/s by expanding the channel width from 20 MHz to 40 MHz and establishing several physical input and output channels using multiple separate antennas.

Generally, an increase in link distance will decrease the maximum data rate provided by the system. Especially for long range links, the installation can become quite complex with very high mast poles and difficult alignment procedures. Link distances can also be doubled with the use of “repeater stations”.

A number of commercial products support data transmission over radio frequency (RF) telemetry links. The systems mainly differ in transmission rates, operating frequencies (VHF 160-200 MHz and UHF at 450 MHz and 900 MHz) and modulation techniques. Running the system involves no operational costs as the frequencies can be used freely, however, some frequency bands need operating permissions by national agencies.

Bidirectional broadband systems can provide transmission rates of 150 kbit/s. Some systems specialize on link distance and provide up to 150 km coverage without repeaters (Figure 8). With increasing link distance, the transmission rate decreases for a given transmitter power. Especially for long range links, the installation can become complex with very high mast poles and difficult alignment. Link distances can be doubled with the use of “repeater stations”.

Compared to WLAN links, RF links are more robust against rain attenuation due to the lower frequencies used. However, terminals are generally more expensive.

If available, it is also possible to use the public mobile phone network for data transmission.

Packet oriented standards extending the Global System for Mobile Communications (GSM), such as GPRS, UMTS and LTE support both transmission of voice and transfer of data. Some available wireless modems have built-in ability to connect to all three services (Figure 9). Cell availability is depending on the provider and usually limited to inhabited areas.

Provided data rates differ depending on the standard



**Figure 8** Example of an RF telemetry modem and a Yagi antenna.



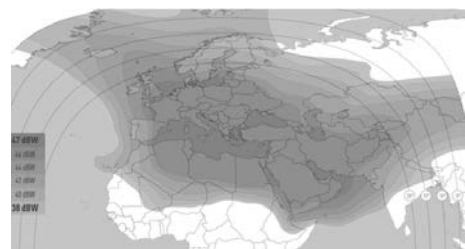
**Figure 9** Example of a mobile phone antenna and modem supporting all standards: GPRS, UMTS and TLE.

used (GPRS 53.6 kbit/s; 384 kbit/s; LTE 100 Mbit/s) but also on saturation of the shared infrastructure. Especially during and after catastrophic events, such as earthquakes, congestion or complete system blackouts are common. Configuration and installation of the modem is straight forward and easy. No antenna alignment is required.

### 4.3 Satellite communication technologies

At some remote stations satellite-based systems are the only alternative for seismic data transmission due to their availability at even very remote locations. Due to their independence from terrestrial infrastructure, they are very robust against failure during natural disasters. However, satellite capacities are generally shared with other companies/ services. Link saturation problems, which may occur, can be overcome by configuring bandwidth allocation methods such as committed information rates (VSAT) or constant bit rate modes (BGAN, Thuraya). Therefore, these systems provide a robust solution for early warning systems that require very high availability even during exceptional circumstances.

VSAT terminals represent the current state of the art measure for permanently installed seismic stations. Depending on their location, a suitable satellite (footprint, see Figure 10) and teleport operator must be selected. In principle, VSATs can be used worldwide, with the exception of the polar regions (latitudes larger than 70 degrees). Links can provide data rates up to several Mbit/s. Depending on data rate requirements of the station or the entire network, suitable bandwidth must be allocated.



**Figure 10** Example satellite footprint of Eutelsat 21B (source: [www.eutelsat.com](http://www.eutelsat.com)).

VSAT systems generally work on Ku-band, C-band, and Ka-band frequencies. C-band systems (smaller rain attenuation, larger antennas) are more common in Southeast Asia, Central and Southern Africa and Latin America, while Ku- and Ka-band systems (smaller antennas, but susceptible to rain attenuation) are more common in Europe, North Africa, Central Asia and North America. The influence of rain attenuation can also be mitigated by the use of adaptive coding and modulation techniques, e. g., utilized in DVB-S2. The required equipment consists of an indoor unit (modem) and an outdoor unit (Figure 11).



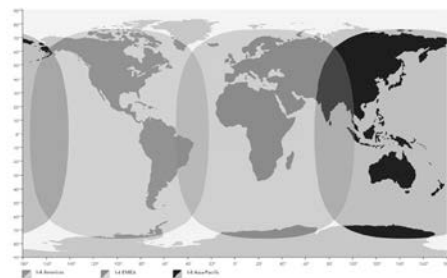
**Figure 11** Example of a VSAT modem (indoor unit) and parabolic antenna with receive and transmit unit (outdoor unit).

The outdoor unit (antenna, transmitter, and receiver) must be chosen depending on the calculated link budget. Transportation of antennas can cause problems, as typical sizes for Ku and Ka-band range between 75 cm and 1.8 m and C-band between 1.8 m and 2.4 m.

Another great disadvantage of a VSAT system is the high power consumption. Modems usually require around 25 W and additional 5-10 W per Watt of transmit unit power (BUC), which adds up to an overall consumption of approximately 35-70 W. Complex tests (cross-polarization and 1 dB compression test) must be performed, together with the satellite or teleport provider, during installation in order to activate the link. Compared to other satellite communication systems, the operational costs are low (see Table 3).

The Inmarsat BGAN satellite communication system uses three geostationary satellites, which provide global coverage up to 70 degrees latitude. The system does not cover the polar regions (Figure 12). It allows the transmission of data with a maximum upload rate of 240 kbit/s. Data volume-based tariffs (usually cheaper) or time-based tariffs (for real-time transfers) can be selected. Compared to VSAT, the transfer costs are very high in both modes. The system should be deactivated for times when no data transfer is taking place.

There are BGAN terminals of different manufacturers on the market. All of the devices are very compact, comparable to the size of a laptop, weighing less than 3 kg and have relatively low power consumption (less than 25 W). Modem and antenna are usually combined in one piece of transportable outdoor equipment (Figure 13). During installation, the terminal antenna must be roughly aligned to one of the three BGAN satellites (5 degrees accuracy sufficient). No complicated activation tests are required, however the SIM card must be commissioned/ activated by the BGAN provider before it can lock to the network.



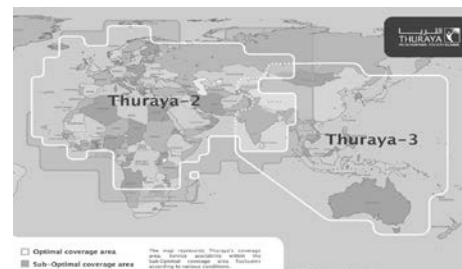
**Figure 12** BGAN coverage map (source: [www.inmarsat.com](http://www.inmarsat.com)).



**Figure 13** Example BGAN modem with integrated antenna.

Thuraya provides a satellite telephone service based on two geostationary satellites covering Europe, Australia, the Middle East, Africa, Central and Southeast Asia (Figure 14). It sells two data transmission terminals: Thuraya IP and Thuraya IP Plus.

Both terminals can be used in two data transmission modes: “Standard IP” has a variable bit rate and offers up to 444 kbit/s data connectivity on a shared mode or best effort basis. “Streaming IP”, also known as Constant Bit Rate offers dedicated data connectivity ranging from 16 kbit/s to 384 kbit/s.



**Figure 14** Thuraya coverage map (source: [www.thuraya.com](http://www.thuraya.com)).

Modem and antenna are integrated in one piece of equipment weighing less than 2 kg (Figure 15). Power consumption of both terminals is less than 25 W while transmitting. Prices for terminals and transmission are comparable to Inmarsats BGAN system.



**Figure 15** Thuraya IP Plus modem with integrated antenna.

The Iridium satellite telephone system provides global coverage for all latitudes, including the poles. Since the 66 cross-linked low-Earth orbiting (LEO) satellites have polar orbits, the coverage above the poles is even particularly good (Figure 16). The antenna of the Iridium terminal need not be aligned with a satellite, hence installation is very simple.



**Figure 16** Iridium global coverage map (source: [www.iridium.com](http://www.iridium.com)).

Iridium offers different data services. The only broadband data transmission OpenPort provides data rates up to 128 kbit/s. The tariff plans of the OpenPort system are more complicated compared to BGAN, and transmission costs are generally higher. Both parts of the system (ADE (above decks equipment) – antenna and BDE (below decks equipment) – modem) are easy to install (Figure 17). No activation tests are required, however SIM cards must get commissioned/ activated by the provider. The power consumption in data transfer mode is less than 25 W.



**Figure 17** Iridium antenna (ADE) and OpenPort terminal (BDE).

Launching in 2015, Iridium NEXT will substantially enhance and extend Iridium services, delivering higher data speeds (L-Band: 1.5 Mbit/s, Ka-band: 8 Mbit/s) (Iridium, Iridium Next, <http://www.iridium.com/about/iridiumnext.aspx>).

## 5 Conclusions

All technologies discussed in the previous paragraph provide certain advantages and disadvantages, depending on the particular issues under consideration. It is important to quantify the differences and compare them to see if they really matter. Moreover, even there may be more than one suitable technology. It is obvious that choosing the right one can reduce the overall costs of the project.

Intra-station networks can be implemented using wires, based on the provided interfaces integrated in the seismic equipment (serial or Ethernet). Using adaptors, distances can be



easily extended up to several hundreds of meters, if needed. To avoid extensive cable laying work, wireless LANs can also be an alternative in unpopulated areas, where interferences are not expected.

**Table 3** Comparative matrix for communication systems used in seismic networks. For column Installation/ Activation, Easy/ Moderate/ Complex have been defined as follows: Easy - no expert knowledge required; Moderate - basic network knowledge sufficient; Complex specialist knowledge/or special tools needed. Numbers indicated in columns Hardware- and Operating Costs are estimations in \$US and should only be used for relative comparison.

|                                    | Type                         | Distance/<br>Coverage                                               | Data Rate                                                                 | Installation/<br>Activation           | Hardware<br>Costs                       | Operating<br>Costs   | Link<br>Reliability                             | Power<br>Consumption |
|------------------------------------|------------------------------|---------------------------------------------------------------------|---------------------------------------------------------------------------|---------------------------------------|-----------------------------------------|----------------------|-------------------------------------------------|----------------------|
| Wired Communication                | RS232<br>(Serial)            | 15 m (19200 Baud), 300 m (4800 Baud)                                | <20 kbit/s (depending on cable length)                                    | Easy                                  | PCI Card <50                            | None                 | Very robust                                     | <1 W                 |
|                                    | RS485<br>(Serial)            | <1200 m                                                             | <12 Mbit/s                                                                | Easy                                  | Adaptor <10                             | None                 | Very robust                                     | <1W                  |
|                                    | 100Base-T<br>(FastEthernet)  | <100 m                                                              | <100 Mbit/s                                                               | Easy                                  | PCI Card <20                            | None                 | Very robust                                     | <1 W                 |
|                                    | 100Base-FX<br>(Fibre Optics) | <2000 m                                                             | <100 Mbit/s                                                               | Complex                               | PCI card<150<br>Adaptor<300             | None                 | Very robust                                     | <1 W                 |
|                                    | VDSL<br>Ethernet<br>Extender | <8000 m                                                             | <100 Mbit/s (depending on cable length)                                   | Easy                                  | <200 per piece                          | None                 | Very robust                                     | <5 W                 |
| Wireless terrestrial Communication | WLAN                         | <70 km                                                              | 802.11a/g <54 Mbit/s<br>802.11n < 600 Mbit/s (depending on link distance) | Moderate, complex on long range links | Modem & antenna<br>50 – 500             | None / private link  | Good                                            | <10 W                |
|                                    | RF Links                     | <150 km                                                             | <200 Kbit/s (depending on link distance)                                  | Complex                               | Modem & antenna<br>300 – 3000           | None / private link  | Good                                            | <10 W                |
|                                    | Mobile phone network         | Provider dependent                                                  | GPRS <53.6 Kbit/s<br>UMTS <384 Kbit/s<br>LTE <100 Mbit/s                  | Moderate                              | Modem & antenna<br>100 – 1000           | Low<br><0.05/MB      | Bad, public shared link                         | <5 W                 |
| Wireless sat-based Communication   | VSAT                         | Global, depending on satellite footprint, excluding pole regions    | <5 Mbit/s                                                                 | Complex and heavy equipment           | Modem & antenna<br>1500-4000            | Low<br><0.05/MB      | Good, using committed information rate settings | >25 W                |
|                                    | Inmarsat BGAN                | Global, excluding pole regions                                      | <240 kbit/s                                                               | Moderate                              | Modem & integrated antenna<br>2000-4000 | High<br>~5,-/MB      | Good, in constant bit rate mode                 | <25 W                |
|                                    | Thuraya IP Plus              | Europe, Australia, Middle East, Africa, central- and southeast Asia | 16 kbit/s to 384 kbit/s in streaming mode                                 | Moderate                              | Modem & integrated antenna<br>~3000     | High<br>~5,-/MB      | Good, in constant bit rate mode                 | <25 W                |
|                                    | Iridium OpenPort             | Global, including pole regions                                      | <128 kbit/s                                                               | Moderate                              | Modem & antenna<br>~4000                | Very high<br>>5,-/MB | Good                                            | <25 W                |

Local inter-station networks can be set up using WLAN and directional antennas if stations are in less than 70 km range. Using RF links, distances can reach up to 150 km at most. Both systems are prone to interference with other wireless systems in urban areas and sound link budget calculations should be performed which require special knowledge and tools (e.g. a

spectrum analyser). The system distance limitations can be extended by the use of repeater stations, but this will add complexity to the communication network and complicate troubleshooting and maintenance.

In case mobile phone reception exists at the station and reliability during earthquakes is not of highest importance, mobile phone modems can be a low priced and very effective alternative.

For regional and global networks, satellite based communication systems are available even in the remotest areas, but come with a large price tag. For applications requiring high availability during global/ regional disasters, it is reasonable to rely on space-based infrastructures. VSAT systems have lower operational costs, but are difficult to transport, install, and activate. The high power consumption of VSATs increases the installation costs further. However, for permanent streaming from seismic stations with sampling rates higher than 10 Hz, VSATs are currently the only financially feasible selection as operational costs of the satellite-based alternatives are very high.

For temporary installations or very low sampling (data) rates Inmarsats BGAN or Thuraya are a good solution. These systems are easy to transport, install and activate. For the polar regions above the 70<sup>th</sup> latitude, Iridium provides the only satellite-based communication solution. These communication technologies may become a good alternative for permanent seismic networks in the future if prices will further decrease. Another way to allow the extensive usage of these technologies will be to complement existing seismic networks with stations running the pre-processing of waveform data on-site, if a computer is available. This would drastically reduce data transfer requirements of the inter-station network, because only a few parameters, such as picks, amplitudes, peak ground acceleration, velocity, etc. may be transmitted to the data centre and not the complete high sampling rate waveforms.

## Acknowledgments and comments

We thank P. Bormann for additional editing of this Information Sheet. We are also grateful to the reviewers C. Milkereit and K. Wendlandt, which improved the text significantly. P.L. Evans and S. Gensch kindly amended our English.

Identifying the name of the instruments' manufacturers is not meant to be an endorsement of their products. Prices and costs mentioned in this Information Sheet can vary a lot in time and depending on regions/ countries. They cannot be used as absolute values but only for relative comparison. The list of manufacturers provided in Appendix 2 is a subjective selection and is by no means complete.

## References

Rekhter, Y., Moskowitz, B., Karrenberg, D., de Groot, G., and E. Lear, Address Allocation for Private Internets, *BCP 5, RFC 1918*, February 1996.

Angermann M., Guenther M., and Wendlandt K., Communication architecture of an early warning system. *Nat. Hazards Earth Syst. Sci.*, **10**, 2215–2228, 2010.

## Appendix 1

### Example design of a communication infrastructure for a seismic network of 25 stations used for tsunami early warning in the Indian Ocean

#### Step 1) Requirement Analysis

Continuous seismic data streams are needed at the data centre to calculate earthquake parameters in real-time, identify if there is a tsunami potential and send out a timely warning to the population under threat.

- Stations must be constantly online, streaming data on permanent communication links to the data centre.

The network is permanently deployed and will continuously be enlarged.

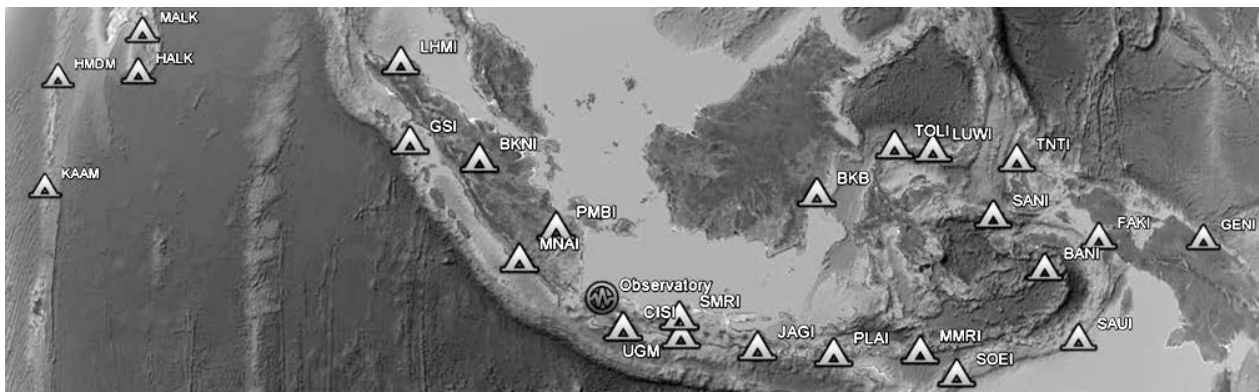
- Operational and maintenance costs outweigh initial investments in equipment, training, transportation, installation and activation.

Each station hosts two types of sensors, a very broad band seismometer and an accelerometer, both having three components. Selected sampling frequencies are 50 Hz for the seismometer and 100 Hz for the accelerometer, both connected to a 24 bit digitizer.

- $3 \text{ components} * 50 \text{ Hz} * 24 \text{ bit} + 3 \text{ components} * 100 \text{ Hz} * 24 \text{ bit} = 10800 \text{ bit/s}$ .

The SeedLink protocol is used to transfer station data to the observatory. SeedLink uses Stein compression, which can almost halve the data quantity (see 8.6.6).

- The data to be transmitted using the communication system varies between 5-10 kbit/s per link depending on compression efficiency.



**Figure 18** Geographical distribution of the stations and observatory location.

The deployment area of the network is along the Sunda arc covering several thousand kilometres (Figure 18).

- Link distances between stations are generally over 500 km.

Stations are in very remote areas and located in three different countries. To reach stations, long and expensive journeys are necessary with several connecting flights.

- A secondary backup communication system would be useful to reduce maintenance costs.

High network availability is required, being specified as above 99 %. Data streaming must be ensured during and immediately after the occurrence of large earthquakes.

- Link robustness is of highest importance, higher than budget constraints.

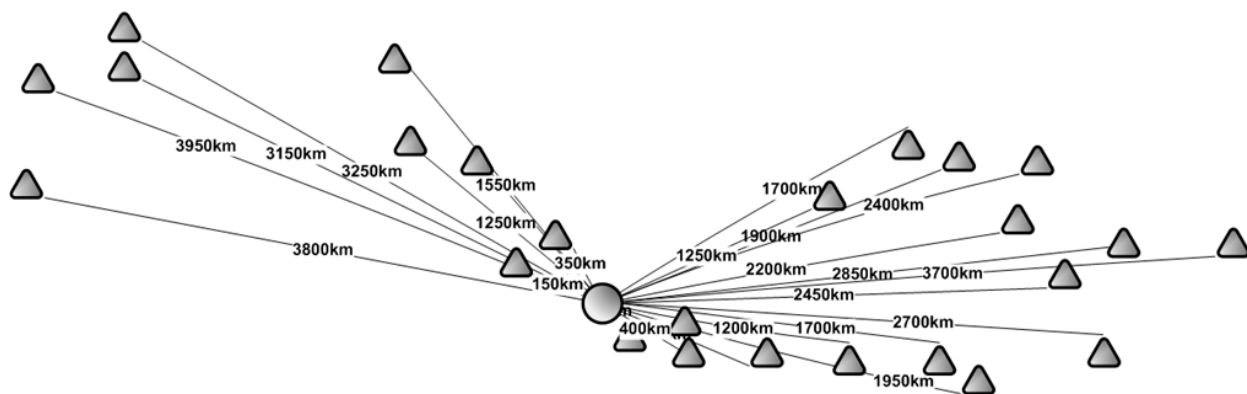
Table 4 summarizes the requirements of the seismic network used for tsunami early warnings that are relevant for the communication system selection.

**Table 4** Similar to Table 1 but tailored for a tsunami early warning system.

| Application Type             | Geographic Spread                                 | Sampling/<br>Data Rates | Link Type                                  | Required Robustness | Remoteness of Stations |
|------------------------------|---------------------------------------------------|-------------------------|--------------------------------------------|---------------------|------------------------|
| Tsunami Early Warning System | Network:<br>>8000 km<br>Inter-station:<br>>500 km | 5-10 kbit/s per station | Continuous link for real-time data streams | Very high           | Very remote            |

### Step 2) General design considerations for the communication system of the seismic network

Requiring no data transfer between stations, but having a central data centre collecting all waveforms directly from remote sites, the “natural” topology in regards to the general data flow is a star topology (Figure 19). The analysis also showed that link reliability and real-time data availability are of high importance. The star topology increases robustness, as a single link outage can only affect one station. Compared to other topologies, it simplifies troubleshooting and maintenance as there are no multi-hop links to reach any of the stations.



**Figure 19** Network topology with link distances indicated in kilometres.

To simplify maintenance efforts, all stations are deployed with common hardware and common configurations. Data rate requirements are the same on each link with a maximum transmission rate of 10 kbit/s (assuming no compression). Depending on the compression efficiency, the data rate will vary over time and, under normal conditions, may be as low as 5-7 kbit/s. This leaves some spare bandwidth that can be used to transmit buffered data

collected during communication/ power outages as soon as the link is re-established. The acquisition protocol (SeedLink) is capable of filling gaps, acquiring old data, and continuing to acquire real-time data as soon as gaps are filled (see 8.6.4). As a general rule of thumb, 30% of the total link budget should be allocated for this purpose. This also allows connecting to a station for trouble shooting purposes without affecting the real-time streaming.

**Table 5** The implemented IP plan ensures scalability and simplifies maintenance in regards to troubleshooting and routing at the data centre.

| Inter-station network IP | Station                          | Intra-station network IP | Number of hosts at station |
|--------------------------|----------------------------------|--------------------------|----------------------------|
| 172.16.0.0/16            | Station 1                        | 172.16.1.0/24            | 254                        |
|                          | Station 2                        | 172.16.2.0/24            | 254                        |
|                          | Station $n$                      | 172.16. $n$ .0/24        | 254                        |
|                          | Last Station(255 <sup>th</sup> ) | 172.16.255.0/24          | 254                        |

Using the scheme shown in Table 5, a total of 255 stations are possible, each with up to 254 devices within the intra-station network. A private IP block was chosen, because equipment at the stations does not access the Internet and does not need to be reachable from there. Data is directly transferred from stations to the data centre and there it is shared with other national and international observatories using a central acquisition server reachable from outside.

### Step 3) Selection of the communication technology

According to Table 3, the geographical extent and large inter-station distances drive the selection of the suitable communication technology towards mobile phone networks or satellite-based solutions. Mobile phone networks are excluded because they cannot provide the required link reliability for a tsunami early warning system.

Out of the four satellite-based solutions introduced in paragraph 4.3, VSAT was chosen as the best candidate. Table 6 shows rough cost estimates, from which it is apparent that VSAT is considerably cheaper in the long term. The high operational costs of BGAN/ Thuraya/ Iridium outweigh the additional investments needed to install a VSAT terminal after just a few weeks.

**Table 6** Relative cost comparison for satellite-based communication systems. Initial investment cost estimation includes: price estimation for communication equipment and investment for independent power supply system (based on photovoltaic and batteries). Operational cost estimation is based on the assumption that a station produces approximately 100 MB data per day. VSAT contracts are usually priced in kbit/s with a minimum duration of one month.

| Estimated Costs per Station | VSAT      | BGAN and Thuraya | Iridium     |
|-----------------------------|-----------|------------------|-------------|
| Initial Investment:         | \$US 7000 | \$US 4000        | \$US 5000   |
| Operational Costs:          |           |                  |             |
| per week                    | \$US 200  | \$US 3500        | \$US 4900   |
| per month                   | \$US 200  | \$US 15000       | \$US 21000  |
| per year                    | \$US 2400 | \$US 180000      | \$US 250000 |

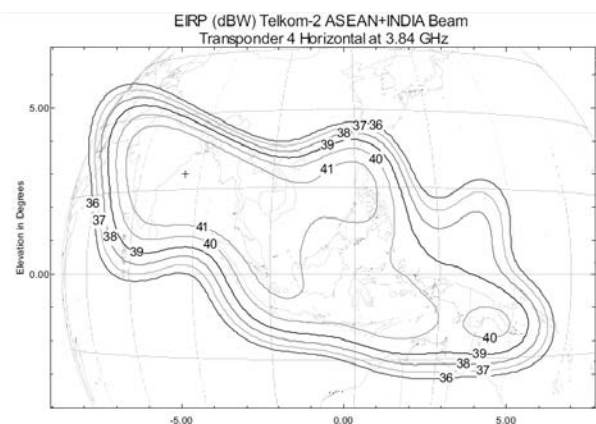
VSAT systems, when set up in star topology, have a dedicated station called hub that is the gateway for all stations. Additionally the hub has management and control functions, such as defining carrier frequencies, creating time plans and orchestrating the overall system by assigning transmit slots to each station. The VSAT hub is installed at the data centre, because qualified sat-com professionals are employed at the tsunami observatory.

Hub services can also be rented from teleport providers, which usually run VSAT hub services on different satellites at their teleport location. However, this will add an additional data link from the teleport to the observatory. Using the Internet to do so will add an unreliable link to the system, while a VSAT-based connection will double the required VSAT bandwidth and involved operational costs.

There are different manufacturers of VSAT platforms on the market that differ in protocols, modulation techniques used, and also in prices for the modems. Hub and stations must run with equipment of the same platform type (VSAT modem). Hub services rented via a teleport provider dictate therefore, which modem model needs to be purchased for a station.

All VSAT platforms can run on Ku-band, C-band, and Ka-band frequencies. The International Telecommunication Union (ITU) has categorized the region as a Region P, which designates regions with very high rain precipitation (ITU, Characteristics of precipitation for propagation modelling, [http://www.itu.int/dms\\_pubrec/itu-r/rec/p/R-REC-P.837-1-199408-S!!PDF-E.pdf](http://www.itu.int/dms_pubrec/itu-r/rec/p/R-REC-P.837-1-199408-S!!PDF-E.pdf)). Generally, rain attenuation increases with the frequency of the transmitted signal, making C-band the standard for reliable communication in tropic regions.

After the selection of the frequency band a suitable satellite must be chosen. This can be done by checking the coverage area of different satellites (Satellite footprint maps, <http://www.satbeams.com/footprints>). A so-called satellite footprint map shows coverage areas enclosed by contour lines indicating the level of EIRP (Effective Isotropic Radiated Power), a value representing the satellite signal strength in those areas. Figure 20 shows the footprint of the satellite selected for this case study. Based on the EIRP levels, link budget calculations done by qualified sat-com professionals will define dish size and transmit unit characteristics needed at each seismic station.

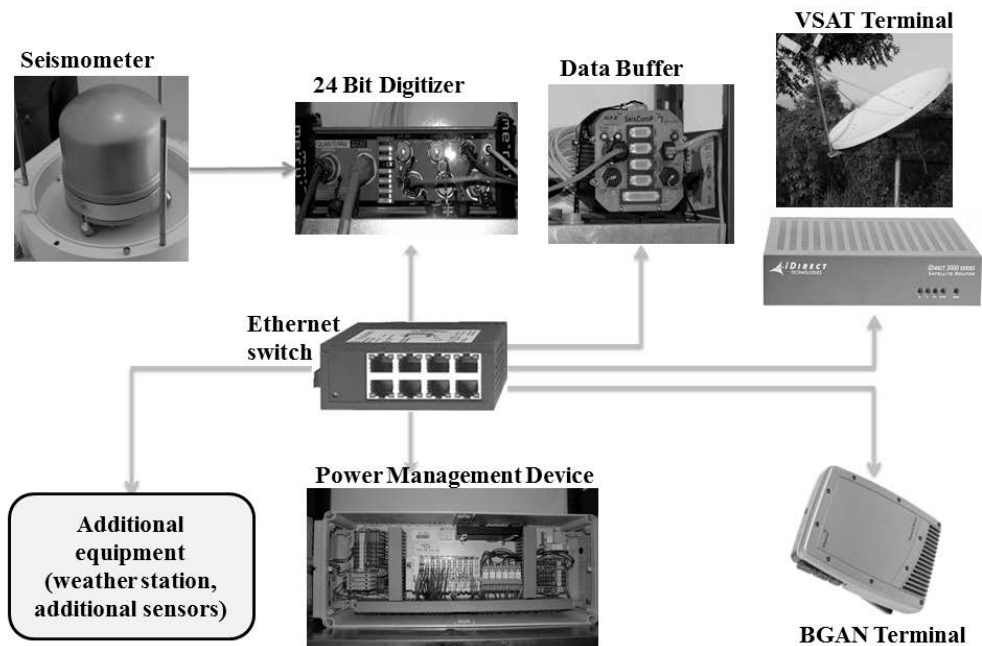


**Figure 20** The footprint of the Telkom 2 communication satellite covers the area of the seismic network.

#### Step 4) Implementation of the communication system

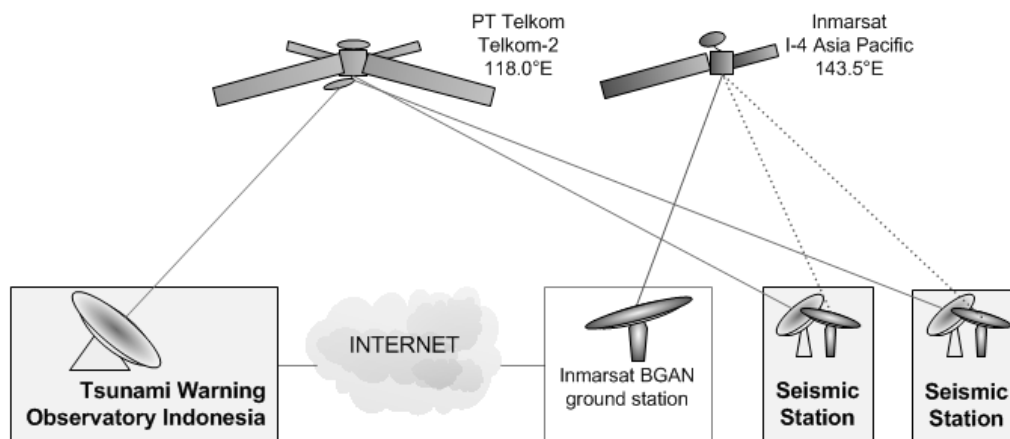
The intra-station network at all stations is based on Ethernet interfaces and a switch (Figure 21). The VSAT modem is configured to be the default gateway for all equipment on-site (intra-station network), so waveform data is transmitted using the VSAT network. In order to reduce maintenance costs, BGAN terminals are added at each station to provide a secondary

backup link. This link is only initiated for remote login and troubleshooting in case of a VSAT link failure.



**Figure 21** Intra-station network based on Ethernet using a switch.

The inter-station VSAT network is operated from the hub, which is located directly at the data centre. The BGAN network terminates at a ground station of the company offering the service. Using BGAN the traffic will pass the Internet before reaching the stations (Figure 22).



**Figure 22** Inter-station network and backup link.

The case study presented was implemented in the GITEWS project (GITEWS; <https://www.gitews.de>) as the backbone seismic network for the Indian Ocean Tsunami Early Warning System and is currently operational. For further information about the communication architecture and the selection and installation process, please refer to Angermann et al. 2010. The network is under continuous development and more seismic stations, as well as other sensors, are continuously added to the system.

## Appendix 2

## List of manufacturers and products

The following list of communication equipment manufacturers provides more details for further reading.

| Category                             | System       | Product                               | Manufacturer     | Web Link                                                                          |
|--------------------------------------|--------------|---------------------------------------|------------------|-----------------------------------------------------------------------------------|
| Wired communication technologies     | RS-232       | PCI cards                             | StarTech:        | <a href="http://www.startech.com/">http://www.startech.com/</a>                   |
|                                      | RS-485       | Cards and converter                   | SerialComm:      | <a href="http://www.serialcomm.com/">http://www.serialcomm.com/</a>               |
|                                      |              |                                       | CommFront:       | <a href="http://www.commfront.com">http://www.commfront.com</a>                   |
|                                      | 100Base-T    | PCI cards                             | D-Link:          | <a href="http://www.dlink.com/">http://www.dlink.com/</a>                         |
|                                      |              | STP cable                             | Belden:          | <a href="http://www.belden.com/">http://www.belden.com/</a>                       |
|                                      | 100Base-FX   | PCI cards                             | D-Link:          | <a href="http://www.dlink.com/">http://www.dlink.com/</a>                         |
|                                      |              | Fibre cable                           | Corning:         | <a href="http://www.corning.com/">http://www.corning.com/</a>                     |
|                                      |              | Transceiver                           | EtherWAN:        | <a href="http://www.etherwan.com/">http://www.etherwan.com/</a>                   |
|                                      | VDSL         | Ethernet extender                     | StarTech:        | <a href="http://www.startech.com/">http://www.startech.com/</a>                   |
|                                      |              |                                       | Allied Telesis:  | <a href="http://www.alliedtelesis.com/">http://www.alliedtelesis.com/</a>         |
| Wireless communication technologies  | Wireless LAN | Modems                                | Cisco:           | <a href="http://www.cisco.com/">http://www.cisco.com/</a>                         |
|                                      |              |                                       | Avalan:          | <a href="http://www.avalanwireless.com/">http://www.avalanwireless.com/</a>       |
|                                      |              | Long range systems                    | Ubiquiti         | <a href="http://www.ubnt.com/">http://www.ubnt.com/</a>                           |
|                                      |              |                                       | Afar:            | <a href="http://www.afar.net/">http://www.afar.net/</a>                           |
|                                      | RF telemetry | Modems                                | Avalan:          | <a href="http://www.avalanwireless.com/">http://www.avalanwireless.com/</a>       |
|                                      |              |                                       | Radiometrix:     | <a href="http://www.radiometrix.com/">http://www.radiometrix.com/</a>             |
|                                      |              | Long range systems                    | Freewave:        | <a href="http://www.freewave.com/">http://www.freewave.com/</a>                   |
|                                      | Mobile phone | Modems                                | Cisco:           | <a href="http://www.cisco.com/">http://www.cisco.com/</a>                         |
|                                      |              |                                       | LyconSys:        | <a href="http://www.lyconsys.com/">http://www.lyconsys.com/</a>                   |
|                                      |              |                                       | NetModule:       | <a href="http://www.netmodule.com/">http://www.netmodule.com/</a>                 |
|                                      |              |                                       | Insys:           | <a href="http://www.insys-icom.com/">http://www.insys-icom.com/</a>               |
| Satellite communication technologies | VSAT         | Parabolic antennas                    | Prodelin:        | <a href="http://www.prodelin.com/">http://www.prodelin.com/</a>                   |
|                                      |              |                                       | Skyware:         | <a href="http://www.skywareglobal.com/">http://www.skywareglobal.com/</a>         |
|                                      |              |                                       | Andrew:          | <a href="http://www.andrew.com/">http://www.andrew.com/</a>                       |
|                                      |              | Receive and transmit units (LNB, BUC) | Belcom:          | <a href="http://www.belcommicrowaves.com/">http://www.belcommicrowaves.com/</a>   |
|                                      |              |                                       | Terrasat:        | <a href="http://www.terrasatinc.com/">http://www.terrasatinc.com/</a>             |
|                                      |              |                                       | New Japan Radio: | <a href="http://www.njr.com/">http://www.njr.com/</a>                             |
|                                      |              |                                       | Advantech:       | <a href="http://www.advantechwireless.com/">http://www.advantechwireless.com/</a> |
|                                      |              | VSAT platforms/ Modems                | IDirect:         | <a href="http://www.idirect.net/">http://www.idirect.net/</a>                     |
|                                      |              |                                       | Hughes:          | <a href="http://www.hughes.com/">http://www.hughes.com/</a>                       |
|                                      |              |                                       | Gilat:           | <a href="http://www.gilat.com/">http://www.gilat.com/</a>                         |
|                                      |              |                                       | Viasat:          | <a href="http://www.viasat.com/">http://www.viasat.com/</a>                       |
|                                      |              |                                       | Nanometrics:     | <a href="http://www.nanometrics.ca/">http://www.nanometrics.ca/</a>               |
|                                      |              | Cables and connectors                 | Belden:          | <a href="http://www.belden.com/">http://www.belden.com/</a>                       |
|                                      |              |                                       | Cablecon:        | <a href="http://www.cabelcon.dk/">http://www.cabelcon.dk/</a>                     |
|                                      | BGAN         | Terminals                             | Thrane & Thrane: | <a href="http://www.cobham.com/">http://www.cobham.com/</a>                       |
|                                      |              |                                       | Addvalue:        | <a href="http://www.wideye.com.sg/">http://www.wideye.com.sg/</a>                 |
|                                      |              |                                       | Hughes:          | <a href="http://www.hughes.com/">http://www.hughes.com/</a>                       |
|                                      | Thuraya      | Terminals                             | Thuraya:         | <a href="http://www.thuraya.com/">http://www.thuraya.com/</a>                     |
|                                      | Iridium      | Terminals                             | Iridium:         | <a href="http://www.iridium.com/">http://www.iridium.com/</a>                     |



| Topic   | Seismic waveform data retrieval                                                                                                                                             |
|---------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Joachim Saul, GFZ German Research Centre for Geosciences,<br>Telegrafenberg, 14473 Potsdam, Germany<br>E-mail: <a href="mailto:saul@gfz-potsdam.de">saul@gfz-potsdam.de</a> |
| Version | December 2014; DOI: 10.2312/GFZ.NMSOP-2_IS_8.3                                                                                                                              |

## Abstract

This information sheet gives an overview of some of the currently most popular and efficient tools for the retrieval of seismic waveform data from data centers. We do not attempt to provide a complete list of tools or exhaustive documentation. Instead it is our aim to provide some guidance and ideally whet the reader's appetite.

## 1 Overview

Digital recording methods have been used in seismology since about 1970. Digital methods have greatly increased data quality but they have also created new challenges. In order to allow efficient data exchange, the data had to be homogenized across different computer software and architectures. At the same time the exchange techniques had to keep pace with evolving new technologies. In particular, the advent of the Internet a little over 20 years ago and the continuously decreasing costs for network bandwidth have had great impacts on the way seismic data are acquired, archived, exchanged and processed today.

This chapter gives an overview of the techniques that are now (2014) “state of the art” for retrieving seismic waveform data from data archives. Data retrieval methods can be basically categorized into three groups:

- interactive, web-based solutions
- non-interactive techniques suitable for automated batch-type requests
- continuous, real-time data transfer

We will present some of the more popular solutions for each of these groups and discuss their individual merits and disadvantages.

### 1.1 Definition of the subject

We want to request seismic waveform data in order to display, analyze or otherwise use them. In general, the request of waveform data takes place at the beginning of a possibly long work flow, which is, of course, very application specific. Data requests nonetheless do very often follow a certain sequence of steps:

#### A. Discovery of

- seismic events based on hypocenter location, magnitude, type, etc.
- seismic stations based on sensor location (often within a specific epicentral distance range), instrument characteristics, etc.

- a combination of both in order to highlight a certain region, e.g., for tomographic studies
- B.** Selection of stations and channels based on what we found in step A. This selection process often combines event and station criteria.
- C.** Selection of time windows, often based on a combination of event and station locations
- D.** Formulation and submission of the request
- E.** Wait
- F.** Pick-up of the data
- G.** Data pre-processing (extraction, conversion, scaling, etc.)
- H.** Data processing - the actual data analysis, display, etc.

The subject of this information sheet is to provide some guidance to the novice user through steps A-F. Our aim is not to provide a comprehensive recipe, but to discuss concepts, weaknesses and strengths of various data request techniques. For more in-depth and technical documentation, the (potential) user is referred to external documents.

## 2 Relevant data formats used for data exchange

Chapter 10.3 gives a comprehensive overview about data formats commonly used in seismology, both for parametric and waveform data. In the following, we briefly summarize only those data formats, which are currently the most relevant for the data retrieval from seismic data archives.

### 2.1 SEED format

The SEED data format (SEED - Standard for the Exchange of Earthquake Data) was designed for use by the earthquake research community, primarily for the exchange between institutions of unprocessed raw waveform data. It's specification is part of the SEED manual (FDSN, 2012). Today, SEED is also the format that is most commonly provided to the end user by all major seismic data centers, including IRIS DMC, GEOFON, ORFEUS DC, GEOSCOPE and other data centers of the FDSN.

SEED is a binary format and while it is in fact quite complicated, there is de-facto standard software available for decoding and thus helping the end user to convert it to formats used by the most common data processing software packages. This conversion software (rdseed), available from IRIS and add the URL, allows the seismologist to conveniently decode SEED and encode the data into application specific formats like the SAC or AH formats.

A SEED volume consists of an arbitrary number of so-called “blockettes”. The format of these blockettes is described in the SEED manual. There are many different blockette types for raw waveform data as well as for metadata describing the raw data (e.g., station location,

channel gain to convert digital counts into units of ground motion, etc.). Raw waveform data and metadata may also be shipped separately, then normally referred to as miniSEED and dataless SEED, respectively.

## 2.2 MiniSEED format

The miniSEED format is the raw data part of the SEED format. It only consists of those blockette types that are used for storing information about the actual waveforms. That includes the encoded waveform data themselves, but also basic metadata like station code, channel code, start time of a data record and -optionally- timing quality. However, neither event nor instrument response information (not even a gain factor) is encoded in MiniSEED, because these metadata generally change very rarely if at all. They are therefore normally transmitted separately from the waveforms (often in the form of dataless SEED). This makes miniSEED a very efficient, light-weight format for archival as well as real-time exchange. MiniSEED consists of fixed-size blockettes as SEED. A miniSEED data blockette is also often referred to as a “record”. The physical size of a record, which is the smallest block containing contiguous data, is normally a multiple of 512 bytes. The most commonly used record sizes are 512 or 4096 bytes, but 512 bytes is presently the most common record size and is also used by data exchange protocols such as SeedLink described below. Generally the number of data samples that can be stored within one miniSEED record depends on the sampling frequency and efficiency of data compression, besides the physical record size. A 512-byte record can store approximately 400 samples of data if data compression is used, corresponding to data snippets of 20 seconds length at a sampling rate of 20 Hz. For 100-Hz data, the data snippets are accordingly shorter (approximately 4 seconds). This causes a latency of typically 20 seconds in acquiring 20-Hz data, which might be of concern for certain early-warning applications. Such applications, however, commonly record data at a higher sampling rate like 100 Hz, which already dramatically reduces the latency to less than 5 seconds. The processing can be further reduced if the data recorder sends out the data records prior to filling them with the maximum possible number of samples. Currently miniSEED is the basis for or is at least supported by all relevant real-time data processing and archival systems.

## 2.3 Dataless SEED format

While miniSEED contains only the raw waveforms (usually in digital counts) without any metadata, dataless SEED contains only metadata such as station location and instrument response without any waveform data. This separation allows storage, exchange and configuration of station inventories with the smallest possible overhead.

The main disadvantage of SEED is that it is not easily extensible. For instance the station identifier consists of up to five alphanumeric characters. This limitation is simply due to a pre-defined fixed space limit. There are many other limitations in SEED and the FDSN has therefore decided to introduce a new, XML (eXtensible Markup Language)-based format for station metadata. Dataless SEED will nevertheless continue to be of importance for the seismological research community for many years to come.

## 2.4 FDSN stationXML

Being an XML (eXtensible Markup Language) based format, FDSN Station XML was designed to overcome the limitations of dataless SEED, particularly the limited extensibility. The format has been approved by the FDSN in 2013 and is meant to become – in the long term -- a replacement of the dataless SEED format.

Some data centers and services make their station inventories available as FDSN StationXML already. Software exists for both the conversion to/from dataless SEED as well as for direct use as inventory data format.

## 2.5 Further information

For further information about the FDSN standard formats the reader is referred to

- the SEED format version 2.4 specification (FDSN 2012)
- the FDSN StationXML specification (FDSN 2013)

## 3 Access to archive data

20 years ago, the access to seismic data could be a lengthy and time-consuming procedure, especially for large data volumes. Requests had to be submitted to the data centers and data were returned, e.g., on magnetic cartridges or CD's. Due to the availability and greatly reduced costs of high-speed Internet connections, the data exchange through physical media is now obsolete (2014). The Internet can be and is nowadays used even for synchronization of large data archives. E.g., seismic noise studies utilize huge amounts of data that can now be transferred over the Internet. The cost of hard disk space has decreased greatly, with the benefit for data centers of being able to now store their entire archives online on hard disk, which has brought additional dramatic speed advantages.

This offers many new possibilities for data centers to offer their data in ways more convenient to the users than ever before. While it may have taken weeks to process data requests at the data center and to ship the data tapes to the user, it is now a matter of minutes or at worst hours between the arrival of the user request and the data shipment. Some request mechanisms are even fast enough to provide data "on the fly" within seconds. New approaches for data archival, like distributed data archives, have become possible.

Below we will describe interactive and non-interactive data access techniques.

### 3.1 Interactive methods

In this section we will explore several web pages that provide interactive access to seismic waveform data. At each of the data portals we will perform an example data request for broadband waveforms of the 2011 Tohoku earthquake. We are interested in teleseismic P waves and want to request all public broadband data from stations within 90 degrees epicentral distance. For the P waves we can restrict our request to the vertical component only. Due to the magnitude of the earthquake and long rupture duration, we request time

windows of 5 minutes before to 10 minutes after the predicted P-wave arrival time. Such a data set may be used, e.g., for magnitude computation or body-wave moment tensor inversion.

### 3.1.1 IRIS Wilber3

IRIS Wilber is a web application for requesting event-oriented data from the IRIS DMC. It allows users to easily find earthquakes from DMC's database and access waveform data based on a particular event. Waveform data are prepared using earthquake travel times and may be downloaded in a variety of data formats including SEED, SAC, and miniSEED. It is worth mentioning that WILBER3 interacts with information at the IRIS DMC totally through FDSN web services deployed at the IRIS DMC.

- URL: <http://www.iris.edu/wilber3> (see Figure 1)
- At *Load Event Data* select *Since 1990, M5.5+*
- At the right panel, *Date* field, enter **2011-03-11** as both start and end date
- You are now offered a list of events that occurred on the day 2011-03-11. Scroll to the bottom of the list and select the Tohoku main shock.
- Go to the *Select Stations* dialog (see Figure 2). A world map with available stations is now displayed, with the seismic travel times as isolines.
- Seismic Network). To select a different (virtual) network, click on the text field and select the network from the pop-up menu.
- On the bottom panel the selection can be fine-tuned on a per-station level if needed.
- Finally, hit *Request Data* to get a pop-up window in which you have to specify time windows and data format (see Figure 3). Most users will request either SEED or MiniSEED format, but options like SAC can be selected, too. In our example we select output as SEED volume. In this case, the request will be processed by the IRIS BREQ\_FAST, which means that once the request is ready (usually within a few minutes) the user is informed by email which SEED file to pick up at the IRIS FTP server.

On the panel to the right of the map, you can further limit the station selections according to distance and azimuth with respect to the event, but also channels (default is BH? - all broadband channels) and (virtual) networks (default is GSN - IRIS Global

**IRIS** INCORPORATED RESEARCH INSTITUTIONS FOR SEISMOLOGY

Resources/Search

Home / Wilber 3

## Wilber 3: Select Event

☒ Wilber Feedback/Questions

Looking for previously requested data? [View recent requests.](#)

Load Event Data: Since 1990, M5.5+

**Show Only**

Location: N, S, E, W

Date: 2011-03-11 - 2011-03-11

Magnitude: 5.5 - 10

147 events listed.

| Date (UTC)          | Region                           | Magnitude | Latitude | Longitude | Depth   | Contributor |
|---------------------|----------------------------------|-----------|----------|-----------|---------|-------------|
| 2011-03-11 06:00:10 | Near East Coast Of Honshu, Japan | mb 5.8    | 36.41°   | 141.69°   | 26 km   | ISC         |
| 2011-03-11 06:05:01 | Near East Coast Of Honshu, Japan | mb 5.8    | 36.41°   | 141.69°   | 26 km   | ISC         |
| 2011-03-11 06:04:00 | Near East Coast Of Honshu, Japan | mb 5.9    | 36.33°   | 141.85°   | 67 km   | ISC         |
| 2011-03-11 06:00:35 | Near East Coast Of Honshu, Japan | mb 6.2    | 38.13°   | 142.50°   | 14 km   | ISC         |
| 2011-03-11 05:59:35 | Near East Coast Of Honshu, Japan | mb 5.9    | 37.04°   | 141.74°   | 33 km   | ISC         |
| 2011-03-11 05:58:02 | Off East Coast Of Honshu, Japan  | mb 6.3    | 37.67°   | 142.04°   | 9.5 km  | ISC         |
| 2011-03-11 05:57:25 | Near East Coast Of Honshu, Japan | mb 5.7    | 39.21°   | 142.35°   | 32 km   | ISC         |
| 2011-03-11 05:55:45 | Off East Coast Of Honshu, Japan  | mb 6.4    | 37.30°   | 143.41°   | 35 km   | ISC         |
| 2011-03-11 05:54:42 | Near East Coast Of Honshu, Japan | mb 5.7    | 36.71°   | 140.58°   | 10 km   | ISC         |
| 2011-03-11 05:54:31 | Near East Coast Of Honshu, Japan | mb 6.3    | 37.51°   | 141.36°   | 35 km   | ISC         |
| 2011-03-11 05:51:20 | Off East Coast Of Honshu, Japan  | mb 6.8    | 37.31°   | 142.24°   | 33 km   | ISC         |
| 2011-03-11 05:46:23 | Near East Coast Of Honshu, Japan | MW 9.1    | 38.30°   | 142.50°   | 19.7 km | ISC         |
| 2011-03-11 00:14:48 | Southern East Pacific Rise       | MW 5.7    | -53.15°  | -117.75°  | 10 km   | ISC         |

**Figure 1** Selection of seismic events using the IRIS Wilber 3 interface

### 3.1.2 EIDA WebDC

The WebDC web interface is a portal to the seismic waveform data archives at GEOFON, including the European Integrated Data Archive (EIDA) data holdings. It allows to request

- seismological waveform data and
- station inventories (Dataless SEED, Inventory XML).

from different data centers across Europe participating in the EIDA initiative. In order to retrieve data for the Tohoku earthquake follow these steps:

- URL: <http://webdc.eu>
- We start our request session at the *Explore events* tab (see Figure 4)
- Select a catalog: GFZ is default, USGS and EMSC are other options. Alternatively, a user-supplied event list may be used and uploaded to the WebDC server. This is particularly useful to request data for events not covered by the above earthquake catalogs.
- As in the Wilber3 example above, we again select **2011-03-11** as start and end date. By limiting the magnitude to 7 and above, we restrict our search to the Tohoku main shock and two major aftershocks.
- Once the main shock is selected, we move on to the *Explore stations* tab (see Figure 5) and select the GEOFON network (GE) and streams by code *BH*, which means broadband streams. By hitting "Search" the station selection is displayed on the map as well as in a list at the bottom of the page, where stations can be (de)selected individually if desired.
- Finally, to submit the request, we move on to the *Submit request* tab, where we can choose between relative and absolute time windows (see Figure 6). Relative time windows are specified relative to a selectable wave type (P by default). The data formats that can be requested include SEED, MiniSEED as well as station metadata (dataless SEED). After specifying a valid email address the "submit" button is hit to submit the data request to WebDC.
- After submitting the request, the user will be notified about it being processed by a blinking "Download data" tab at the top of the page. By clicking on that tab, a list of the most recent data request of the user (identified by the email address) will be displayed, along with the processing status and -once the request is completed- one or several download links (see Figure 7). Note, however, that the data may have been split into several volumes as EIDA is an integrated data center with several nodes in Europe, each of which is responsible for its own data. This also means that the user downloads the data directly from the respective node.



INCORPORATED RESEARCH INSTITUTIONS FOR SEISMOLOGY

[Resources/Search](#)



[Home](#)
[About IRIS](#)
[Programs](#)
[Educational Resources](#)
[Data & Software](#)
[Multimedia](#)
[IRIS Community](#)

[Home / Wilber 3 / 2011-03-11 MW9.1 Near East Coast Of Honshu, Japan](#)

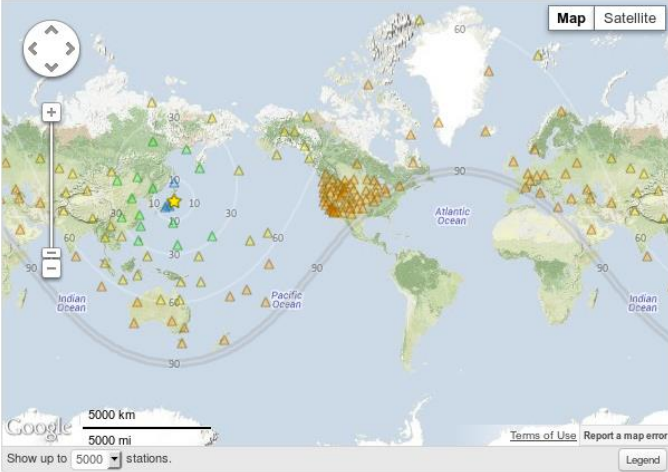
## Wilber 3: Select Stations

[Wilber Feedback/Questions](#)

### 2011-03-11 MW9.1 Near East Coast Of Honshu, Japan

| Latitude   | Longitude  | Date                    | Depth   | Magnitude | Description                      | Related Pages                      |
|------------|------------|-------------------------|---------|-----------|----------------------------------|------------------------------------|
| 38.2963° N | 142.498° E | 2011-03-11 05:46:23 UTC | 19.7 km | MW9.1     | Near East Coast Of Honshu, Japan | <a href="#">IRIS Data Products</a> |

The map below shows stations operational during this event, filtered by the criteria in the form to the right.



Show up to 5000 stations.

#### Request Only

**Networks**

**Channels**

Set default networks/channels

**Distance Range**

-

**Azimuth Range**

☐ Invert

-

#### Actions

[Show Record Section](#)
[Request Data](#)

Use the checkboxes below to add/remove individual stations from your request.

Selected 145 out of 145 stations.

Select

All
None
One station every

|                                     | Station | Network | Latitude | Longitude | Distance | Azimuth  | Elevation | Name                                       |
|-------------------------------------|---------|---------|----------|-----------|----------|----------|-----------|--------------------------------------------|
| <input checked="" type="checkbox"/> | TSK     | PS      | 36.21°   | 140.11°   | 2.82°    | -136.92° | 350 m     | Tsukuba, Japan                             |
| <input checked="" type="checkbox"/> | MAJO    | IU      | 36.55°   | 138.20°   | 3.83°    | -115.86° | 405 m     | Matsushiro, Japan                          |
| <input checked="" type="checkbox"/> | INU     | G       | 35.35°   | 137.03°   | 5.28°    | -122.28° | 132 m     | Inuyama, Japan                             |
| <input checked="" type="checkbox"/> | YSS     | IU      | 46.96°   | 142.76°   | 8.66°    | 1.19°    | 150 m     | Yuzhno Sakhalinsk, Russia                  |
| <input checked="" type="checkbox"/> | OGS     | PS      | 27.06°   | 142.20°   | 11.24°   | -178.65° | 20 m      | Chichijima, Bonin Islands, Japan           |
| <input checked="" type="checkbox"/> | MDJ     | IC      | 44.62°   | 129.59°   | 11.53°   | -52.67°  | 270 m     | Mudanjiang, Heilongjiang Province, China   |
| <input checked="" type="checkbox"/> | PET     | IU      | 53.02°   | 158.65°   | 18.47°   | 31.89°   | 110 m     | Petropavlovsk, Russia                      |
| <input checked="" type="checkbox"/> | SSE     | IC      | 31.09°   | 121.19°   | 18.88°   | -105.98° | 40 m      | Shanghai, China                            |
| <input checked="" type="checkbox"/> | HIA     | IC      | 49.27°   | 119.74°   | 19.64°   | -48.69°  | 620 m     | Hallar, Neimenggu Autonomous Region, China |
| <input checked="" type="checkbox"/> | BJT     | IC      | 40.02°   | 116.17°   | 20.41°   | -76.86°  | 197 m     | Baijiatuan, Beijing, China                 |
| <input checked="" type="checkbox"/> | MA2     | IU      | 59.58°   | 150.77°   | 21.92°   | 11.25°   | 339 m     | Magadan, Russia                            |
| <input checked="" type="checkbox"/> | TATO    | IU      | 24.97°   | 121.50°   | 22.21°   | -120.73° | 160 m     | Taipei, Taiwan                             |
| <input checked="" type="checkbox"/> | GUW     | IU      | 13.58°   | 144.87°   | 24.80°   | -174.50° | 170 m     | Guam, Mariana Islands                      |

**Figure 2** Station selection using the IRIS Wilber 3 interface



**Build Data Request**

Requesting **606** channels from **145** stations. Estimated size: **43 megabytes**  
*\* Rough estimate only; variability in compression and data gaps may result in download size up to 2x this value.*

**Time Range**

Starting: 5 minutes  
before: P arrival

Ending: 10 minutes  
after: P arrival

**Data Output**

Output Format: SEED  
*SEED output is requested through the **BREQ\_FAST** system. These requests cannot be tracked using Wilber.*

Institution: GFZ Potsdam

Address:

Phone:

Data Quality: E - Everything

**Request Information**

Your Name: Joachim Saul

Request Label: 2011-03-11 MW9.1 Near East Coas

*Your name and the request label are used to uniquely identify your request, so you cannot reuse a label you used before.*

Email Notification: saul@gfz-potsdam.de

☐ Save these settings as defaults **Submit** **Cancel**

**Figure 3** Formulation of the actual data request using the IRIS Wilber 3 interface

**GFZ** **GEOFON** **EIDA** **DEUTSCHES GEOFORSCHUNGSZENTRUM**

Explore events Explore stations Submit request Download data View console

[OLD webdc.eu portal](#) (Preliminary Doc.: [PDF](#) [Help](#))

**Events Controls**

**Event Information**

Catalog Services: User Supplied

Catalog Service: GFZ

Quick look-up of last 10 events at GFZ: With M<sub>z</sub> 6.0

Date Interval (yyyy-mm-dd): 2011-03-11 to 2011-03-11

Minimum Magnitude: 7

Depth from 0 to 999 km

Coordinates: (Use -ve for S/W; +ve for N/E)

N 90

W -180 180 E

-90

S

**Event and Station Map**

Map data ©2013 MapLink - Terms of Use

Use left SHIFT + drag mouse to select regions. [Legend](#) [Help](#)

**Event and Station List**

Request: Freeze Delete Stations Delete Events

**Events (3 events)**

| Origin Time                                             | Mag. | Lat.  | Long.  | Depth | Region                           |
|---------------------------------------------------------|------|-------|--------|-------|----------------------------------|
| <input type="checkbox"/> 2011-03-11T06:25:52            | 7.6  | 38.37 | 144.66 | 10.0  | Off East Coast of Honshu, Japan  |
| <input type="checkbox"/> 2011-03-11T06:15:38            | 7.8  | 36.32 | 140.71 | 30.0  | Near East Coast of Honshu, Japan |
| <input checked="" type="checkbox"/> 2011-03-11T05:46:23 | 8.9  | 38.23 | 142.53 | 15.0  | Near East Coast of Honshu, Japan |

**Stations (-)**

No Stations loaded

WebDC3 Interface © (2013) Helmholtz-Zentrum Potsdam - Deutsches GeoForschungsZentrum GFZ

**Figure 4** Explore seismic event catalogs using the WebDC interface




Access to GEOFON and EIDA Data Archives



HELMHOLTZ-ZENTRUM POTSDAM  
DEUTSCHES  
GEOFORSCHUNGSZENTRUM

[Explore events](#)
[Explore stations](#)
[Submit request](#)
[Download data](#)
[View console](#)

[OLD webdc.eu portal](#) [Preliminary Doc.: [PDF](#)] [Help](#)

Stations Controls

Networks

Year from 1980 to 2013:

Network Type:
All permanent nets

Network Code:
GE (1993) - GEOFON Prog

\* = temporary network; + = restricted access

Stations

by Code by Region by Events

Filter stations by region:
N
90
W -180 180 E
-90
S
Clear

Streams

by Code by Sampling

Choose the desired set of channels:  
Use SHIFT and CTRL to extend the set.

BH  
LH  
VH  
HH

Reset Append

Event and Station Map



Map data ©2013 MapLink - Terms of Use  
Use left SHIFT + drag mouse to select regions.

[Legend](#) [Help](#)

Event and Station List

Request:
Freeze Delete Stations Delete Events

Events (3 events)

| Origin Time                                             | Mag. | Lat.  | Long.  | Depth | Region                           |
|---------------------------------------------------------|------|-------|--------|-------|----------------------------------|
| <input type="checkbox"/> 2011-03-11T06:25:52            | 7.6  | 38.37 | 144.66 | 10.0  | Off East Coast of Honshu, Japan  |
| <input type="checkbox"/> 2011-03-11T06:15:38            | 7.8  | 36.32 | 140.71 | 30.0  | Near East Coast of Honshu, Japan |
| <input checked="" type="checkbox"/> 2011-03-11T05:46:23 | 8.9  | 38.23 | 142.53 | 15.0  | Near East Coast of Honshu, Japan |

Stations (110 stations)

| Network                             | Station | Lat. | Long. | O/R    | Streams        |
|-------------------------------------|---------|------|-------|--------|----------------|
| <input checked="" type="checkbox"/> | GE      | APE  | 37.07 | 25.53  | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | APEZ | 34.98 | 24.89  | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BGIO | 31.72 | 35.09  | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BKB  | -1.11 | 116.90 | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BKNI | 0.33  | 101.04 | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BNDI | -4.52 | 129.90 | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BOAB | 12.45 | -85.67 | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CART | 37.59 | -1.00  | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CEU  | 35.90 | -5.37  | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CISI | -7.56 | 107.82 | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CSS  | 34.96 | 33.33  | .BHE,.BHN,.BHZ |

WebDC3 Interface © (2013) Helmholtz-Zentrum Potsdam - Deutsches GeoForschungsZentrum GFZ

Figure 5 Explore stations using the WebDC interface




Access to GEOFON and EIDA Data Archives



HELMHOLTZ-ZENTRUM POTSDAM  
DEUTSCHES  
GEOFORSCHUNGSZENTRUM

[Explore events](#)
[Explore stations](#)
[Submit request](#)
[Download data](#)
[View console](#)

[OLD webdc.eu portal](#) [Preliminary Doc.: [PDF](#)] [Help](#)

**Make Request**

**Time Window selection:**

Relative Mode Absolute Mode

Use time windows relative to events, by phase and onset time.

Start (minutes before)  
 -

End (minutes after)  
 +

**Request Information:**

Request type:

☐ Waveform (Mini-SEED)  
☒ Waveform (Full SEED)  
☐ Metadata (Dataless SEED)  
☐ Metadata (Inventory XML)

Use compression?  
☐ Yes ☒ No

Use response dictionary?  
☒ Yes ☐ No

Your e-mail address:

☐ Remember me?

**Event and Station Map**



Map data ©2013 MapLink - [Terms of Use](#)

Use left SHIFT + drag mouse to select regions. [Legend](#) [Help](#)

**Event and Station List**

**Request:**

**Events (3 events)**

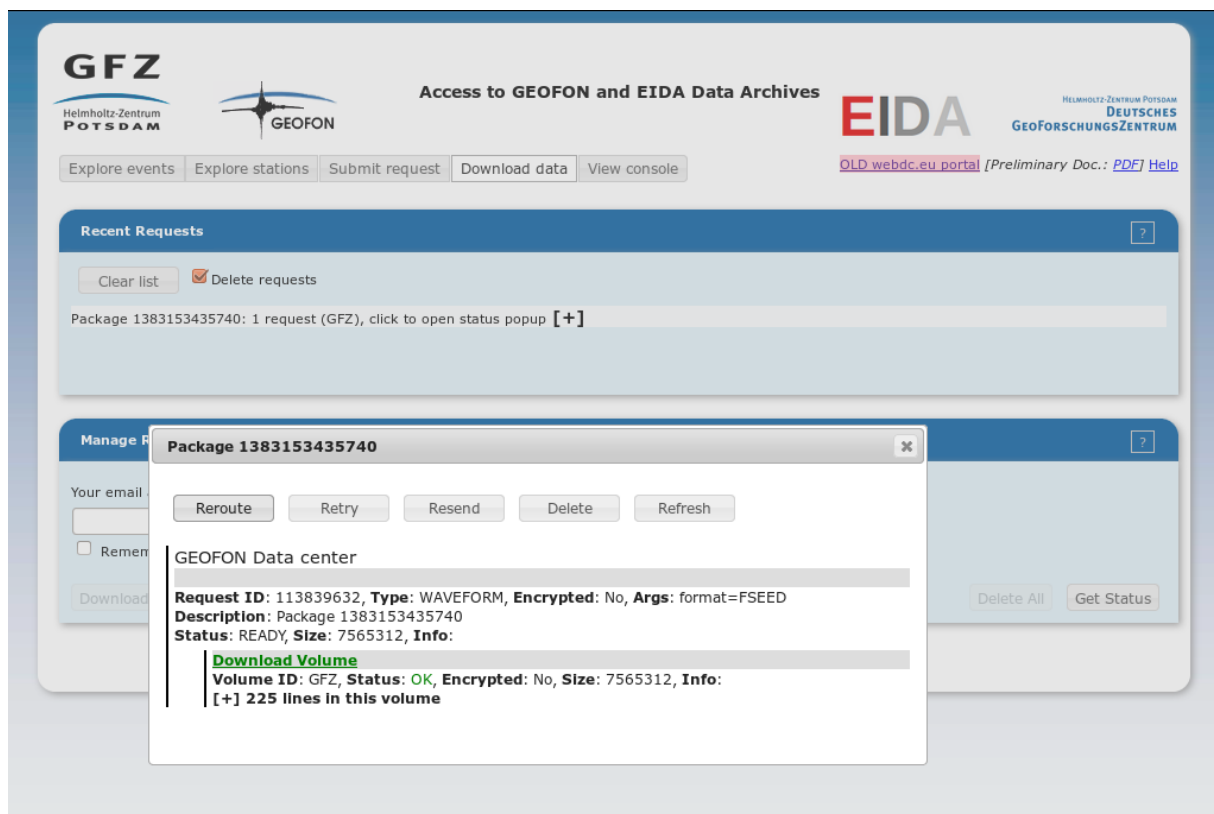
|                                     | Origin Time         | Mag. | Lat.  | Long.  | Depth | Region                           |
|-------------------------------------|---------------------|------|-------|--------|-------|----------------------------------|
| <input type="checkbox"/>            | 2011-03-11T06:25:52 | 7.6  | 38.37 | 144.66 | 10.0  | Off East Coast of Honshu, Japan  |
| <input type="checkbox"/>            | 2011-03-11T06:15:38 | 7.8  | 36.32 | 140.71 | 30.0  | Near East Coast of Honshu, Japan |
| <input checked="" type="checkbox"/> | 2011-03-11T05:46:23 | 8.9  | 38.23 | 142.53 | 15.0  | Near East Coast of Honshu, Japan |

**Stations (110 stations)**

|                                     | Network | Station | Lat.  | Long.  | O/R | Streams        |
|-------------------------------------|---------|---------|-------|--------|-----|----------------|
| <input checked="" type="checkbox"/> | GE      | APE     | 37.07 | 25.53  |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | APEZ    | 34.98 | 24.89  |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BGIO    | 31.72 | 35.09  |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BKB     | -1.11 | 116.90 |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BKNI    | 0.33  | 101.04 |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BNDI    | -4.52 | 129.90 |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | BOAB    | 12.45 | -85.67 |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CART    | 37.59 | -1.00  |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CEU     | 35.90 | -5.37  |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CISI    | -7.56 | 107.82 |     | .BHE,.BHN,.BHZ |
| <input checked="" type="checkbox"/> | GE      | CSS     | 34.96 | 33.33  |     | .BHE,.BHN,.BHZ |

WebDC3 Interface © (2013) Helmholtz-Zentrum Potsdam - Deutsches GeoForschungsZentrum GFZ

Figure 6 Data request submission



**Figure 7** Finally, download the data as a SEED volume

### 3.1.3 Remarks

Both presented interactive request methods are similar in that they greatly facilitate the whole process of event selection, station discovery and retrieval. They are the methods of choice for the retrieval of a limited data set for one or few earthquakes, especially if it is unclear which stations shall be used. If large data sets need to be compiled, the interactive sessions may be too time consuming. Fully automated processing is not possible at all, at least not until the data have been downloaded. This is where the use of non-interactive methods is required, which we will look at next.

## 3.2 Non-interactive methods

Any user who wants to retrieve large quantities of data, like many stations for many events, will quickly realize that while interactive methods provide an excellent starting point for exploring data archives, often at the beginning of larger projects, they are normally too inefficient to use if hundreds or thousands of data sets need to be retrieved. This is where automated requests using non-interactive request methods come into play. Automated requests are also needed in automated seismic monitoring environments.

A variety of automatic request tools exist, some of which have been in existence for many years and are widely known and used, others are newer and reflect technical developments of the recent years.

Here we will provide an overview and a discussion about some of the more important request tools.

### 3.2.1 Email based methods

Being one of the earliest Internet communication protocols, email was quickly adopted as a means for requesting and transmitting seismic data. Several different techniques have been in use over the years, some of which are now obsolete but mentioned here for the sake of completeness:

#### **AutoDRM**

developed at the Swiss Seismological Service (SED) at ETH Zürich. AutoDRM was a purely email based request service in which both the request as well as the compiled data volume were transmitted as text messages through email in GSE format. AutoDRM pioneered seismological data access but is now obsolete due to the availability of much more efficient techniques. However, some data centers especially in Europe, continue to provide this service.

#### **BREQ\_FAST**

is another email based tool and was developed by IRIS. Data requests are sent in an email message to a data request manager, where they are processed. Upon completion, the data are delivered via a medium as specified in the request. Nowadays the FTP Internet protocol is the most common medium, but for large data volumes magnetic tapes might occasionally still be the preferred media type.

A BREQ\_FAST request message consists of a header, in which the user provides some personal information like name, postal and email address. The user can specify a preferred media type for the data delivery and a request label, under which the request can be tracked.

Following the header, the request message contains an arbitrary number of lines, each of which consists of a combination of station/network/channel codes and associated time windows to be extracted. Each of the textual fields may contain wild cards, e.g., for matching more than one channel code to get all available components of a sensor.

For more information about BREQ\_FAST, see  
[http://www.iris.edu/dms/nodes/dmc/manuals/breq\\_fast/](http://www.iris.edu/dms/nodes/dmc/manuals/breq_fast/)

A detailed example will be shown below.

#### **NetDC - Networked Data Center Protocol**

Like BREQ\_FAST, requests are email based, whereas data are delivered to the user via either the FTP Internet protocol or magnetic tapes (Exabyte). NetDC is in principle superior to BREQ\_FAST because it allows the user to submit a data request without knowing where the data are stored physically. It introduced the concept of "networked data centers", where requests are automatically routed to the data center responsible for the data.



When making the data request, the user can choose between either receiving data from individual data centers or receiving the data as a single merged product from the data center originally contacted. Either way a user interested in data will have to contact only one site when making a request, which is certainly a big advantage.

Due to the availability of more modern request techniques, NetDC is now considered obsolete. While it is still in use at a few data centers (e.g. GEOSCOPE) it has already been shut down at the IRIS DMC in 2013.

Out of these three most popular email based request techniques, nowadays only BREQ\_FAST continues to be of importance and is supported by GEOFON, IRIS and EIDA. We will therefore briefly explain the basics of a data request using BREQ\_FAST.

### Example BREQ\_FAST waveform data request

In the following example, we want to request a recording of the 2011 Tohoku earthquake recorded at the GEOFON station Wanagama, Indonesia, which has the station code UGM. The GEOFON network code is GE. We want to retrieve all three components in a time window starting 10 minutes before and ending one hour after the earthquake origin time, which is 2011-03-11 05:46:23 UTC.

```
.NAME Joe Seismologist
.INST GFZ Potsdam
.MAIL Telegrafenberg, Potsdam, Germany
.EMAIL joe.seismologist@seismology.org
.MEDIA: Electronic (FTP)
.LABEL Tohoku-UGM
.END
```

```
UGM GE 2011 03 11 05 36 23.0 2011 03 11 06 46 23.0 1 BH?
```

As data for this particular station are archived both at GEOFON and IRIS data centers, the request may be submitted to either the GEOFON/WebDC (email: [breq\\_fast@webdc.eu](mailto:breq_fast@webdc.eu)), the ORFEUS DC (email: [breq\\_fast@knmi.nl](mailto:breq_fast@knmi.nl)) or the IRIS DMC (email: [breq\\_fast@iris.washington.edu](mailto:breq_fast@iris.washington.edu)). In general, however, when using BREQ\_FAST for requesting data, it must be known beforehand which data center provides what data, because the BREQ\_FAST service itself provides no functionality for station discovery. This is often not a problem, especially if the same set of stations (or “virtual network”) is always used for a certain analysis and if the number of stations is small enough to be easily manageable using e.g. home grown station inventories like station lists. When BREQ\_FAST was developed more than 20 years ago, this was possible simply because the digital seismic networks consisted of much fewer stations than today.

But things have changed. What if there are many stations in our region of interest and we want to use just as many of them as possible? What if the station/network configuration changes nearly on a daily basis like in the case of the EarthScope network? It then becomes increasingly tedious and time consuming to keep a local station inventory up to date. Fortunately, there are now services that make this task easier. We will explore these in more detail below.

### 3.2.2 ArcLink

ArcLink is a data request protocol developed within EIDA especially to facilitate access to distributed, de-centralized data archives. High-quality, digital seismic waveform data are nowadays openly available from many institutions and data centers world-wide. Data are therefore physically stored in different places. This requires tools that facilitate the discovery and access to such distributed archives. In particular, data requests need to be routed from a client to often several server nodes. ArcLink can therefore be considered an enhancement of the now discontinued NetDC.

ArcLink uses its own protocol language between an ArcLink server and an ArcLink client program. ArcLink is also the architecture behind the EIDA WebDC data portal. It consists of dedicated server and client software. For data retrieval, a client software is needed. There are several options, with the most popular being

#### **arclink\_fetch**

is a stand-alone download client, which is distributed as part of the SeisComp software package (Hanka et al., 2010). It is invoked either from the command line prompt or from within shell scripts:

```
$ echo '2011, 3, 11, 5, 36, 23 2011, 3, 11, 6, 46, 23 GE UGM BHZ *' | \
arclink_fetch -u "joe@seismology.org" -o GE.UGM..BHZ.mseed
```

The resulting MiniSEED file GE.UGM..BHZ.mseed can then be saved and viewed by using any seismogram viewer or be further processed.

#### **obspy.arclink**

is a library of client functions that are invoked from programs written in the Python programming language. Example:

```
$ python
Python 2.7.3 (default, Feb 27 2014, 19:58:35)
[GCC 4.6.3] on linux2
Type "help", "copyright", "credits" or "license" for more information.
>>> from obspy import UTCDateTime
>>> from obspy.arclink.client import Client
>>> client = Client(user='joe@seismology.org')
>>> t = UTCDateTime("2011-03-11 05:46:23")
>>> st = client.getWaveform("GE", "UGM", "", "BHZ", t-600, t+3600)
>>> st.plot()
```

Instead of plotting the data, it is of course also possible to write them to a file:

```
>>> client.saveWaveform("GE.UGM..BHZ.mseed",
... "GE", "UGM", "", "BHZ",
... t-600, t+3600, format='MSEED')
```

This will save the raw MiniSEED data to a file GE.UGM..BHZ.mseed, which can then be processed. Also other formats commonly used by data analysis packages are supported (e.g., SAC, SEISAN, GSE etc.)

### SeisComP3

is software for seismological data archival, exchange and processing. It has built-in ArcLink client support. In fact, ArcLink is the default archive access protocol of SeisComP (Hanka et al., 2010).

NOTE: ArcLink is a popular protocol in Europe within the EIDA initiative and within the SeisComP ecosystem. It is not supported by major data centers like IRIS DMC, however, and has therefore not become a world-wide standard. A standard archive access mechanism is provided by web services.

### 3.2.3 Web Services

Web services in general are a method of communications over the World Wide Web, hence the name. The fact that web services use the HTTP Internet protocol allows access to resources offered by web services usually even in tightly security controlled environments (e.g., behind network firewalls, etc.), because the Internet ports used by the web services are the same as for interactively browsing the WWW. Many Internet technologies nowadays are based on web services and as a natural consequence web services also have been adopted in seismology as a means for exchange of waveform and parametric data.

Web services for seismic waveform data retrieval have been developed over the last approximately 10 years in several institutions in Europe and the United States, resulting in initially different and incompatible solutions. Fortunately in 2013, an agreement on a common web service standard could be achieved by the members of the FDSN. The set of web services defined in this standard are referred to as “FDSN web services” (FDSN, 2013):

**fdsnws-station** - For access to station metadata in FDSN Station XML format

**fdsnws-dataselect** - For access to time series data in MiniSEED format

**fdsnws-event** - For access to event parameters in QuakeML format

The access to the FDSN web services is very simple and there exists a variety of techniques to choose from. In fact, the HTTP protocol allows the use of any web browser to access data from an FDSN web service, which is very convenient especially for the novice user to explore the web service features. Non-interactive HTTP client programs like “wget” and “curl” may easily be used as FDSN web service clients. In addition, specialized client software exists to make the data retrieval simple and transparent.

The data request can be specified in two different ways. One is simply via a URL, which essentially must consist of the stream to load (possibly containing wild cards matching several streams at once) and the time window. The stream is specified by the network, station, location and channel codes. See the SEED manual for more details about these fields.

For example, if we want to request the same data as in the above example BREQ\_FAST waveform data request (70 minutes of data for the 2011 Tohoku earthquake recorded at GEOFON station UGM), the URL to fetch these data would be <http://geofon.gfz-potsdam.de/fdsnws/dataselect/1/query?sta=UGM&net=GE&start=2011-03-11T05:36:23.0&end=2011-03-11T06:46:23.0>



[11T06:46:23.0&cha=BH?](http://11T06:46:23.0&cha=BH?) This URL may simply be entered in any web browser and the resulting data file may be immediately viewed or processed.

Using the same URL, the data can also be fetched with command line utilities like “wget” or “curl”, which allows partial automation of data requests, e.g. using shell scripts. Since the URLs are very easy to generate, this is a very convenient method for requesting and immediately retrieving seismic data.

When many different stations or time windows shall be retrieved at one time, these cannot all be specified on a single URL. Here an alternative flavor of the web service comes into play, which uses the HTTP-POST technique to upload a request file to the web service. This request file contains lines with individual combinations of streams and time windows, thus similar to the BREQ\_FAST email requests. The text file containing the request is uploaded to the web service and the data are retrieved immediately.

NOTE that the data retrieved using FDSN web services are formatted as MiniSEED, which means that unlike full SEED volumes, they lack metadata such as instrument responses which, if needed, have to be requested separately using the fdsnws-station web service.

#### **wget**

is a general-purpose HTTP client that can also be used to request seismic data using the FDSN web services. Invocation simply as

```
$ wget -O data.mseed "http://..."
```

#### **curl**

is an alternative to “wget”. Invocation equally simple as

```
$ curl -o data.mseed "http://..."
```

#### **obspy.fdsn**

is a client library that may be used from programs written in the Python programming language. Example:

```
$ python
Python 2.7.3 (default, Feb 27 2014, 19:58:35)
[GCC 4.6.3] on linux2
Type "help", "copyright", "credits" or "license" for more information.
>>> from obspy import UTCDateTime
>>> from obspy.fdsn import Client
>>> client = Client("GFZ")
>>> t = UTCDateTime("2011-03-11 05:46:23")
>>> st = client.get_waveforms("GE", "UGM", "", "BHZ",
... t-600, t+3600)
>>> st.plot()
```

Note the great similarity with the obspy.arlink example provided above. Instead of plotting the data, it is of course also possible to write them to file:

```
>>> client.get_waveforms("GE", "UGM", "", "BHZ", t-600, t+3600,
... filename="GE.UGM..BHZ.mseed")
```

This will save the raw MiniSEED data to a file GE.UGM..BHZ.mseed, which can then be processed. Note that unlike in the above obspy.arclink example, the data format is not specified because here the only available format is MiniSEED, as it is not possible to download full SEED using the FDSN web services. Note that the above call only retrieves raw waveform data. In order to retrieve metadata, the above ObsPy client provides a “get\_stations” method. Please refer to the ObsPy documentation for more details (e.g., Beyreuther et al. 2010, and the [ObsPy home page](#)).

### IRIS Fetch scripts

can be used on the command line or from shell scripts. Invocation example:

```
$ FetchData -N GE -S UGM -C BHZ -s 2011-03-11T05:36:23 ¥  
-e 2011-03-11T06:46:23 -o UGM.mseed -m UGM.metadata
```

In the above call both raw waveform data as well as basic metadata like sensor coordinates and gain are downloaded.

Data for GEOFON station UGM happens to be available from both GEOFON and IRIS. By default, the FetchData script connects to the IRIS fdsnws server. In order to connect to the GEOFON fdsnws server, this default can be overridden using so-called environment variables. The above call then becomes:

```
$ export SERVICEBASE=http://geofon.gfz-potsdam.de  
$ FetchData -N GE -S UGM -C BHZ -s 2011-03-11T05:36:23 ¥  
-e 2011-03-11T06:46:23 -o UGM.mseed -m UGM.metadata
```

### SeisComP3

comes with a fdsnws server implementation included, which can be configured to access the data through ArcLink or directly from a mounted file system. Additionally, SeisComP has built-in support for fdsnws-datasetselect as client. All SeisComP programs, which use waveform data as fixed time windows can be configured to retrieve waveforms through fdsnws-datasetselect.

### Other web service clients

for MatLab and Java are available for download under <http://service.iris.edu/clients>

As the number of data centers supporting FDSN web services increases, it should be possible to develop clients that can discover data holdings across the entire system of centers. For instance, IRIS is currently developing a web service that will look at the holdings of all data centers running FDSN web services, and return information to a client application that will enable parallel recovery of waveforms from multiple data centers. Again the data will be sent from multiple data centers directly to the client application allowing each data center to understand its customer base. A similar “routing” facility is also being developed in Europe within EIDA. Any such solution will require special “smart client” software to make this data center integration transparent to the user.

### 3.2.4 Data discovery

In the previous section about web services, we have assumed that the user knows which data streams to request. In general, however, this may not always be the case and we may want to make a selection of stations/streams based on ad-hoc criteria such as a geographical region or epicentral distance range around an earthquake. This “discovery” process becomes actually quite simple using the “fdsnws-station” web service. If for instance we want to request data from IRIS for all stations within 50 degrees around the 2011 Tohoku earthquake. Given the epicenter coordinates and the time window it is very simple to request a list of stations within that distance range:

[http://service.iris.edu/fdsnws/station/1/query?level=channel&format=text&channel=BHZ&network=I\\*&include\\_restricted=false&start=2011-03-11&end=2011-03-11&longitude=142.5&latitude=38.2&maxradius=50](http://service.iris.edu/fdsnws/station/1/query?level=channel&format=text&channel=BHZ&network=I*&include_restricted=false&start=2011-03-11&end=2011-03-11&longitude=142.5&latitude=38.2&maxradius=50)

This is a long URL, but the format is actually quite simple and almost self-explanatory. `level=channel&format=text&channel=BHZ` requests a text list containing BHZ channels. Stations without BHZ channels will therefore not be listed, allowing a crude pre-selection of broadband sensors. Here we want to list only stations from the IRIS networks (II,IU,IC).

The full specification for the FDSN web services can be found under <http://www.fdsn.org/webservices/>.

### 3.2.5 Outlook

Nowadays automated data retrieval is possible using email based techniques as BREQ\_FAST, ArcLink and web services. All of these techniques have their specific advantages and will continue to be available for years to come. BREQ\_FAST and ArcLink allow data retrieval as full SEED volumes, which the web services don't. Instead, the web services promote a stronger separation of waveforms (MiniSEED format) and metadata (FDSN StationXML). This may appear like a disadvantage and for many users used to working with full SEED it currently is. Work flows need to be adopted, which is not always trivial. However, the data separation also has advantages. Erroneous metadata may more easily be corrected if the metadata are loaded “on the fly” during processing. Use of XML as metadata format will allow extensions to the metadata schema more easily.

It is likely that in the future the web services become by far the most important request mechanism, as they rely not only on standard data formats but are also based on standardized protocols. The web services are therefore being adopted and implemented worldwide at a growing number of institutions. What the web services are currently lacking is a possibility for routing requests as in ArcLink. It is therefore still necessary to know where the data are physically located. Also it is not yet possible to request data depending on quality metrics. But at the time of this writing, there are several initiatives working in that direction and working solutions will likely become available within the year 2015.

## 4 Conclusions

We have presented some of the main methods for requesting seismic data. The interactive web portals provide an excellent way to conveniently and quickly retrieve the data for a

limited data set in a nearly self-explanatory way. If you have done it once you know how it works. Very little learning is required.

The non-interactive methods - web services or `arclink_fetch` but also old-fashioned email - are much more powerful whenever the requests can be automated, e.g. if very large and/or numerous data sets have to be assembled. Embedding these mechanisms into a work flow may require a little more work at the beginning. Client software like ObsPy helps to minimize the work, which will anyway soon pay off due to the greatly reduced work to retrieve the data.

## Acknowledgment

The very careful and constructive reviews by W. Hanka, T. Ahern and J. Wassermann are acknowledged with thanks, as well as the reviews and discussions at several stages of the developing draft with the Manual editor, who also cared for the final formatting.

## References

Beyreuther, M., Barsch, R., Krischer, L., Megies, T., Behr Y. and Wassermann, J. (2010). ObsPy: A Python Toolbox for Seismology SRL, 81(3), 530-533 DOI: 10.1785/gssrl.81.3.530

Hanka, W., Saul, J., Weber, B., Becker, J., Harjadi, P., Fauzi, GITEWS Seismology Group (2010). Real-time earthquake monitoring for tsunami warning in the Indian Ocean and beyond. - Natural Hazards and Earth System Sciences (NHESS), 10, 12, p. 2611-2622. DOI: 10.5194/nhess-10-2611-2010

FDSN (2012). SEED reference manual, Standard for the Exchange of Earthquake Data, SEED format version 2.4, August 2012, [http://www.fdsn.org/seed\\_manual/SEEDManual\\_V2.4.pdf](http://www.fdsn.org/seed_manual/SEEDManual_V2.4.pdf)

FDSN (2013). Web Services Version 1.0 Specification, available as online document under <http://www.fdsn.org/webservices/FDSN-WS-Specifications-1.0.pdf>

Links to software packages:

ObsPy home page: <http://www.obspy.org>

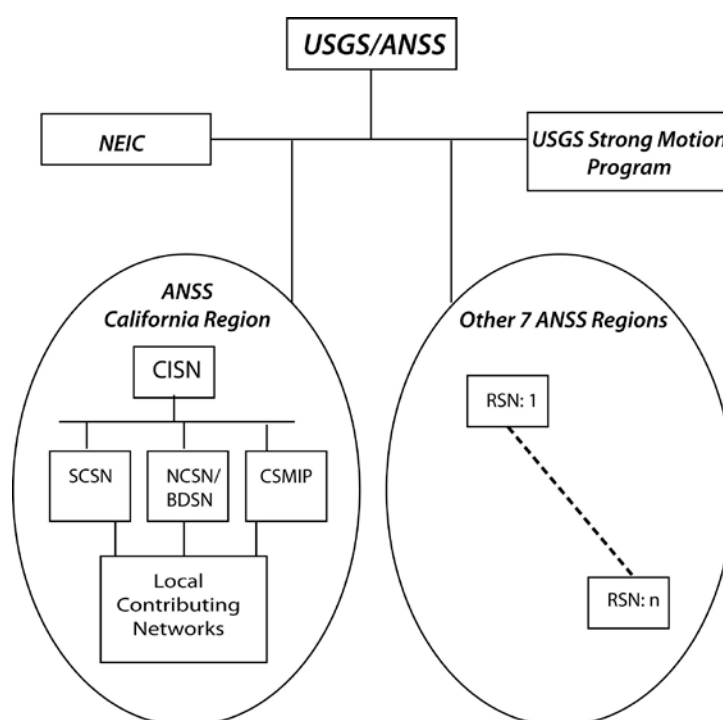
SeisComP home page: <http://www.seiscomp3.org>

IRIS rdseed manual: <http://www.iris.edu/ds/nodes/dmc/manuals/rdseed>

| Topic   | California Integrated Seismic Network (CISN)                                                                                                                                                                             |
|---------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Egill Hauksson</b> , Seismological Laboratory MC 252-21, California Institute of Technology, 1200 East Calif. Blvd, Pasadena, CA 91125, USA<br>E-mail: <a href="mailto:hauksson@caltech.edu">hauksson@caltech.edu</a> |
| Version | June 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_8.4                                                                                                                                                                               |

## 1 How earthquake monitoring is organized in the United States

Earthquake monitoring in the US is organized under the Advanced National Seismic System (ANSS), a program operated by the US Geological Survey. The mission of the ANSS is to provide timely and accurate earthquake information products, including rapid estimates of their effects on the built environment across the Nation. The United States is divided into eight ANSS regions to facilitate monitoring on a regional scale and to foster partnerships with local universities and other seismic network operators. The California Integrated Seismic Network (CISN) is the umbrella organization for earthquake monitoring in the California region of ANSS (Figure 1).



**Figure 1** Schematic organizational diagram illustrating how the ANSS, and as an example the CISN and other regional networks (RSNs, numbered 1 through n) form a system. **Bottom left:** The main components of the CISN; **Bottom right:** Summary diagram to indicate the other seven regions of ANSS; BDSN – Berkeley Digital Seismic Network; CISN – California Integrated Seismic Network; CSMIP – California Strong Motion Instrumentation Program; NCSN – Northern California Seismic Network; NEIC – USGS National Earthquake Information Center; and SCSN – Southern California Seismic Network.

## 2 California Integrated Seismic Network (CISN)

There are more than 15 seismic network operators in California that share data real-time as part of CISN operations. The CISN consists of three management centers. First, the California Geological Survey (CGS) operates the California Strong Motion Program, and provides near real-time strong motion data and associated products from all CISN partners to earthquake engineers. Second, the Northern California Seismic Network (NCSN) in Menlo Park and the Berkeley Digital Seismic Network (BDSN) at UC Berkeley provide real-time reporting for northern California. Third, the Caltech/USGS Southern California Seismic Network (SCSN) provides real-time reporting for southern California. The CISN partners also contribute data to the USGS/NEIC, develop standards, and carry out joint development and implementation of new technologies.

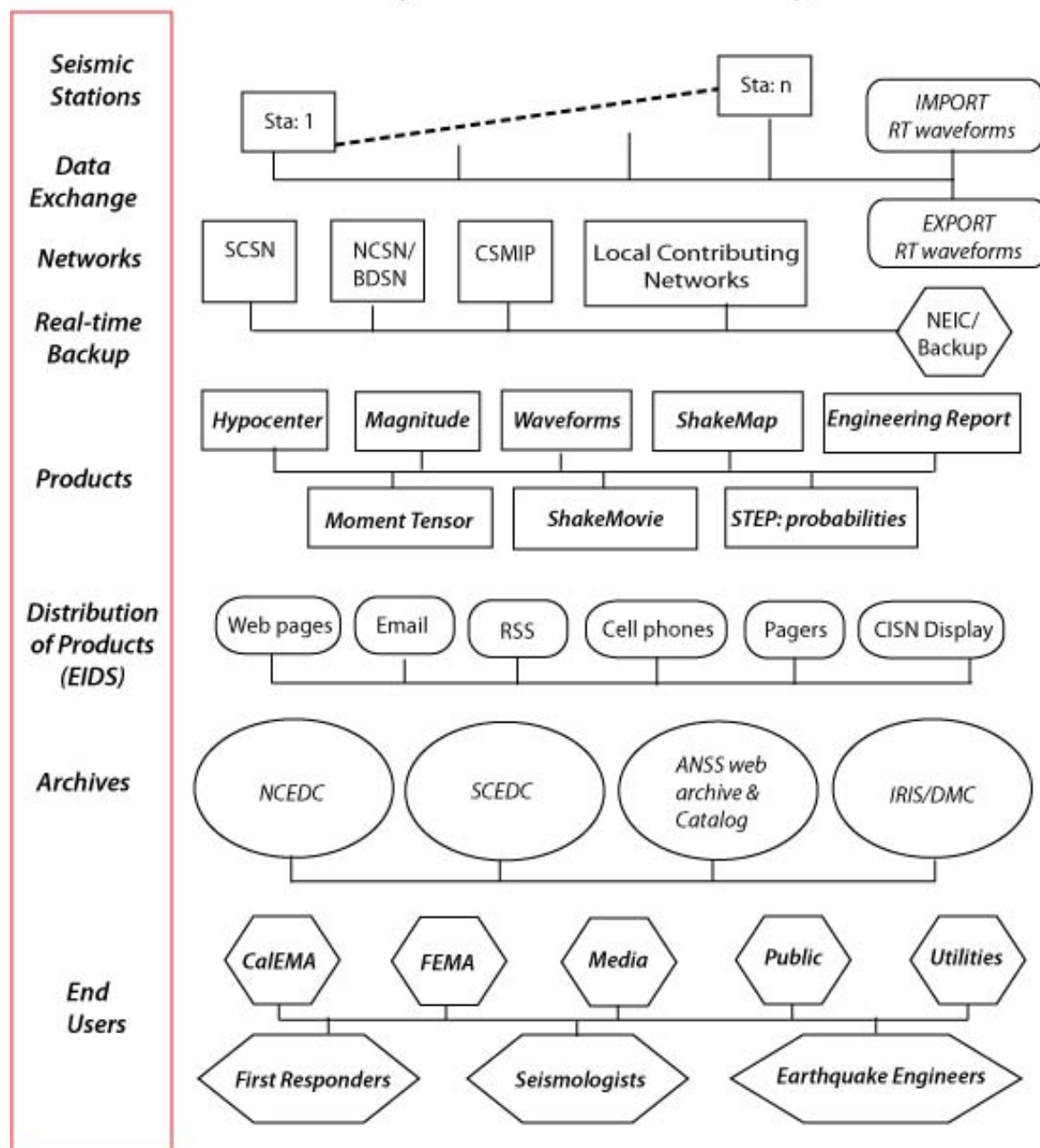
Extensive real-time data sharing takes place within the CISN. Subsets of real-time waveforms are shared automatically for processing by partner networks between central processing sites. In addition, about 12 broadband and strong motion stations operated by the SCSN transmit data real-time to both the SCSN in Pasadena and the BDSN in northern California, and comparable 12 BDSN stations are also recorded in Pasadena for backup purposes. Also, several strong motion stations provide simultaneous real-time data feeds to the SCSN Pasadena and the CGS in Sacramento (Figure 2).

Amplitudes are produced at all three CISN processing centers and shared automatically to facilitate seamless production of ShakeMap. Most recently, proxy wave-servers are used in post-processing to access waveforms in remote data archives, and improve earthquake locations and magnitudes. All products are transmitted via the USGS operated middleware Earthquake Information Distribution System (EIDS) for publishing on the USGS/ANSS web pages. To ensure high availability of earthquake information following major earthquakes, the USGS contracts with a commercial Internet provider to host the ANSS/USGS web pages.

The CISN partners have developed and operate the ANSS Quake Monitoring Software (AQMS) for their earthquake monitoring needs. In 2009, the ANSS began deploying the AQMS software at five other major ANSS data processing centers across the US. The AQMS software is based on the Oracle database, Earthworm modules, a variety of CISN developed software modules, and open source IT software modules for state of health monitoring. A CISN working group curates the Oracle CISN schemas to maintain standardization. The AQMS software includes data acquisition modules, broadcast and capture modules, data analyst and duty-seismologist review modules, and modules for various processing threads such as generation of earthquake locations, magnitudes, moment tensors, and alarming functionality. The AQMS software also has various facilities for data archiving. Recent additions to the software include incorporation of QuakeXML and StationXML to facilitate catalog data and station metadata exchange.

Today the USGS National Earthquake Information Center (NEIC) is the backup for the CISN. The CISN partners share waveforms from selected seismic stations, as well as parametric products such as arrival time picks, hypocenters, and magnitudes in real-time with NEIC. If the CISN does not report within several minutes on an earthquake that just occurred, the NEIC will release available information to ensure timely and accurate reporting. Once CISN has recovered from a failure or significant new information is available, it may update the information previously posted to the USGS hosted and other web sites.

### California Integrated Seismic Network: Region of ANSS



**Figure 2** Schematic diagram showing the major elements and components of the CISN. (Left) list of major elements; (right) the various components of the CISN; CalEMA – California Emergency Management Agency; FEMA – Federal Emergency Management Agency; IRIS/DMC – IRIS Data Management Center; NCEDC – Northern California Earthquake Data Center; RSS- Real Simple Syndication; SCEDC – Southern California Earthquake Data Center; STEP – short-term earthquake probabilities.

### 3 Southern California Seismic Network (SCSN)

The authoritative earthquake-monitoring region of the Southern California Seismic Network (SCSN) extends across southern California, from the US-Mexico international border to Coalinga and Owens Valley in central California. The SCSN also reports on earthquakes in northern Baja California, Mexico because these could possibly cause damage in the United

States. The region is home to almost 20 million inhabitants, including two of the ten largest cities in the United States (Los Angeles and San Diego) and the two largest harbors (Los Angeles and Long Beach) in the Nation. Emergency managers use SCSN information to coordinate rescue operations, guide inspectors in the search for damage, and to facilitate disaster recovery. The media broadcasts the earthquake information to satisfy public need for information.

The SCSN processes and records real-time data from about 425 seismic stations (station map shown in Chapter 8.7). It operates about 300 seismic stations and receives real-time waveform data from an additional 125 stations operated by partner networks. The SCSN seismic stations can be divided into three groups. First, the most modern group of 167 SCSN stations consist of broadband sensors, strong motion sensors, and high resolution dataloggers for digitization and data communications. The data from these stations are used for real-time determination of earthquake locations, magnitudes, moment tensors, and amplitudes for ShakeMap. The second group consists of about 35 stations that are equipped with only strong motion sensors and a datalogger. These stations provide amplitudes for ShakeMap and occasional arrival-time picks. The third group consists of 125 short-period stations that use legacy analog, frequency-modulated (FM) audible tone technology and Earthworm digitizers for data acquisition. These stations are low-cost and can be installed at seismically quiet sites with minimal security. The waveform data can be transmitted for long distances using analog FM radio with very low bandwidth. Although, the amplitude data have limited dynamic range and are often only calibrated using nominal response values, these stations provide arrival-time picks for improved location and depth determination as well as coda duration for determining magnitudes of small earthquakes. Eight Earthworm hubs that are spread across southern California aggregate the SCSN short-period stations and transmit digital waveform data to the central site in Pasadena.

We use SeisNetWatch, initially developed by Caltech and later adopted by ANSS, to track the state of health of the stations. Various tools are used to check delivery and the presence of current waveform data within the processing systems. The tracking of mass centering and issuing of recentering commands is semi-automated. In addition, firmware upgrades on dataloggers are done remotely using command and control tools.

The data flow path from the remote seismic stations includes the on-site equipment, the data communications path, and a data acquisition computer at the central recording site. The last mile data communications to each site may use copper wire, spread-spectrum radios, or optical fiber. The main long-haul data communications include digital phone lines (frame relay or T1 lines) and digital microwave circuits. The waveform data are acquired by front-end processors that broadcast the data on a private local network. Many computers that are connected to this network capture and process the waveforms or capture the data for archiving. For instance, a processing thread of Earthworm modules carries out the real-time picking of arrival-times, association, and hypocenter determination. A different thread harvests amplitudes and determines local magnitudes. Similarly, an earthquake-early-warning processing thread captures the waveform data to facilitate testing of algorithms for ultra rapid determination of source parameters.

The AQMS software provides various buffers to allow for intermittent failure or maintenance of the acquisition and processing system. The dataloggers at broadband and strong motion stations can store several months worth of waveform data on site. In addition, each import server has some amount of waveform data storage as a backup feature. At the central



recording site in Pasadena, wavepools that are populated with waveform data in real-time provide redundant storage for up to four months. These wavepools are accessed by many processes including the processes that perform final data archiving.

The Southern California Earthquake Data Center (SCEDC) archives all the SCSN data and distributes the products of the SCSN. The waveform archiving is done within minutes of the occurrence of the earthquake using a system of request cards. The data center issues request cards for both event triggered waveforms where each request card specifies a time window of waveform data for a selected station channel of data. It also issues request cards for six hour windows of continuous data to the wavepool that is operated as part of the real-time systems. All parametric data determined by the real-time systems are stored in the database. During post-processing, data analysts use tools to display the waveforms, modify or add arrival time picks or amplitudes. They can also relocate the earthquake and update the magnitude. All of these actions result in updates in the database for that particular event. As the SCSN processes the data for detected earthquakes, outside researchers can also access all the data in the database, within minutes or as soon as the data are entered into the archive.

The SCSN products include the southern California earthquake catalog, event gathers (triggers) of waveforms for nearly all events back to 1981 and for some events extending back to 1976, continuous broadband waveforms from 1999 to present, continuous instrumental record of earthquake occurrence in southern California that includes four earthquakes of  $M > 7$  extends back to 1932. The SCSN and the SCEDC have maintained and published a catalog of earthquakes with magnitude of completeness above magnitude 3.25 since 1932 and above magnitude 1.8 since 1980 with consistent magnitudes over the whole time (Hutton, *et al.* 2010). The catalog is used to produce societally useful products such as USGS earthquake hazards maps as well as by seismologists for research.

Subsets of SCSN waveforms are also archived at the IRIS/DMC. In the 1990s Caltech started collecting broadband data using a new seismic network called TERRAscope (network code: TS). These data were at first only archived at the IRIS/DMC. In the late 1990s TERRAscope was merged with the SCSN under the network code CI. In 2004 the SCSN began data sharing with Earthscope/USArray and those data are also archived at the IRIS/DMC. At present, continuous waveform data from a subset of about 65 SCSN broadband and strong motion stations are being archived at the IRIS/DMC.

For more information see related web sites:

<http://earthquake.usgs.gov/monitoring/anss>; <http://www.strongmotioncenter.org>

<http://www.cisn.org>; <http://www.scsn.org>; <http://www.data.scec.org>

<http://www.ncedc.org>; <http://www.iris.edu>

## References

Hutton, L. K., Woessner, J., and Hauksson, E. (2010). Seventy-seven years (1932 – 2009) of earthquake monitoring in southern California., *Bull. Seismol. Soc. Am*, **100** (2); 423-446; doi: 10.1785/0120090130.

| Topic   | An operational authoritative location scheme                                                                                                                                             |
|---------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Rémy Bossu and Gilles Mazet-Roux, EMSC, c/o CEA, Centre DAM IDF, Bruyères le Châtel, 91297 Arpajon Cedex France;<br>E-mail: <a href="mailto:bossu@emsc-csem.org">bossu@emsc-csem.org</a> |
| Version | May 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_8.5                                                                                                                                                |

## 1 Introduction

When an earthquake draws public and media attention, the slightest differences in published epicentral location or magnitude can lead to basic questions such as “Who is right?” As a scientist, the seismologist answers as best he can what are the uncertainties and what are their causes. And invariably the same questions come back: “OK, but who is right? Was the earthquake located 2 km or 5 km to the north of the city?”

This example – which is common for felt events affecting border regions – illustrates how the differences in earthquake parameters are often taken by journalists and citizens as proof that at least one of the publishing organisations must be wrong, thereby raising doubt in the trustworthiness of the seismological community as a whole! Improving the accuracy of earthquake parameters will not solve the problem as small differences will always persist. In essence, this is a public communications issue rather than a scientific problem. To address this issue, a scheme must be defined that specifies the conditions under which an earthquake location determined by a given agency should be considered “authoritative” and should be used as it is by other agencies.

This article describes past efforts to address this issue and the authoritative location scheme approved by the members of the European-Mediterranean Seismological Centre (EMSC) and applied in EMSC real time information services ([www.emsc-csem.org](http://www.emsc-csem.org)) since November 2010. Not considered are the location procedures itself, the model assumptions on which they are based and their inherent errors. They are discussed in detail in IS 11.1.

## 2 Previous authoritative location schemes

Past efforts to define an authoritative location scheme at the EMSC and at other agencies were based on static geographical areas defined from the geographical extents of the seismic networks. A location is generally considered to be reliable when the azimuthal gap of the reporting stations is small. By definition, locations that fall within the bounds of reporting networks exhibit good azimuthal coverage. Therefore, these areas based on the network extents were supposed to identify the source for the *best* locations. Furthermore, this strategy implicitly recognises that local institutes are generally in the best position to accurately locate local earthquakes thanks to their knowledge of the local conditions.

This scheme properly addresses the majority of earthquake locations. However, it proved problematic in several cases. For example, it does not define which network should prevail in areas monitored by several networks. Such an overlap is not restricted to border regions: there

are also a number of European countries in which several seismic networks are in operation. Choosing *the* authoritative network for such regions is touchy and may become a political issue: how to prevent any national institute from claiming to be authoritative over its entire national territory even when part of that territory is poorly covered by its network?

In the end, these difficulties result from the fact that the underlying assumptions – that location accuracies are uniform within network boundaries and constant over long periods of time – are not always satisfied. Location accuracies depend on both the network geometry and on the relative spatial distribution and number of stations used in the calculation of the earthquake location.

The proposed authoritative scheme described below is based on the quality of the geographical coverage of the stations which actually report the considered event. It takes into account the actual operating status of each individual station at the time of the earthquake's occurrence, and only considers those stations that reported the event and were used in the calculation of the location.

### **3 An operational authoritative location scheme**

#### **3.1 Assumptions**

The simple and basic assumption of the proposed scheme is that locations which are both reliable and accurate must be considered to be authoritative and must therefore be published without modification for public consumption. Relocations should be restricted only to cases where a significant quality improvement can be expected. In essence, defining an authoritative location scheme amounts to the definition of the criteria for what can be considered a reliable and accurate location.

In the discussion below, we only consider the epicentral location. It is assumed that if the epicentral location is accurate and reliable, the focal depth is well constrained. Theoretically, this may not always be the case. In practice however, as we will see later, accurate and reliable locations are generally produced by local institutes which are in the best position to provide reliable depth estimates. Furthermore, as focal depth is often ignored by the public, it is a less critical element of a public communication scheme than the epicentral location or magnitude.

#### **3.2 Criteria for accurate locations**

Locations are considered to be accurate if they satisfy either of the 2 sets of criteria defined by Engdahl et al. (2001), or Bondár and McLaughlin (2009). Both criteria characterize the geometry and azimuthal distribution of the reporting stations at short distances (250 km), and were both derived from the study of explosions. The considered criteria are referred to as GT5 (for Ground Truth). If satisfied, the epicentral location is expected to be within 5 km of the actual epicentre, within a formal confidence level. For more details on the GT concept, global project and GT data base at the International Seismological Centre see section 7 in IS 11.1.

Table 1 and 2 present the GT5 criteria as defined by Engdahl et al. (2001) at a 90% confidence level and by Bondár and McLaughlin (2009) at a 95% confidence level, respectively, hereinafter referred to as GTA and GTB. The GTB criteria are more restrictive

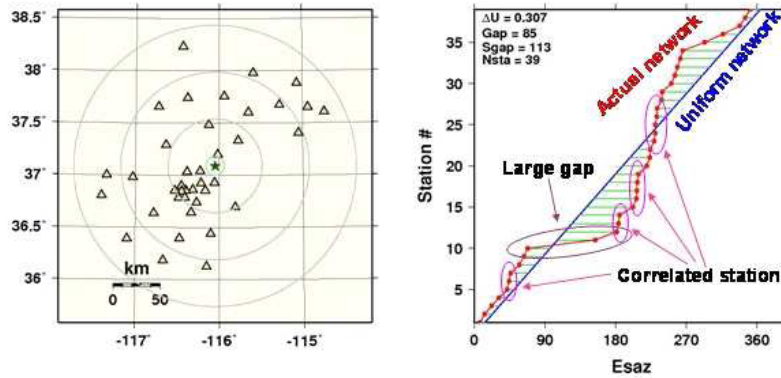
than those of the GTA, but at the same time they select events not identified previously. Allowing the choice of either of the two sets of criteria maximizes the number of potential authoritative locations. Table 2 and Figure 1 present the alternative criteria defined by Bondár and McLaughlin (2009)

**Table 1** GT5 criteria at 90% confidence level for local networks, as defined by Engdahl et al. (2001).

|                                                                         |
|-------------------------------------------------------------------------|
| At least 10 stations at epicentral distance $\leq 250$ km               |
| At least one station at epicentral distance $\leq 30$ km                |
| Azimuthal gap defined with stations closer than 250 km $\leq 110^\circ$ |

**Table 2** GT5 criteria at 95% confidence level for local networks, as defined by Bondár and McLaughlin (2009). The secondary gap is the largest azimuthal gap filled by a single station. The criterion on the secondary gap implies a minimum of 5 stations. The definition of the network quality metric,  $\Delta U$ , is given in Figure 1.

|                                                                                   |
|-----------------------------------------------------------------------------------|
| At least one station at epicentral distance $\leq 10$ km                          |
| $\Delta U \leq 0.35$ when considering stations closer than 150 km                 |
| Secondary azimuthal gap defined with stations closer than 150 km $\leq 160^\circ$ |

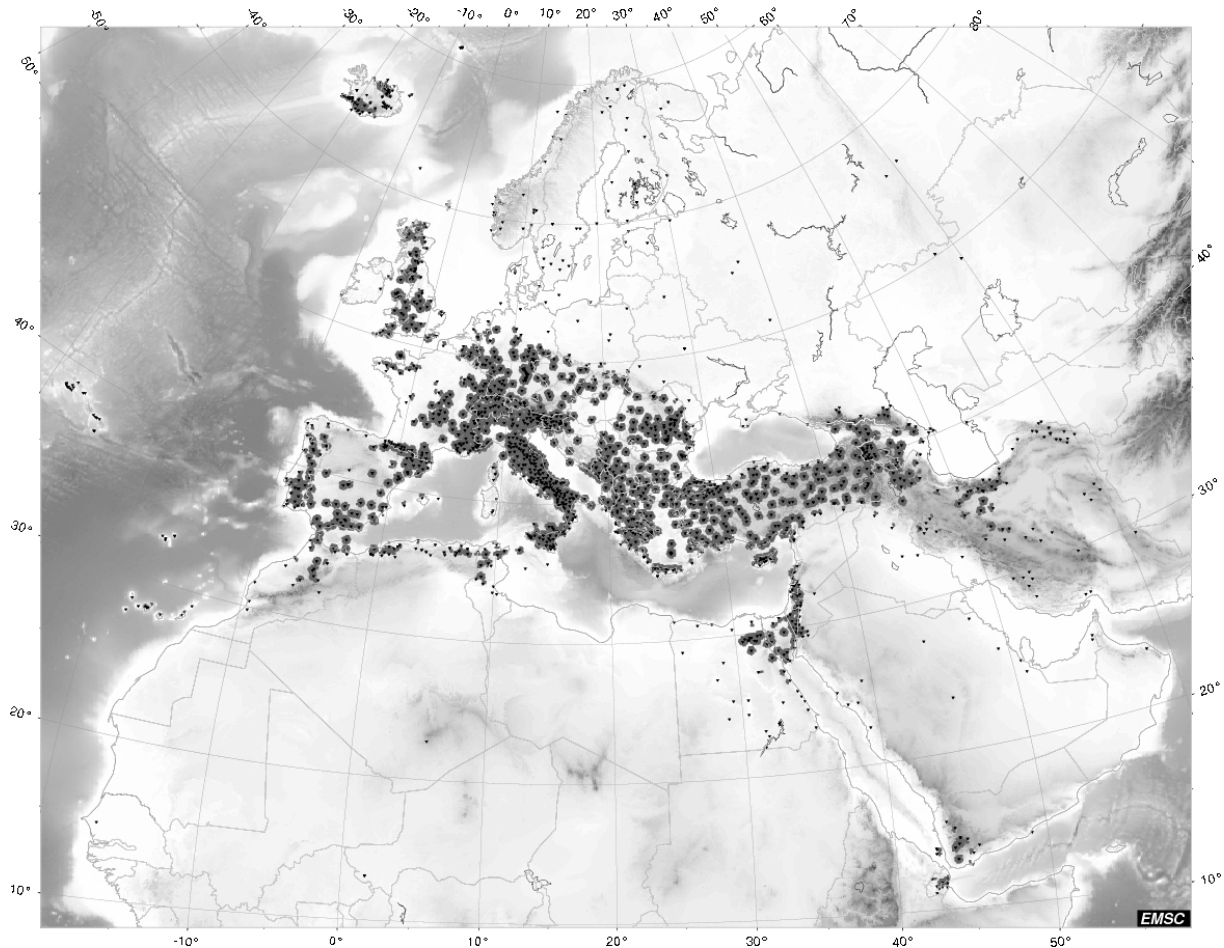


$$\Delta U = \frac{4 \sum |esaz_i - (unif_i + b)|}{360N}, \quad 0 \leq \Delta U \leq 1;$$

$$unif_i = \frac{360i}{N}; \quad b = \overline{esaz} - \overline{unif}$$

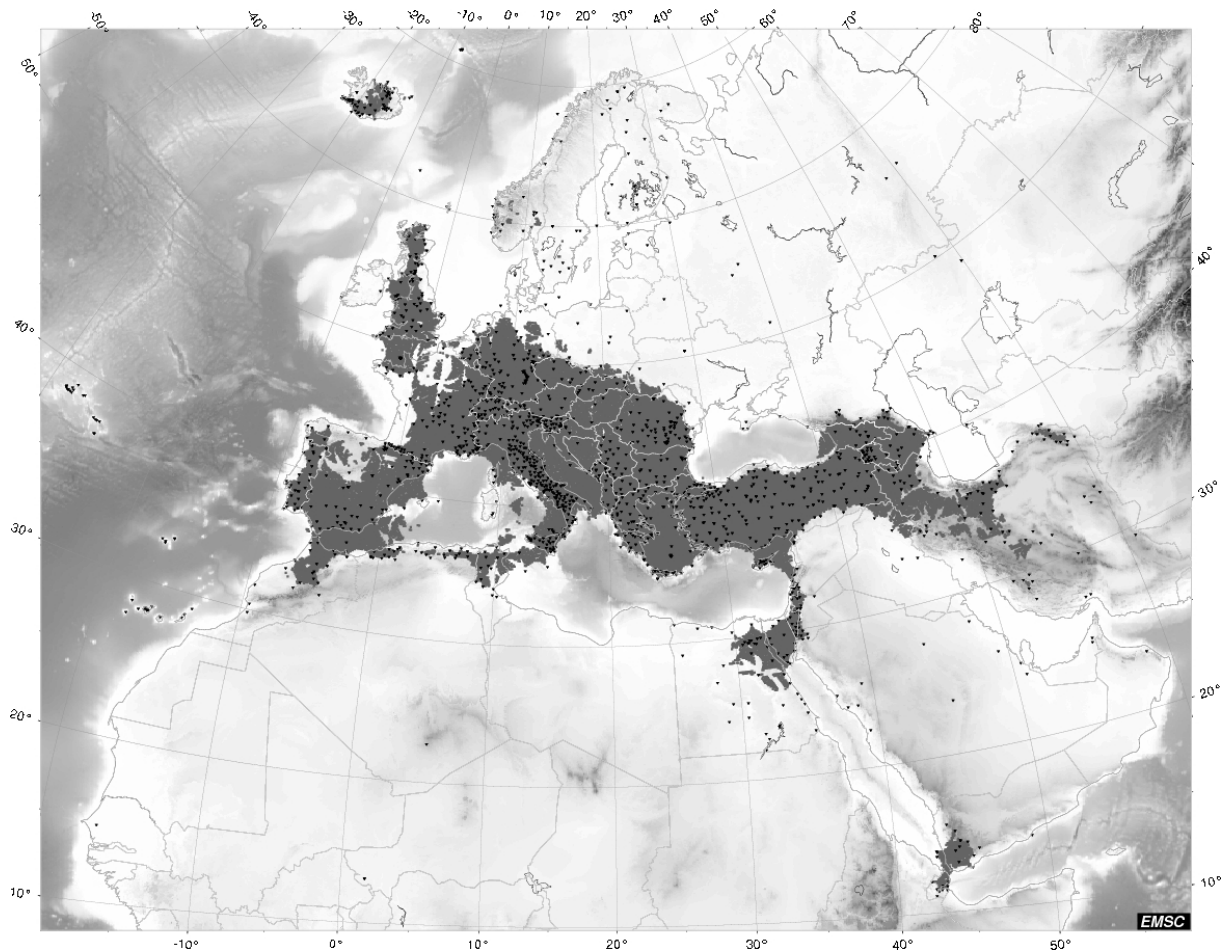
**Figure 1** Definition of the network quality metric,  $\Delta U$ . If the stations, ordered by event-to-station azimuth were distributed uniformly they would align along a straight diagonal line and  $\Delta U$  would be zero. The dots in the right-hand panel represent the actual geometry of the network as plotted in the left-hand panel. The network quality metric is defined as the normalised area between the best fitting uniformly-spaced network and the actual network, shown shaded by grey lines in this example. The lower the  $\Delta U$  value, the more uniform the network geometry. Copy of Figure 5 in Bondár and McLaughlin (2009, p. 468), with © granted by the Seismological Society of America.

The constraint on the closest station, within 30 km for GTA and 10 km for GTB, strongly restricts the geographical area where authoritative locations could potentially be found for a given network (Figure 2). In practice, less than 5% of the locations collected in real time by the EMSC satisfy these criteria.



**Figure 2** Geographical area where GTA or GTB criteria are satisfied when considering all the seismic stations which contribute to the EMSC real time earthquake information services. Effectively, there cannot be any GTA or GTB locations in regions where inter-station distance is greater than 30 km.

The constraint on the closest station improves the confidence in depth for crustal events, but has no significant effect on epicentral location accuracy (e.g., Bondár et al., 2004). Therefore, in agreement with our basic assumption that our authoritative scheme is based on epicentral location only, this constraint is dropped for the purposes of defining what may be considered an accurate epicentral location. It doubles to about 10% the number of locations provided in real time which are potentially authoritative. More importantly, it greatly enlarges the geographical regions where such authoritative locations could potentially be found (Figure 3).



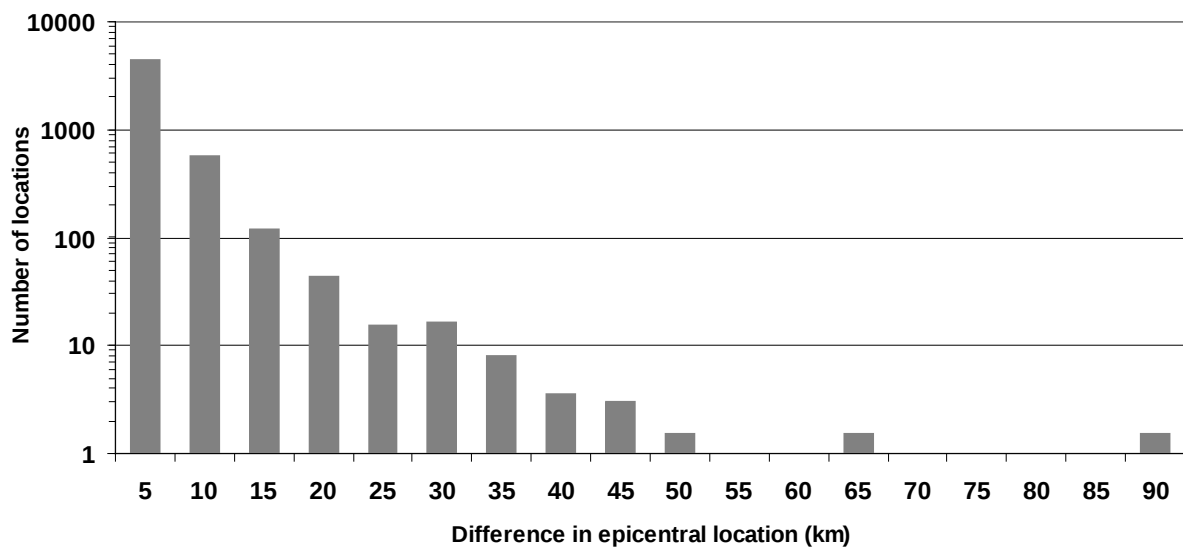
**Figure 3** Geographical area where authoritative locations could potentially be found for the EMSC real time services after relaxing the closest station constraint. This result is based on the assumption that earthquakes are reported by all the stations contributing to the EMSC real time services and located within 250km of distance.

### 3.3 Criteria for reliable locations

As for any measurement, a location is considered to be reliable if it can be independently and consistently reproduced within its inherent uncertainties. The estimation of location uncertainties is a complex issue (e.g., Thurber and Engdahl, 2000). By taking as reference locations the final solutions produced by the local networks and published in their final bulletins, Mazet-Roux et al. (2010) show that on average the locations produced in real time by the EMSC are within 10 to 12 km of the reference locations. The location procedures currently in place at the EMSC (velocity model and algorithm) are therefore very unlikely to produce a mislocation greater than 15 km when using the same dataset. We therefore set the criterion for a reliable location as being reproducible to within less than 15 km; larger differences are assumed to indicate an unreliable location.

The vast majority (98%) of the accurate locations (per definition in section 3.2) are also effectively reproduced to within less than 15 km, and can therefore also be labelled as reliable (Figure 4). However, there are still cases in which the differences in location are significant. Potential causes for these discrepancies were investigated. In some cases, the differences in

location were caused by data format or station code errors resulting in incorrect data inputs. However, in several cases no clear cause was identified. It was noted, however, that several of the submitted locations did not appear in the final bulletin of the reporting institute, or the locations in the bulletin were more than 15 km from the initially reported locations. These examples reinforce the importance of independent evaluation of location accuracy and reliability to avoid the publication of incorrect locations, even if the actual causes remain unidentified.



**Figure 4** Distribution of the difference in epicentral location between the accurate locations (per definition in section 3.2) provided to the EMSC and EMSC relocations using the same dataset. Results obtained from the 5,265 accurate locations provided to the EMSC in real time between Sept. 2008 and Sept. 2009.

### 3.4 Discussion

The authoritative location scheme presented in this work defines the conditions under which a real time epicentral location provided by one of the contributing agencies to the EMSC will be published without any modification in the EMSC public real time services. It does not intend to identify the best location, nor does it address any scientific issue. The intent is to avoid possible communication difficulties with the media and the public caused by differences in published earthquake locations, even if the differences are scientifically and statistically insignificant.

In practice, we consider a location to be authoritative when it is both reliable and accurate. The accuracy criteria are derived from the GT5 criteria. An accurate location is considered to be reliable when it can be reproduced to within 15 km, using the same dataset but different location procedures (velocity model, location algorithm).

The advantages of the proposed scheme are numerous. It depends only on the spatial distribution of the stations which have been used to compute the locations. It does not require specific criteria for border regions. Each location is evaluated independently and instantaneously, which fits within the requirements of rapid information services where

updates are numerous. The most significant improvement when compared to the previous procedures in place at the EMSC is that the new scheme does not depend on the magnitude of the earthquake. The old scheme, applied up through November, 2010, considered a location as authoritative if it was reported by only one network and located within the network geographical boundaries; Earthquakes reported by more than one network had not authoritative location and the published location was computed merging all available data. Larger earthquakes being recorded at larger distances are generally reported by several networks. Then, under the old scheme, the large earthquakes which typically attract public and media attention had no authoritative location.

The new scheme solves this problem. It was formally accepted by the EMSC General Assembly in September, 2010, and implemented in November of the same year in the EMSC real time information services.

Authoritative locations have also been identified within the Euro-Med bulletin (Godey et al., 2006) which merges 78 individual bulletins. Data contributors are encouraged to report, when possible, restricted data obtained through bilateral exchanges to ensure authoritative locations in border regions are identified as such. These restricted data are not published by the EMSC. We next plan to use the GT5 and possibly GT10 criteria to define the best dataset to be used to relocate earthquakes when no authoritative location is available.

## 4 Conclusion

The EMSC has developed and implemented a simple authoritative location scheme for its rapid earthquake information services. This development addresses a long-standing issue in public and media communications. Further improvements in the way locations are computed are also foreseen. So far, no complaints have been received from EMSC members, and the scheme seems to offer a pragmatic and satisfactory solution. The next step, and probably the biggest challenge for rapid earthquake information services, will be to propose a convincing and acceptable authoritative magnitude scheme.

## Acknowledgement

The authors would like to thank the EMSC members and the members of the IASPEI working group on authoritative locations for the fruitful discussions. We also thank Linus Kamb and Peter Bormann for improving the manuscript.

## References

- Bondár I. and McLaughlin, K. L. (2009). A new ground truth data set for seismic studies. *Seismol. Res. Lett.*, **80**(3), 465-472
- Bondár, I., Myers, S. C., Engdahl, E. R., and Bergman, E. A. (2004). Epicenter accuracy based on seismic network criteria. *Geophys. J. Intern.*, **156**, 483–496; doi10.1111/j.1365-246X.2004.02070.x.
- Engdahl, E. R., Bondar, I., Bergmann, E., and Firbas, P. (2001). IASPEI Working group on reference events: 1999-2001 activity report.



- Godey, S., Bossu, R., Guilbert, J., and Mazet-Roux, G. (2006). The Euro-Mediterranean bulletin: a comprehensive seismological bulletin at regional scale. *Seism. Res. Lett.*, **77**, 460-474.
- Mazet-Roux, G., Bossu, R., Carreño, E., Guilbert, J., Gilles, S., and Roussel, F. (2010). Report on 2010 real time activities. *EMSC report*, pp. 38, [http://www.emsc-csem.org/DOC/EMS\\_DOCS/EMSC\\_RT\\_activities\\_2010.pdf](http://www.emsc-csem.org/DOC/EMS_DOCS/EMSC_RT_activities_2010.pdf).
- Thurber, C. H., and Engdahl, E. R. (2000). Advances in global seismic event location. In: in Thurber and Rabinowitz (2000), 3-22.

| Topic   | Identification and collection of ground truth events                                                                                                                                                    |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>István Bondár</b> , International Seismological Centre,<br>Pipers Lane, Thatcham, RG19 4NS, United Kingdom<br>Fax: +44 1635 872351<br>E-mail: <a href="mailto:istvan@isc.ac.uk">istvan@isc.ac.uk</a> |
| Version | October, 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_8.6                                                                                                                                                          |

## 1 Introduction

Accurate event locations are essential for seismic hazard studies and for developing and testing three-dimensional velocity models of the Earth. However, as the main objective of producing earthquake catalogues is to achieve completeness down to the lowest detectable magnitude levels, published bulletins inevitably contain a mixture of accurate, good and poor locations. Thus, event locations in earthquake catalogues should always be treated with caution.

Published bulletins give little indication about the accuracy of the event locations. One common mistake is to consider the formal uncertainties (error ellipse, origin time and depth uncertainties) as measures of *accuracy*. The error ellipsoid provides a statistical estimate of *precision* derived from the *a posteriori* (Flinn, 1965), *a priori* (Evernden, 1969), or a mixture of the two (Jordan and Sverdrup, 1981) distributions of travel-time residuals of the observations with respect to the underlying travel-time and error-distribution models and the procedure of calculation used in the location (see Glossary terms *accuracy* and *precision*).

The calculation of formal uncertainties relies on the assumption that the error processes are Gaussian, have zero mean and are uncorrelated. Unfortunately, these assumptions are often violated in seismic locations. Picking errors may exhibit heavy tails (Buland 1986) and onset times tend to be picked late with decreasing event size (Billings et al., 1994; Douglas et al., 1997, 2005).

More crucially, travel-time predictions suffer from systematic errors due to three-dimensional velocity structures left unmodeled by the commonly used Earth models. One of the classic examples of location bias due to unmodeled 3D Earth structure is the Longshot explosion (29, October 1965, Aleutian Islands). The ray paths to nearly all stations sample the Aleutian slab, a subducted oceanic lithosphere with high seismic velocity, causing arrival-time predictions for these paths to be systematically late, which results in a 26 km location error to the north-northwest (Herrin and Taggart, 1968; Bormann, 1972). Note that such a travel-time prediction bias can be further amplified by the fact that similar ray paths from closely spaced stations will generate correlated travel-time prediction errors (Chang et al., 1983; Myers and Schultz 2000; Bondár and McLaughlin, 2009a; Bondár and Storchak, 2011).

Since the error ellipse measures precision but not accuracy, researchers considered various metrics to assess the location accuracy of events based on network geometry. We summarize these efforts below and describe the IASPEI Reference Event List maintained by the International Seismological Centre (ISC; [www.isc.ac.uk](http://www.isc.ac.uk)).

## 2 Identifying Ground Truth (GT) events

Sweeney (1996) investigated the feasibility of selecting ‘reference events’ (events where the hypocentres can be considered known to high accuracy) in continental regions from global bulletins, such as those published by the International Seismological Centre and the US Geological Survey National Earthquake Information Centre (NEIC) that contain predominantly teleseismic arrival time data. He suggested that locations from these catalogues have an accuracy of 10–15 km when the largest azimuthal gap between stations surrounding the epicenter is less than  $200^\circ$  and at least 50 first-arriving P phases are used. Sweeney (1998) revised these selection criteria and found 15 km (or better) epicentre accuracy for teleseismic networks with an azimuthal gap of less than  $90^\circ$  and with at least 50 first-arriving P phases that were used in the location.

Engdahl et al., (EHB; 1998) produced a “groomed” ISC catalogue by using the *ak135* travel-time tables (Kennett et al., 1995), later phase (including depth phases) arrival times, and station-specific travel-time corrections. Myers and Schultz (2001) estimated that the epicentre accuracy in the EHB catalogue is 15 km or better at the 95% confidence level for events not in subduction zones when the largest azimuthal gap is less than  $90^\circ$ .

Bondár *et al.* (2001) introduced the nomenclature of ‘ground truth’ categories (GT<sub>x</sub>, where ‘x’ designates epicentre location accuracy in kilometres (i.e., the true epicentre lies within ‘x’ km of the estimated epicentre) to describe the location accuracy of events in the Ground Truth data set assembled at the Prototype International Data Centre, Arlington, USA (1995-2002).

Bondár et al. (2004a) developed GT selection criteria based on network geometry to assess the location accuracy of events published in earthquake bulletins. The selection criteria identify GT5 candidate events at the 95% confidence level if they are located:

- (1) with at least 10 stations, all within 250 km;
- (2) with an azimuthal gap of less than  $110^\circ$ ;
- (3) with a secondary azimuthal gap of less than  $160^\circ$ ;
- (4) with at least one station within 30 km from the epicentre.

The GT5 selection criteria above have been successfully applied to identify GT5 candidate events by introducing the notion of secondary azimuthal gap (defined as the largest azimuthal gap which has been filled by just a single station). The secondary azimuthal gap is particularly sensitive to a station with disproportionately large data importance that may introduce location bias. The secondary azimuthal gap criterion implicitly invokes constraints on both the azimuthal gap and the minimum number of stations, but it is insensitive to clustering of stations that may generate correlated travel-time prediction error structures.

To remedy these issues, Bondár and McLaughlin (2009b) revisited the GT5 selection criteria. To avoid the *Pg/Pb* and *Pg/Pn* cross-over distance ranges, which are prone to phase identification errors, they considered stations up to 150 km only. They also introduced a new metric that measures the deviation from the optimal, azimuthally uniformly distributed network of local stations. The network quality metric is defined as the mean absolute deviation from the best-fitting uniformly distributed network of stations and the actual network. Provided that the event-to-station azimuths are sorted by increasing values, the metric is given by the expression

$$\Delta U = \frac{4 \sum_{i=1}^N |esaz_i - (unif_i + b)|}{360N}, 0 \leq \Delta U \leq 1$$

where  $N$  is the number of stations,  $esaz_i$  is the  $i$ th event-to-station azimuth,  $unif_i = 360i / N$  for  $i = 0, \dots, N - 1$ , and  $b = \text{avg}(esaz_i) - \text{avg}(unif_i)$ . The network quality metric is normalized;  $\Delta U$  is 0 when the stations are uniformly distributed in azimuth and it is 1 when all the stations are aligned at the same azimuth. Since large azimuthal gaps or potentially correlated stations (stations at similar azimuth) introduce deviations from the optimal, uniformly distributed network, the metric is sensitive to both sources of potential bias. The metric is similar to the non-parametric Kolmogorov-Smirnov statistic, which measures the maximum absolute deviation between two distributions.

The revised selection criteria identify GT5 candidate events at the 95% confidence level if an event is (re)located:

- (1) with stations within 150 km;
- (2) with  $\Delta U \leq 0.35$ ;
- (3) with a secondary azimuthal gap  $\leq 160^\circ$ ;
- (4) with at least one station within 10 km from the epicentre.

Depth and origin time typically have lower accuracy than the epicentre, since the accuracy of these parameters is not governed by the network geometry but depends on the accuracy of the velocity model as well as the availability of depth-sensitive phases (e.g., depth phases such as pP and sP, or of core reflections such as PcP). The criterion for the existence of a station virtually on top of the event is aimed to ensure that there is sufficient resolution in the data to constrain the depth, so that depth will not trade-off with the epicentre.

The Bondár et al. (2004a, 2009b) criteria were used to identify candidate GT5 events in the EHB bulletin for the IASPEI Reference Event List hosted by the ISC ([www.isc.ac.uk](http://www.isc.ac.uk)). Note that the location accuracy criteria described in this section are most relevant to continental earthquakes. Subduction zone events are likely to have a larger (and systematic) location uncertainty due to the travel-time bias introduced by the subducting slab, even if the station coverage satisfies the location accuracy criteria. Nevertheless, identifying GT5 events in subduction zones becomes more viable owing to the increasing number of ocean bottom seismometer arrays deployments worldwide. Agencies operating ocean bottom seismometer deployments are encouraged to share their data with the international scientific community.

It should be emphasized that the GT selection criteria above assume that no special effort has been made to remove location bias through the use of an optimal velocity model or path-specific travel-time predictions, or through special analysis of waveforms to improve phase picks. With such an expert seismic analysis, location accuracy of a few kilometers can be achieved (e.g. Boomer et al., 2010; Richards et al., 2006).

Detailed location studies typically make use of multiple-event methods (e.g. Douglas, 1967; Dewey 1972; Jordan and Sverdrup, 1981; Pavlis and Booker, 1983; Got et al., 1994; Waldhauser and Ellsworth, 2000; Zhang and Thurber, 2003; Engdahl et al., 2006; Bondár et al., 2008; Myers et al., 2007, 2009). Multiple event methods are known for precise *relative*

locations, but a loss of accuracy – manifested as a consistent bias for all locations – is well documented (Douglas, 1967; Jordan and Sverdrup, 1981; Pavlis and Booker, 1983). Therefore modern multiple event location techniques utilize independent GT information such as existing reference events (e.g. Ritzwoller et al., 2003; Bondár et al., 2004b), seafloor bathymetry (Pan et al., 2002), satellite imagery (e.g. Bennett et al., 2010; Fisk 2002), InSAR (e.g. Biggs et al., 2006; Parsons et al., 2006), and active fault lines (Waldhauser and Richards, 2004) to provide absolute location constraints.

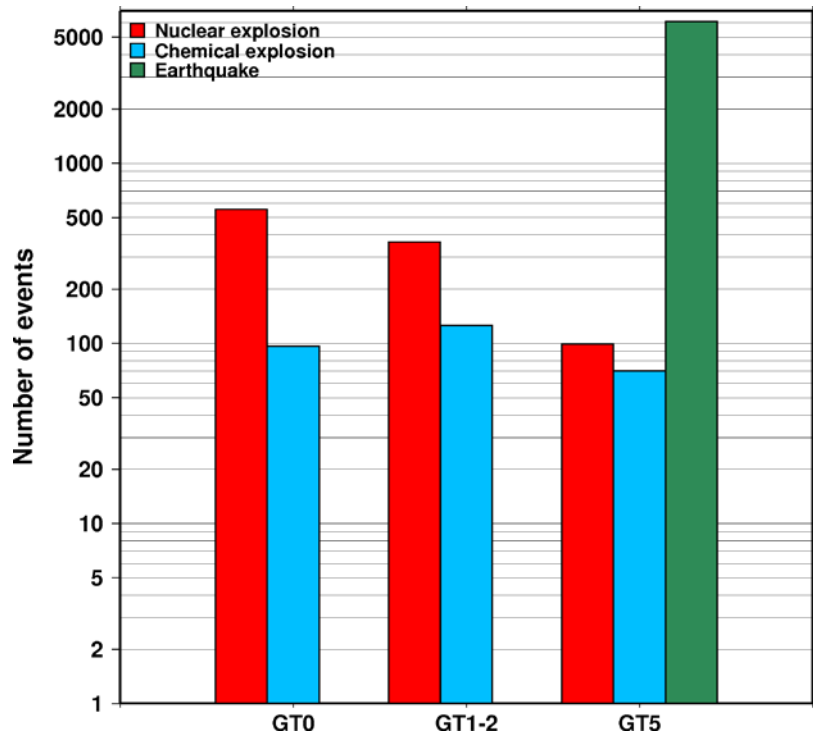
Despite the utilization of multiple event methods, location precision of a few kilometers still requires the attention of an expert to construct a well-posed multiple-event problem, develop or validate travel-time models, and groom the data set.

### **3 IASPEI Reference Event List**

The IASPEI Commission on Seismic Observation and Interpretation (CoSOI) Working Group on Reference Events for Improved Locations has coordinated the global effort of collecting and validating ground truth events in the past decade. The International Seismological Centre (ISC) hosts the database of the IASPEI reference event list. Ground truth events can be searched, downloaded or submitted through the ISC website, [www.isc.ac.uk/GT/index.html](http://www.isc.ac.uk/GT/index.html).

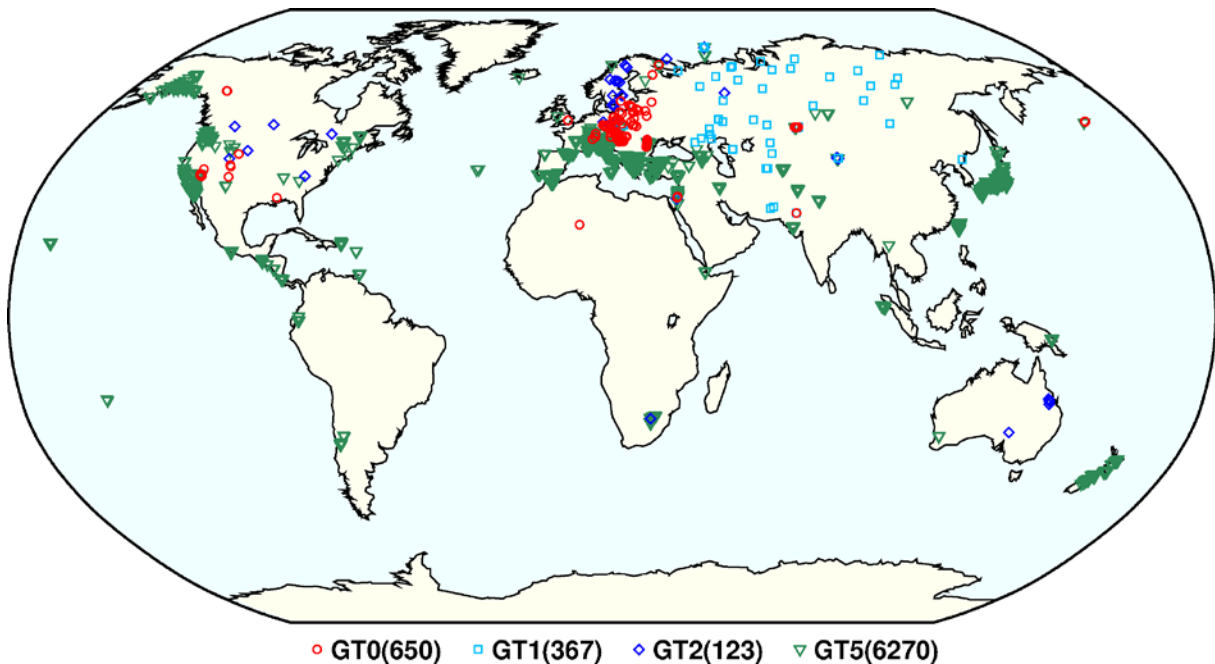
The ISC updates the ground truth database annually. Because the ISC bulletin contains parametric data reported by various agencies with little indication of the procedures of how the actual measurements were made, it is advisable to follow a more conservative approach when identifying GT events. Therefore the Bondár and McLaughlin (2009b) ground truth selection criteria, that were specifically developed for assessing the quality of locations in a bulletin without any further special analyses, are applied to select GT candidate events from the reviewed ISC bulletin. The GT candidate events are relocated using the local stations only (within 150 km of the epicentre). The subsequent outlier analysis may reject some of the GT candidates. The members of the IASPEI CoSOI Working Group review the identified GT events, and the accepted GT events are added to the IASPEI Reference Event List. The Working Group also vets events submitted by researchers.

The IASPEI Reference Event List contains 7,410 GT0-5 events (as of August 2011) with some 525,000 phase onset time readings. The GT database consists of both earthquakes and explosions (nuclear or chemical). Figure 1 shows the number of events by type and category.



**Figure 1** Number of events by event type and GT category.

Figure 2 shows the geographical distribution of the GT events color, and symbol coded by their GT category. Note the uneven distribution of events due to the lack of dense local networks on many parts of the globe. In order to fill in gaps in GT coverage, we encourage researchers who run temporary local network deployments to submit GT event candidates. This particularly applies to Africa, South America, and South-East Asia.



**Figure 2** Geographical distribution of GT0-5 events in the IASPEI Reference Event List.

## Acknowledgement

Thanks go to Bob Engdahl for his careful review and proposed valuable amendments.

## References

- Bennett, T. J., Oancea, V., Barker, B. W., Kung, Y.-L., Bahavar, M., Kohl, B. C., Murphy, J. R., and Bondár, I. K. (2010). The Nuclear Explosion Database (NEDB): A new database and web site for accessing nuclear explosion source information and waveforms. *Seism. Res. Let.*, **81**, 12- 25.
- Biggs, J., E. Bergman, B. Emmerson, G.J. Funning, J. Jackson, B. Parsons and T.J. Wright. (2006). Fault identification for buried strike-slip earthquakes using InSAR: the 1994 and 2004 Al Hoceima, Morocco earthquakes. *Geophys. J. Int.*, **166**, 1347–1362, doi:10.1111/j.1365-246X.2006.03071.x.
- Billings, S.D., M.S. Sambridge and B.L.N. Kennett. (1994). Error in hypocenter location: picking, model, and magnitude dependence. *Bull. Seism. Soc. Am.*, **84**, 1978–1990.
- Bondár, I., X. Yang, R.G. North and C. Romney. (2001). Location calibration data for CTBT monitoring at the Prototype International Data Center. *Pure appl. Geophys.*, **158**, 19–34.
- Bondár, I., S.C. Myers, E.R. Engdahl and E.A. Bergman. (2004a). Epicenter accuracy based on seismic network criteria. *Geophys. J. Int.*, **156**, 483–496; doi 10.1111/j.1365-246X.2004.02070.x.
- Bondár, I., E.R. Engdahl, X. Yang, H.A.A. Ghalib, A. Hofstetter, V. Kirichenko, R. Wagner, I. Gupta, G. Ekström, E. Bergman, H. Israelsson and K. McLaughlin. (2004b). Collection of a reference event set for regional and teleseismic location calibration. *Bull. Seism. Soc. Am.*, **94**, 1,528–1,545.
- Bondár, I., E. Bergman, E.R. Engdahl, B. Kohl, Y.-L. Kung and K. McLaughlin. (2008). A hybrid multiple event location technique to obtain ground truth event locations. *Geophys. J. Int.*, **175**, 185–201; doi:10.1111/j.1365- 246X.2008.03867.x.
- Bondár, I. and K. McLaughlin. (2009a). Seismic location bias and uncertainty in the presence of correlated and non-Gaussian travel-time errors. *Bull. Seism. Soc. Am.*, **99**, 172-193.
- Bondár, I. and K. McLaughlin (2009b). A new ground truth data set for seismic studies, *Seism. Res. Let.*, **80**, 465-472.
- Bondár, I. and D. Storchak. (2011). Improved location procedures at the International Seismological Centre. *Geophys. J. Int.*, **186**, 1220-1244, doi:10.1111/j.1365-246X.2011.05107.x
- Boomer, K.B, R.A. Brazier and A.A. Nyblade. (2010). Empirically based ground truth criteria for seismic events recorded at local distances on regional networks with application to Southern Africa. *Bull. Seism. Soc. Am.*, **100**, 1785-1791.
- Bormann, P. (1972). A study of travel-time residuals with respect to the location of teleseismic events from body-wave records at Moxa station. *Gerlands Beitr. Geophysik*, **81**(1/2), 117-124.
- Buland, R. (1986). Uniform reduction error analysis. *Bull. Seism. Soc. Am.*, **76**, 217–230.

- Chang, A.C., R.H. Shumway, R.R. Blandford and B.W. Barker. (1983). Two methods to improve location estimates - preliminary results. *Bull. Seism. Soc. Am.*, **73**, 281–295.
- Dewey, J.W. (1972). Seismicity & tectonics of Western Venezuela. *Bull. Seism. Soc. Am.*, **62**, 1711–1751.
- Douglas, A. (1967). Joint epicentre determination, *Nature*, **215**, 47–48.
- Douglas, A., D. Bowers and J.B. Young. (1997). On the onset of P seismograms. *Geophys. J. Int.*, **129**, 681–690.
- Douglas, A., J.B. Young, D. Bowers and M. Lewis. (2005). Variation in reading error in P times for explosions with body-wave magnitude. *Phys. Earth Planet. Inter.*, **152**, 1–6.
- Engdahl, E.R., R. van der Hilst and R. Buland. (1998). Global teleseismic earthquake relocation with improved travel times and procedures for depth determination. *Bull. Seism. Soc. Am.*, **88**, 722–743.
- Engdahl, E.R., J.A. Jackson, S.C. Myers, E.A. Bergman and K. Priestley. (2006). Relocation and assessment of seismicity in the Iran region. *Geophys. J. Int.*, **167**, 761–778, doi: 10.1111/j.1365-246X.2006.03127.x.
- Evernden, J. (1969). Precision of epicenters obtained by small numbers of world-wide stations. *Bull. Seism. Soc. Am.*, **59**, 1365–1398.
- Fisk, M. (2002). Accurate locations of nuclear explosions at the Lop Nor test site using alignment of seismograms and IKONOS satellite imagery. *Bull. Seism. Soc. Am.*, **92**, 2911–2925.
- Flinn, E. (1965). Confidence regions and error determinations for seismic event location, *Rev. Geophys.*, **3**, 157–185.
- Got, J.-L., J. Fréchet and F.W. Klein. (1994). Deep fault plane geometry inferred from multiple relative relocation beneath the south flank of Kilauea, *J. Geophys. Res.*, **99**, 15375–15386.
- Herrin, E. and J. Taggart. (1968). Source bias in epicenter determinations. *Bull. Seism. Soc. Am.*, **58**, 1791–1796.
- Jordan, T.H. and K.A. Sverdrup. (1981). Teleseismic location techniques and their application to earthquake clusters in the South-central Pacific. *Bull. Seism. Soc. Am.*, **71**, 1105–1130.
- Kennett, B.L.N., E.R. Engdahl and R. Buland. (1995). Constraints on seismic velocities in the Earth from travel times. *Geophys. J. Int.*, **122**, 108–124.
- Myers, S.C. and C.A. Schultz. (2000). Improving sparse network seismic location with Bayesian kriging and teleseismically constrained calibration events. *Bull. Seism. Soc. Am.*, **90**, 199–211.
- Myers, S.C. and C.A. Shultz. (2001). Statistical characterization of reference event accuracy, *Seism. Res. Let.*, **72**, 244.
- Myers, S.C., G. Johannesson and W. Hanley. (2007). A Bayesian hierarchical method for multiple-event seismic location, *Geophys. J. Int.*, **171**, 1049–1063, doi:10.1111/j.1365-246X.2007.03555.x.
- Myers, S.C., G. Johannesson and W. Hanley. (2009). Incorporation of probabilistic seismic phase labels into a Bayesian multiple-event seismic locator, *Geophys. J. Int.*, **177**, 193–204, doi: 10.1111/j.1365-246X.2008.04070.x.



- Pan, J., M. Antolik and A.M. Dziewonski. (2002). Locations of midoceanic earthquakes constrained by seafloor bathymetry, *J. Geophys. Res.*, **107**(B11), 2310, EPM8.1–EPM8.13, doi:10.1029/2001JB001588.
- Parsons, B., T. Wright, P. Rowe, J. Andrews, J. Jackson, R. Walker, M. Khalib, M. Talebian, E. Bergman and E.R. Engdahl. (2006). The 1994 Sefidabeh (eastern Iran) earthquakes revisited: new evidence from satellite radar interferometry and carbonate dating about the growth of an active fold above a blind thrust fault, *Geophys. J. Int.*, **164**, 202–217, doi:10.1111/j.1365-246X.2005.02655.x.
- Pavlis, G.L. and J.R. Booker. (1983). Progressive multiple event location (PMEL), *Bull. Seism. Soc. Am.*, **73**, 1753–1777.
- Richards, P.G., F. Waldhauser, D. Schaff and W.-Y. Kim. (2006). The applicability of modern methods of earthquake location, *Pure appl. Geophys.*, **163**(2–3), 351–372, doi:10.1007/s00024-005-0019-5.
- Ritzwoller, M.H., N.M. Shapiro, E.A. Levshin, E.A. Bergman and E.R. Engdahl. (2003). Ability of a global three-dimensional model to locate regional events, *J. Geophys. Res.*, **108**(B7), 2353, doi: 10.1029/2002JB002167.
- Sweeney, J.J. (1996). Accuracy of teleseismic event locations in the Middle East and North Africa, *Lawrence Livermore National Laboratory*, UCRL- ID-125868.
- Sweeney, J.J. (1998). Criteria for selecting accurate event locations from NEIC and ISC bulletins, *Lawrence Livermore National Laboratory*, UCRL-JC-130655.
- Waldhauser, F. and W.L. Ellsworth. (2000). A double-difference earthquake location algorithm: Method and application to the Northern Hayward fault, California, *Bull. Seism. Soc. Am.*, **90**, 1353–1368.
- Waldhauser, F., and P.G. Richards. (2004). Reference events for regional seismic phases at IMS stations in China, *Bull. Seism. Soc. Am.*, **94**, 2,265–2,279.
- Zhang, H. and C.H. Thurber. (2003). Double-difference tomography: The method and its application to the Hayward fault, California, *Bull. Seism. Soc. Am.*, **93**, 1875–1889.

| Topic   | Assessment of theoretical approaches to seismic network optimization (with PPT Tutorial)                                                                                                                      |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Peter Bormann</b> , formerly at Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Telegrafenberg, D-14473 Potsdam, Germany; now E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version | December 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_IS_8.7">10.2312/GFZ.NMSOP-2_IS_8.7</a>                                                                                                       |

## 1 Introduction

The United Nations Decade on Natural Disaster Reduction (IDNDR; 1990-99) but also the terrible sequence of subsequent devastating earthquakes with often heavy death tolls (Izmit-Düzce, Turkey 1999; Boumerdez, Algeria 2003; Sumatra-Andaman 2004, Pakistan-Kashmir, 2005; Wenchuan, China 2008; Haiti and Chile 2010; Fukushima, Japan 2011) has in a dramatic way increased the world-wide awareness of the tremendous disaster potential of earthquakes and the urgent need for improved seismic hazard assessment, mitigation and preparedness measures. Part of these efforts is the improved monitoring and analysis of seismicity in the countries at risk, which includes improved event detection, location, magnitude determination and assuring catalog completeness down to as small as possible earthquakes (e.g., Woessner and Wiemer, 2005; Schorlemmer and Woessner, 2008; Schorlemmer et al., 2010; Kraft et al., 2011; Mignan et al., 2011).

Accordingly, the number of seismic stations and networks deployed, especially in earthquake-prone developing countries, has grown tremendously. Connected with these often costly investments in hard and software as well as qualified manpower is the expectation to get a maximum of useful information out of these “owned” physical networks, also by taking additionally into account already existing data acquisition and sharing facilities in neighbouring countries and by accessing “virtual” regional and global networks (see Chapter 8 and IS 8.3 of this Manual). In a similar way, this urgent need for fast and best quality data acquisition and information extraction also applies to the International Monitoring System (IMS) of the Comprehensive Nuclear Test-Ban Treaty (CTBT) Organization with its associated national and international components, e.g., for onsite inspection. Accordingly, related publications on the optimization of seismic networks on a local (e.g., Uhrhammer, 1980; Hardt and Scherbaum, 1994; Steinberg and Rabinovitz, 2003), regional (e.g., Kijko, 1977; Steinberg et al., 1995; Kraft et al., 2011) and global scale (e.g., Soriau and Woodhouse, 1985) have been flourishing during the last few decades.

The theoretical basis for such studies is provided by the statistical theory of optimal experimental design (e.g., Kiefer and Wolfowitz, 1960; Kijko, 1977; Silvey, 1980; Rabinovitz and Steinberg, 1990; Steinberg and Rabinovitz, 2003). This theoretical approach aims at maximizing the determinant  $\det(A^T A)$  with  $A$  the  $N \times p$  matrix of partial derivatives of observables with respect to the parameters that characterize the object under investigation, called the *D-criterion*. In the case of event location, e.g., these are the observable  $t$  (travel-

times) with respect to the 4 unknown source parameters latitude ( $\phi$ ), longitude ( $\lambda$ ), source depth  $z$  and origin time  $t_0$ . Then the D-criterion is (det of the information matrix of the hypocenter parameters) and a D-optimal network configuration maximizes the information matrix of these hypocenter parameters.

Travel-time observations may also be used for improving the Earth's velocity model (tomography) or for joint improvement of both event location and velocity structure (see IS 11.1). And other observables allow to derive improved information about other "objects" of interest, e.g., spectral amplitudes of different types of seismic waves for calculating different types of earthquake magnitude, or the 6 components of the seismic moment tensor, or observations of the distribution of P-wave polarities or amplitude ratios of P to S waves in their dependence on source azimuth and distance to derive fault plane solutions. Thus, different combinations of relevant observables with respect to looked-for parameters of the seismic sources, or of the propagation medium such as seismic velocity distribution, inhomogeneity, and anisotropy, or wave attenuation and scattering properties are in principle possible.

In recent years also algorithms of simulated annealing (SA) have been applied for calculating optimization scenarios, e.g., by Hardt and Scherbaum (1995), both for minimizing the volume of the error ellipsoid of the **linearized** earthquake location problem as well as for optimizing network configurations for best source mechanism estimates and velocity tomography studies. SA is a discrete inversion technique, such as genetic algorithms, is fast, easy to program and not restricted to the linearity of the problem. It also allows optimizing of seismic networks for difficult and combined tasks. Kraft et al. (2011), describe SA in more detail. They also extended the algorithm of Hardt and Scherbaum that was restricted to 1D layered velocity models and upgoing direct P and S waves, by

- calculating travel-times of seismic body waves using a finite difference ray tracer and the 3D velocity model of Switzerland;
- calculating seismic body-wave amplitudes at arbitrary stations using the Brune (1970) source model and using scaling relations recently derived for Switzerland, and
- estimating the noise level at arbitrary locations within Switzerland by using a first-order ambient seismic noise model based on satellite-imagery-derived land-use classes and open GIS data.

However, the complexity and degree of non-linearity of the problem increases with the number of observables and parameters to be optimized. Also, in most countries, detailed information about the local and regional velocity structure, topography, signal-to-noise conditions and other crucial boundary conditions are not yet available. This limits the full exploitation of potential benefits, which are offered by the application of more advanced optimization algorithms. Therefore, even rather recent publications, such as Steinberg and Rabinovitz (2003) still apply the simple "layers-above-half-space" model and assume that the layer thicknesses and velocities are known. They refer to Mitchell's (1974) DETMAX algorithm for the problem of finding D-optimal networks when constraints on the use of possible station sites do not permit the direct use of the "pure" theoretical results. They also have implemented these ideas in the software program OPTINET (Shimshoni et al., 1992). Thus they derive first bounds on the efficiency of any proposed network but leave the application of more complex algorithms deliberately for the time when a more detailed model parameterization is available.

This has also been our concept for the complementary PPT tutorial linked to this information sheet.

Note, that the D-criterion is applicable only for linearized design problems. Geoscientific surveys, however, are usually of a more complex non-linear nature. To handle large-scale non-linear optimization problems is still beyond current possibilities, however, respective efforts for small-scale industrial surveys are already promising. E.g., Coles and Curtis (2011) presented a new method, termed  $D_N$  optimization, for a fully non-linear Bayesian statistical experimental design. It is based on a generalization of the D criterion, takes advantage of efficient linear methods and has lower computation and data storage cost than other non-linear algorithms. When comparing their results for an optimal seafloor microseismic sensor network to monitor a fractured petroleum reservoir they could demonstrate reduced spatial bias and hypocenter uncertainty in comparison with the results of a network designed on heuristic principles and of another one based on a linearized Bayesian design. For a more general overview paper on recent advances in optimized geophysical survey design we refer to Maurer et al. (2010). However, dwelling more on non-linear design optimization is beyond the scope of this Information Sheet.

## 2 The optinet.ppt tutorial

The tutorial annexed to this information sheet outlines the very essentials of the two most common approaches to network optimization and presents some basic results derived by applying these algorithms to simplified models and sometimes restrictive boundary conditions. None-the-less, these results are more or less representative for initial network optimization calculations based on still insufficient knowledge of the seismicity, environmental, logistic and other conditions in the considered country or area of network deployment. But most importantly, the PPT aims at demonstrating that the optimization result strongly depends on both the main aim or the multiple tasks of the network as well as on the “ambient” conditions under which the network has to be installed and operated. Accordingly, **one cannot expect a unique multipurpose optimal network solution** from any theoretical approach to network optimization.

However, in order to avoid overloading the PPT presentation about principal optimization goals, algorithms and results with text and comments related to more sophisticated approaches and results we have elaborated on this in the following section of **Comments**. They are numbered in the order as they appear in red as **Comment 1, Comment 2, etc.** on the respective slides of the PPT presentation. These comments aim at qualifying or complementing some of the statements in the PPT presentation and hint to more recent (yet still linearized) optimization procedures and results of their application. Therefore, reading these comments will ease proper understanding and assessment of the PPT presentation.

The PPT tutorial can be opened from here via hyperlink by right mouse click with the cursor set in front of the file name [optinet.ppt](#). The presentation can also be downloaded via the summary listing [Download Programs & files](#) (see Overview on the NMSOP-2 cover page and follow related instructions).

### 3 Comments

#### Comment 1 on slide 4

The idealized assumption of the **first bullet point** is not an inherent requirement of the D-criterion. Rather it has to be qualified depending on the task, environmental conditions, infrastructure, data transmission links, logistic facilities, maintenance requirements, cost and other factors. When the aims and boundary conditions are better known one can “preset” a suitable or affordable number of possible station sites and then run under these refined conditions the search for an D-optimal network.

Also the assumptions made under the **second and third bullet point** are no prerequisites for applying the D-criterion. Rather, they are “set” by the way in which the matrix elements are calculated, e.g., by the algorithm used for the seismic ray-tracing. Kraft et al. (2011) base their search for a D-optimal solution on a 3D velocity model and realistic station heights. But also variable depths of borehole seismometer installations could be handled. Accordingly, also the **warning** at the end of this slide holds only for the case that the algorithm is run under the mentioned simplifying or restrictive conditions.

#### Comment 2 on slide 6

The  $v_p/v_s$  ratio tends to be more stable than the absolute velocities of P and S waves. Therefore, the joint use of P- and S-wave arrivals holds the promise of more reliable hypocentral distance and location estimates. This is usually the case for locations with modest precision requirements. However, in the case of significant variations of the  $v_p/v_s$  ratio (found out, e.g., by constructing Wadati diagrams; see IS 11.1) the inclusion of S-waves into the hypocenter location procedure would require the availability also of an independent good S-wave velocity model. Regrettably, the latter is usually not the case because of the dominating trend in both routine interactive and automatic seismogram analysis to pick only P-wave first motions. This notwithstanding, according to Gomberg et al. (1990), the inclusion of S waves is a powerful tool to significantly improve the hypocenter location accuracy and to reduce the trade-off between source depth and origin time. The most important findings of these authors can be summarized as follows:

- In the absence of detailed knowledge of the regional velocity structure, the most certain way to minimize the solution’s sensitivity to model parameter inaccuracies and theoretical simplifications is to record and pick the onsets of S phases within the aforementioned distance range.
- Using an S phase recorded close to the event almost always improves the accuracy of depth determination. E.g., including at least one reliable S-phase recorded at an epicentral distance  $D$  within about 1.4 times focal depth  $h$  yields depth estimates accurate within approximately 1.5 km. The accuracy will be only slightly inferior when the vertically averaged velocity model assumed in the location procedure differs from the true one by less than a few percent.
- In contrast, when systematic model errors are in the order of 4% and no S phases have been identified close to the source and included into the location procedure then the depth error can be larger than 3 km. This holds even in the case of good azimuthal coverage and the inclusion of several P-phases recorded at  $D < h$ .

- The largest relative improvement in focal depth accuracy is achieved in the case of poor azimuthal station coverage.
- **HOWEVER:** Just because of S-wave readings being potentially so powerful a constraint, systematic S-phase timing errors, already as small ones as 0.2 s, may degrade the accuracy of  $h$  estimates by several kilometres, even if the azimuthal coverage is good. Moreover, such S-wave timing errors may even misleadingly result in a stable - but wrong - solution with reduced standard error, i.e., improved *precision* of the solution although the actual *accuracy* is worse and may even lead to the construction of an incorrect S-wave velocity model.
- Thus, in order to take full advantage of the potential of S wave readings for improving hypocenter solutions one should aim at an optimal station spacing not larger than three times the most shallow earthquake depth which has to be determined accurately, and further, at assuring correct identification and time-picking of the S-wave onset. The latter can be assured best when readings are made on a transverse component record of sufficient dynamic range to avoid clipping.

### Comment 3 on slide 7

Experiments by Kraft et al. (2011) have shown that optimization runs for an N-station network and an N+1 network with free choice of location in both cases still show significant differences in network geometry even in the case of large N. In this sense, the procedure described in the last bullet point on slide 7 is not D-optimal for a “free” N+1 network but only in the sense of adding one station to a given network of already fixed station positions. Yet, in practice, the latter is the usually looked for.

### Comment 4 on slide 8

One major problem, which has great influence on the proposed optimal networks, is the normalization of the used objective cost functions and their relative weight with respect to other parameters. Another problem is the dependence on the random numbers and the high degree of non-linearity of systems with more than 10 stations. This requires careful tests and control of the starting conditions. E.g., Kraft et al. (2011) looked for the optimal network for hypocenter location of events down to magnitude  $M_l = 1$  in northern Switzerland by adding between 10 to 35 new stations to an existing net of 67 stations in Switzerland, Germany and Austria. Repeated optimization runs over a grid net of 952 potential station sites, separated by a spacing of some 4.5 km, yielded fluctuating results of the optimal station sites. Although their positions usually did not differ by more than two grid spacings this hints to inherent non-linearity of the problem. This made it advisable to search within these fluctuation zones of about 10 km diameter for the best suited sites. While the overall hypocentral resolution of the network configurations calculated by several optimization runs are about the same, some larger variations in the actual network geometry are possible. Generally, it is advisable to test the stability of solutions by repeated initialization of the optimization procedure.

**Comment 5 on slide 9**

One basic concept of SA (Kirkpatrick et al., 1983) is to occasionally accept "bad" solutions with  $E(j) > E(j-1)$  depending on the preset and slowly decreasing virtual temperature  $T$  of the system (termed "annealing schedule"). This allows to overrun local minima. The criterion to accept "bad" solutions is:  $\exp(dE/T) \geq \text{rand}(0, 1)$ . The criterion depends on the random number  $\text{rand}(0,1)$  and is more likely fulfilled if the virtual temperature  $T$  is high.

**Comments 6 on slide 10**

Hardt and Scherbaum (1994) considered only 1D-layered or homogeneous velocity models and stations in the distance range of Pg/Sg first arrivals, i.e., within epicentral distances typically less than about 150 km.-As said already in slide 5, for  $v = \text{const.}$  the optimal station distance radius  $r$  around the source would be infinite, because this would assure the smallest relative error in onset-time picking with respect to the total travel time. In reality the dimensions of the network are finite and determined by the scale of the problem, i.e., by the size of the search grid. Therefore, an optimal network geometry in the finite homogeneous case will consist of stations symmetrically distributed on the circumference of the search grid. This is clearly seen in the 4 solutions presented on this slide.

Further, the D-criterion tends to cluster stations at optimal sites, e.g., along the outer circumference of the search grid, if the number of looked-for stations is large with respect to the scale of the problem such as the size of the earthquake source region. This behaviour can be changed by introducing a weighting function for station spacing (e.g., Rabinowitz and Steinberger 1990). Kraft et al (2011) made use of the clustering feature of the D-criterion to identify important sites in an optimal network geometry. They performed repeated optimizations for the same large scale problem with slightly different starting conditions, i.e. re-initialized random numbers. Although the network geometries of individual runs differed due to the inherent non-linearity of the problem, stations of the different solutions were found to cluster at specific locations, which were therefore considered important optimal station sites for the considered problem. For suitable action see Comment 4 and the final comments.

In contrast to the examples shown in slide 10 with event numbers  $N_e$  less or only somewhat larger than the number of looked-for stations sites  $N_s$ , Kraft et al. (2011) searched for the best location of 10 to 35 new stations (to be added to the existing 67 stations) for locating 2240 synthetic catalog earthquakes that were equally distributed over the area under investigation. Although equal distribution of earthquakes is surely not the case in reality, this assumption assures an equally representative assessment of the resolving power of the looked-for network over the hypocentral volume covered by the network and within which earthquakes are likely to occur.

**Comment 7 on slide 15**

See last paragraph of Comments 6. The gain to be expected by later optimization according to the actually observed distribution of earthquakes is considered by T. Kraft (personal communication 2011) as likely being not so large.

**Comment 8 on slide 20**

The results of optimization strongly depend on the assumed ambient noise model. Therefore, Kraft et al. (2011) developed a first order ambient noise model for Switzerland on the basis of land-use-data derived from satellite imagery. Further, they decided to place, in a first implementation step, only part of the recommended new stations at locations with the lowest noise expected according to the three considered noise classes. With actual noise data at these stations, measured over a representative time window, they plan to update the noise model for the remaining required new station sites and to re-evaluate the initial optimization results.

However, it is often overlooked that low ambient noise does not guarantee an improved signal-to-noise ratio (SNR). Low noise hard rock sites often have also reduced signal amplitudes, whereas sites on soft rock with usually higher ambient noise typically show, especially in the short-period range, increased (up to about a factor of 10) signal amplitudes. Moreover, even for given hard-rock site the short-period SNR may, due to lateral velocity and/or topographic inhomogeneities, significantly depend on source distance and azimuth (see Chapter 4, Figs. 4.34-4.36). Investigating such site-specific signal and noise amplification has meanwhile become a widely used method for microzonation studies (see Chapter 14). Yet, these effects have so far rarely been taken into account in network optimization algorithms, although they affect especially network optimization aimed at assuring as low as possible homogeneous “completeness magnitude”  $M_c$ . Therefore, when new information about noise level and SNR conditions become available, even for a small number of the stations, then the optimization procedure should be repeated until the desired hypocentral resolution and detection/magnitude threshold sensitivity of the network is achieved.

For some newer works and results on the variable SNR and magnitude completeness problem see Mignan et al. (2010), Schorlemmer and Woessner (2008), Schorlemmer et al. (2010), Woessner and Wiemer (2005) as well as the final comments.

**Final comments related to slide 21**

With reference to slide 4 it should be mentioned that also the rather advanced Kraft et al. (2011) optimization approach does not account for correlation errors. Therefore, the resulting D-optimal networks often contained stations that were closely spaced with already existing or newly placed stations. The clustering increases with the number of stations to be placed in a network. This redundancy effect, however, can also be interpreted as an up-weighting of important placement regions for improving the hypocentral resolution. If such a cluster placement region contains already a station but with high noise (or bad average SNR) level one should either find within this placement region a better site for re-installation or consider placement of the seismometer in a shallow borehole.

Also the investigations by Kraft et al. (2011) have shown that achieving optimal network design goals for different parameters requires different network size and efforts. While for their regional network it would be sufficient to add only 10 stations to assure an epicentral resolution of 0.5 km, a minimum of 20 new stations is required to achieve a hypocentral resolution of 2 km. And the envisaged constant magnitude completeness of  $M_c = 1$  for an area of several 100 km<sup>2</sup> could only partially be achieved by adding 26 new stations. This



highlights that results from optimization algorithms, when being available already in the planning phase, may help to set more realistic goals of what can reasonably be done and provide useful guidance for final decision making prior to hardware purchases, site selection and installation efforts, depending also on available financial resources.

Although the algorithm by Kraft et al. (2011) had been developed for optimizing a rather large-scale multi-station regional network in Switzerland, it is also applicable to smaller optimization problems, e.g., for small-scale local networks aimed at surveillance of the induced seismicity from geothermal, oil or gas extraction operations, due to high-rise dam impoundment, or for volcano monitoring. Also, the algorithm is especially useful to optimize networks in populated areas with heterogeneous noise conditions, and/or complex velocity structures, as well as for complementing already existing networks. The algorithm is freely available for research and teaching from the author (contact: [toni.kraft@sed.ethz.ch](mailto:toni.kraft@sed.ethz.ch))

Finally, we have to emphasize that practically all network optimization cases discussed so far in this information sheet related to short-period high-gain systems, which are usually deployed for monitoring local and regional seismicity. However, several aspects discussed above for assuring high network performance are irrelevant or less important for long-period broadband and global seismic networks. E.g., some sites with significantly improved SNR in the short-period range and thus signal detection from weak events may show no SNR improvement at all for long-period signals (see Chapter 7, section 7.2.4.4, Fig. 7.33-7.36). Also, whereas SP sensors are rather insensitive to air-pressure, tilt and temperature fluctuations, high-gain broadband sensors requires very careful installations and shielding efforts to minimize these kinds of disturbances. However, some of the task-dependent optimal site positions identified by the algorithms may not permit to assure very stable broadband sensor placement and optimal shielding. Moreover, the optimization of the configuration of global seismic networks has to be made under rather different boundary conditions. On a global scale, earthquakes are even more inhomogeneously distributed, but also the continents and oceans, where the sensors have to be deployed. Also, the climatic, topographic, logistic and “political” environments differ more, as do the research goals when studying the whole planet Earth. This requires to consider many additional boundary conditions for optimization scenarios (e.g., Souriau and Woodhouse, 1985) as well as implementation strategies.

## 4. Final recommendations

Embarking on the optimization of seismic network configuration one should practise a step-wise and problem-/task-oriented approach. Theoretical algorithms on design optimization maybe very helpful in such efforts. Yet, one should accept that their importance lies not so much in the accuracy of their results in absolute terms, which is model-dependent and thus usually still insufficient, but rather in their qualitative value. Even answers to questions such as “Which measures may significantly improve or deteriorate the network performance?”, without giving precise numbers, may provide already useful guidance for sound decision making, as do expert recommendations based on rich seismological experience and sound intuition. The tutorial and the above comments aim at developing this needed problem understanding.

## Acknowledgment

The author owes great thanks to Dr. Toni Kraft for critical reviewing and commenting on the power point presentation and hinting to several most recent publications on network optimization. This led to the decision to complement the original PPT tutorial by a more elaborated introductory text and comment section, which were accepted in a second review.

## References

- Brune, J. N. (1970). Tectonic stress and the spectra of shear waves from earthquakes. *J. Geophys. Res.*, **75**, 4997-5009.
- Coles, D., and Curtis, A. (2011). Efficient nonlinear Bayesian survey design by DN-optimization. *Geophysics*, **76**(2), pp. Q1–Q8; doi: 10.1190/1.3552645.
- Curtis, A. (1999). Optimal experiment design: cross-borehole tomographic examples. *Geophys. J. Int.*, **136**, 637-650.
- Gomberg, J. S., Shedlock, K. M., and Roecker, St. W. (1990). The effect of S-wave arrival times on the accuracy of hypocenter estimation. *Bull. Seism. Soc. Am.*, **80**(6), 1605-1628.
- Hardt, M., and Scherbaum, F. (1994). The design of optimum networks for aftershock recordings. *Geophys. J. Int.*, **117**, 716-726.
- Kiefer, J., and Wolfowitz, J. (1960). The equivalence of two extremum problems. *Canadian J. Mathematics*, **12**, 363-366.
- Kijko, A. (1977). An algorithm for the optimum distribution of a regional seismic network – I. *Pageoph*, **115**, 999-1009.
- Kissling, E., Ellsworth, W. L., Eberhard-Phillips, D., and Kradolfer, U. (1994). Initial reference model in local earthquake tomography. *J. Geophys. Res.*, **99**, 19,635-19,646.
- Kraft, T., Husen, St., Mignan, A., Bethmann, F., and Spada, M. (2011). Optimization of a large-scale microseismic monitoring network in northern Switzerland. Report of the Swiss Seismological Service to NAGRA, July 2011, SED-SPECNet-P040016-15072011-TK, pp. 27.
- Lilwall, R.C., and Francis, T.J.G. (1978). Hypocentral resolution of small scale ocean bottom seismic networks. *Geophys. J. R. Astron. Soc.*, **54**, 721-728.
- Maurer, H., Curtis, A., and Boerner, D. (2010). Recent advances in optimized geophysical survey design. *Geophysics*, **75**(5), pp. 75A177–75A194; doi: 10.1190/1.3484194
- Menke, W. (1989). Geophysical data analysis: Discrete inverse theory. Academic Press, pp. 289.
- Mignan, A., Werner, M. J., Wiemer, S., Chen, C.-C., and Wu, Y.-M. (2011). Bayesian estimation of the spatially varying completeness magnitude of earthquake catalogs. *Bull. Seism. Soc. Am.*, **101**(3), 16 pp.; doi: 10.1785/0120100223.
- Mitchell, T. J. (1974). An algorithm for construction of “D-optimal” experimental design. *Technometrics*, **16**, 203-210.
- Rabinowitz, N., and Steinberg, D. M. (1990). Optimal configuration of a seismographic network: A statistical approach. *Bull. Seism. Soc. Am.*, **80**, 1, 187-196.
- Schorlemmer, D. F., and Woessner, J. (2008). Probability of detecting earthquakes. *Bull. Seism. Soc. Am.*, **98**(5), 2103-2117; doi: 10.1785/0120070105.

- Schorlemmer, D. F., Mele, F., and Marzocchi, W. (2010). A completeness analysis of the National Seismic Network of Italy. *J. Geophys. Res.*, **115**, B04308; doi: 10.1029/2008JB006097.
- Shimsoni, Y., Mizrachi, E., Steinberg, D. M., and Rabinowitz, N. (1992). OPTINET: A computer program for designing the optimal configuration of a seismographic network. *Inst. Petroleum Res. and Geophys.*, Report No 321/47/91(4).
- Silvey, S. D. (1980). *Optimal design*. New York, Chapman and Hall.
- Souriau, A., and Woodhouse, J. H. (1985). A strategy for deploying a seismological network for global studies of earth structure. *Bull. Seism. Soc. Am.*, **75**, 1179-1193.
- Steinberg, D. M., Rabinowitz, N., Shimshoni, Y., and Mizrachi, D. (1995). Configuring a seismograph network for optimal monitoring of fault lines and multiple sources. *Bull. Seism. Soc. Am.*, **85**, 6, 1847-1857.
- Uhrhammer, R. A. (1980). Analysis of small seismographic station networks. *Bull. Seism. Soc. Am.*, **70**, 1369-1379.
- Woessner, J., and Wiemer, St. (2005). Assessing the quality of earthquake catalogues: Estimating the magnitude completeness and its uncertainty. *Bull. Seism. Soc. Am.*, **95**(2), 684-698; doi: 10.1785/0120040007.

# Complementary PPT tutorial to IS 8.7: Assessment of theoretical approaches to seismic network optimization

DOI: 10.2312/GFZ.NMSOP-2\_IS\_8.7

by

**Peter Bormann**

Formerly: *Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences,  
14473 Potsdam, Germany; now: E-mail: [pb65@gmx.net](mailto:pb65@gmx.net)*

## Aim

- During the last decades the installation of seismic networks has been flourishing on local, regional and global scale.
- Efforts are made to assure best possible data products from such major investments.
- Since the 1970s theoretical modelling aimed at seismic network optimization has become fashionable.
- Yet, pretended benefits were often oversold and raised unrealistic expectations of customers.

## Therefore:

- This information sheet outlines some of the common theoretical concepts on which such optimization procedures are based as well as the assumptions that have been made to run the algorithms.
- Oversimplified assumptions or restricting conditions may not be met under real conditions.
- Moreover, optimal determination of **different** seismological parameters require **different** network configurations. Therefore, no single solution can satisfy the usually multi-purpose aim of seismic networks. ➔

- Yet, if one does not expect correct all-optimal solutions from theoretical network modelling in absolute terms but rather some guidance in a more qualitative sense for specific tasks, then such solutions may be useful.
- Two examples for specific tasks are given. For high quality location performance see also criteria outlined in IS 8.5 and 8.6. For publications, more elaborated approaches and results on network optimization see **introduction**, **comments**, and **references** in the text to this information sheet.

## Topics of this tutorial PPT:

1. Optimal configuration of seismic networks for location
2. Design of multi-task optimum networks for aftershock recordings
3. Network optimization with respect to main seismogenic faults
4. Detectability and EQ location accuracy modelling of seismic networks
5. Summary conclusions and recommendations for practical network optimization

# 1. Optimal configuration of seismic networks for location

**Approach A:** Hypocenter/epicenter positions are given  
⇒ search for **optimal siting of stations**

**Approach B:** Studying the **potential of a given network** with respect to variations in epicenter/hypocenter locations, noise conditions, addition or deletion of seismic stations

## 1.1 Statistical theory of optimal experimental design

(Kiefer and Wolfowitz, 1960; Kijko, 1977; Silvey, 1980; Rabinovitz & Steinberg, 1990)

The **D-criterion** ⇒ **maximizing  $\det(A^T A)$**  with **A** the  **$N \cdot p$  matrix** of partial derivatives which are, **in the case of event location**, the observables **t** (travel time) with respect to the **p** components of the unknown source parameters  $\varphi$ ,  $\lambda$ ,  $z$  and origin time  $t_o$

⇒ **Then the D-criterion is  $\sim \det$  of the information matrix** of the hypocenter parameters

**Thus, a D-optimal network configuration maximizes the information matrix of the hypocenter parameters.**

## Offered D-optimal solutions are often based on too simple assumptions

(yet most are not required by the D-criterion itself, but rather **aim at simplifying the algorithm**)

- e.g.:**
- stations can be placed at any location (*i.e., free choice, although it depends on the task as well as ambient and logistic conditions*)
  - stations are placed at the same height on the surface (*not suitable in terrains with rough topography*)
  - validity of a standard layered **1D velocity model** above the half-space (*although lateral velocity inhomogeneities and/or anisotropy may be significant*)
  - model errors are
    - a) uncorrelated or
    - b) correlated in a known way at closely spaced stations
- (although model inaccuracies and local anomalies are perfectly correlated at the same site and closely correlated for nearby stations)*

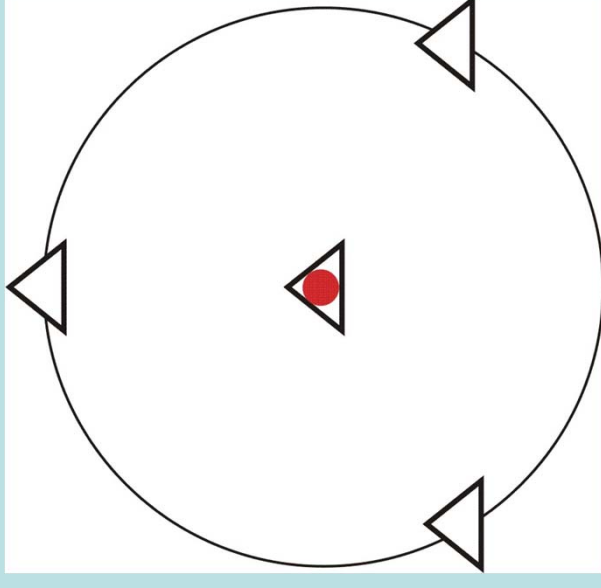
⇒ Design criterion is to maximize  **$\det(\mathbf{A}^T \mathbf{W}^{-1} \mathbf{A})$**  where  **$\mathbf{W}$**  is the **error correlation matrix**

**Warning:** **Such restrictive or idealized assumptions may severely limit the practical value of theoretically „optimized“ configurations**

## 1.2 Results of modelling according to the D-criterion

(for sources in the half-space)

- For a **given epicentre** • the **stations**  $\Delta$  of a D-optimal network are:
  - symmetric about the epicentre;
  - placed **equidistant on concentric circles** in agreement with „**seismological intuition**“.
- For a **given hypocentre in a half-space** the
  - triangular **quadrupartite (4-station) network**  
(according to Lilwall & Francis, 1978, and Uhrhammer, 1980)  
**is D-optimal**;
  - The **optimal radius depends on the velocity model** since the D-criterion is proportional to the take-off angle of the seismic ray from the hypocentre to the station;
  - Therefore, the radius of the „station circle“  $r \rightarrow \infty$  for  $v = \text{const.}$





- For a **hypocentre in a layer above the half-space** (single layer crust) the D-optimal configuration consists of

- **A center station** (*although now less important than in the first case, because of...*)

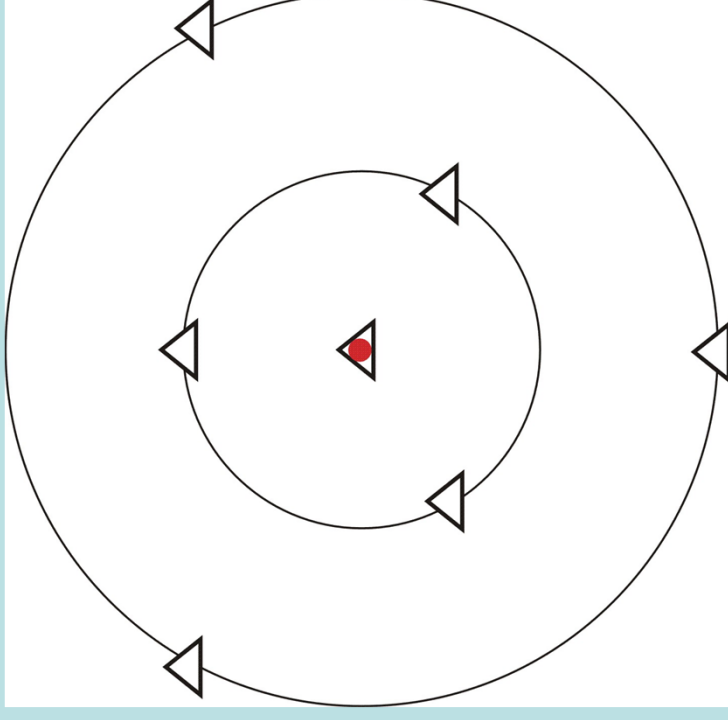
- **Two concentric rings** with three equidistant stations each

- The radii of the two rings have to be chosen so that the

**direct (e.g., Pg) waves** ( $\partial t/\partial z$  positive!) arrive at stations of the **inner ring** and **critically refracted waves** (e.g., Pn) ( $\partial t/\partial z$  negative!) at stations of the **outer ring**.

- Then estimates of depth  $z$  will be uncorrelated with estimates of origin time  $t_0$ .  
➔ **best estimates of source depth!**

- **Additional observations of S waves may substantially improve location** estimates, especially for sub-optimal configurations and for events outside of the network, **provided that** the ratio  $v_p/v_s$  is constant or a good  $v_s$  model is available (*which is, however, usually not the case because of the regrettably dominating routine practice of P-wave first motion picks only; see Gomberg et al., 1990*). **Comment 2!**



## 1.3 Conclusions

- When using only direct P waves, the high correlation between  $z$  and  $t_0$  **may** lead to an **ill-conditioned problem** and large uncertainty in the hypocenter estimate if the number of stations is insufficient and the network geometry bad.
- Networks with **mixed wave arrivals** (e.g., Pg and Pn) **have superior resolution**, in particular for focal depth  $z$ , since  $\partial t/\partial z$  has different signs for the upgoing direct Pg and the downgoing critically refracted Pn waves.
- This allows to design networks so that estimates of depth  $z$  will be uncorrelated with estimates of origin time  $t_0$ .
- **Additional observations of S waves may substantially improve location** under the conditions mentioned before and discussed in detail by Gombert et al. (1990).
- Adding just **one station** to an existing N-station network may already significantly improve its performance. For more detailed discussion on this issue, however, see IS 8.7 text and **Comment 3!**

## 2. Design of multi-task optimum networks for aftershock recordings

**Approach:** **Simulated annealing** (as applied by Hardt & Scherbaum, 1994)

**Prerequisite:** List of **M possible stations locations** is available (with  $M > N$ ; N-number of selected station locations)

**Challenge:** Search for **multi-task optimal network** (e.g., for EQ location, source mechanism, tomography, assuming site specific noise levels, dynamic range, frequency band of recording systems, etc.)

**Advantages:**

- **Deployment after mainshock**, i.e., the likely **area of aftershocks** is  $\pm$  **known** and thus the network aperture can be chosen such that the aftershocks falls **within the network**.
- Only local distance range with  $P_g$  and  $S_g$  as first arrivals considered.
- **Prior event scenario modelling may help (?)** to select the most appropriate design for **rapid deployment** after the event and to **upgrade** the network design in response to the actual development of the aftershock distribution.

**Difficulties:**

- The fine tuning of the annealing parameters **requires experience**.
- Multi-task **optimization is not a trivial problem** and not yet satisfactorily solved.

- The normalization of **cost functions**, their **weighting**, the high **non-linearity** of systems with more than 10 stations (→ see **Comment 4!**)

## 2.1 Method of simulated annealing (SA)

- SA is a discrete inversion technique such as genetic algorithms.
- SA is fast, easy to program and not restricted to the linearity of the problem.
- SA allows **optimizing** of seismic networks for **difficult tasks**.
- SA originates from statistical mechanics and **describes the way a liquid freezes** and forms crystals; if cooling is slow enough, nature **finds a minimum energy state**
- In seismological application **ENERGY** is replaced by the **OBJECTIVE FUNCTION (OF)** which describes the **PERFORMANCE** of a seismic network.
- Thus, the process of **SA can be applied to the problem of finding an optimum network configuration: a) for a given task; b) for a combination of tasks.**
- SA starts from a randomly chosen network configuration selected out of a list of **Np possible sites for Ns stations with  $N_p \gg N_s$** .  
(However, the **Np possible sites are usually not known before the deployment of a net but assumed, e.g., to form a dense regular grid, as in the following demonstration!**)
- For this **OF is computed**, i.e., the ‘energy state’ of the **starting configuration  $E_0$** .
- Then **step by step new configurations** are produced by moving randomly one **station** to another site.
- **New configurations** are accepted if  $dE_i = E_i - E_0 < 0$ , **until  $E_{\min} \ll E_0$  is found**.  
Nos is then the number of optimally distributed station sites
- SA accepts also **conditions** such as  $E(i) > E(i-1) \rightarrow$  **avoid getting trapped in local minima**  $\rightarrow$  see **Comment 5!**

## 2.2 SA optimal networks for hypocenter location **within** network

(Plot examples from Hardt and Scherbaum (1994) were color-coded for eased recognition)

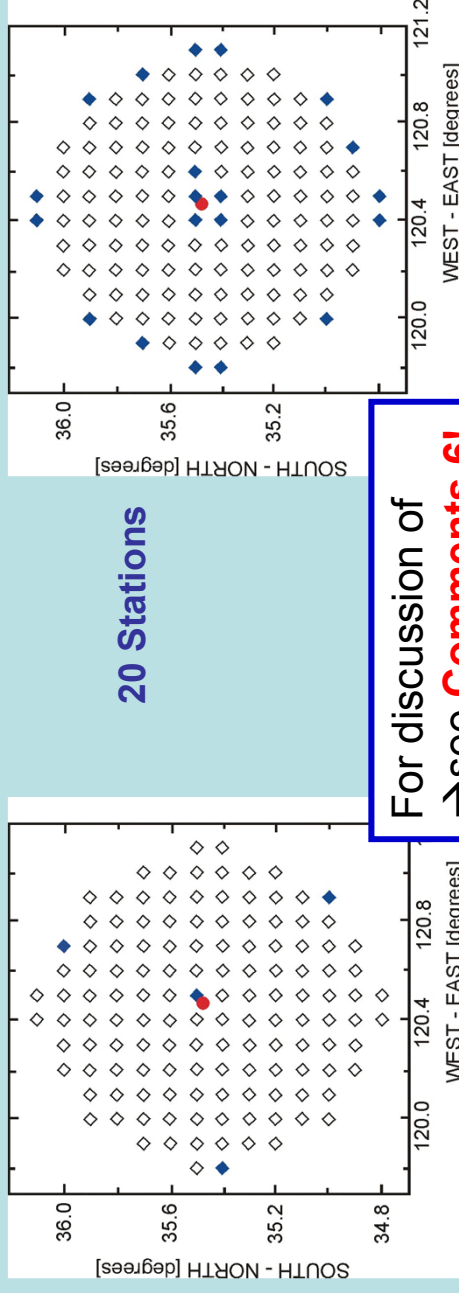
**Red dots:** number **Ne** of assumed event epicenters; **open diamonds:** number **Np** of assumed possible station sites; **blue diamonds:** number **Nos** of calculated optimal station sites; **Ns** = number of available station sites.

**Single event**      4 Stations

- Results as for D-criterion

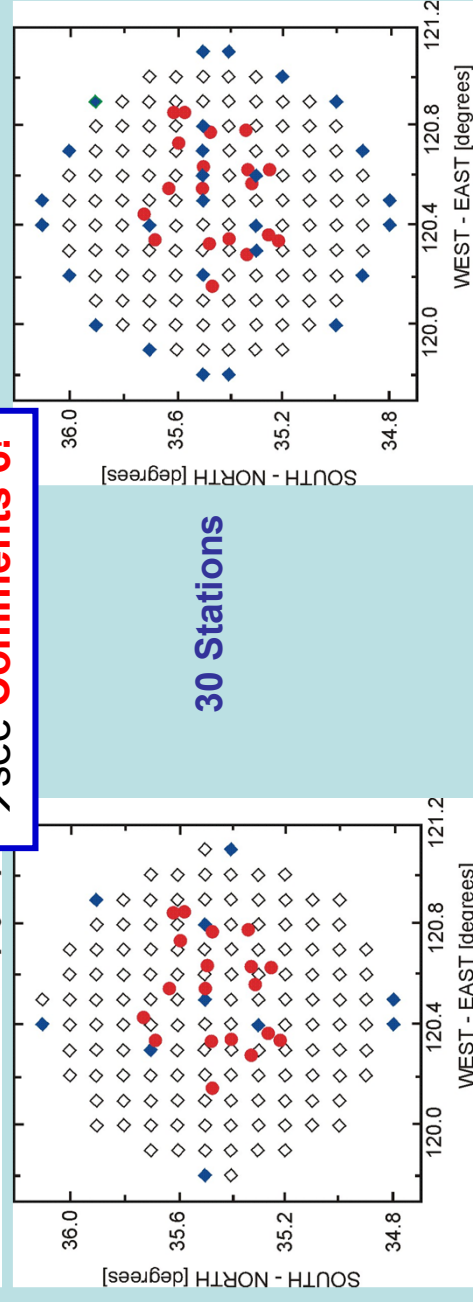
→ **quadrupartite network**

(forms an equal-sided triangle  
with one station in the centre)



For discussion of  
→ see **Comments 6!**

**20 events**      10 Stations



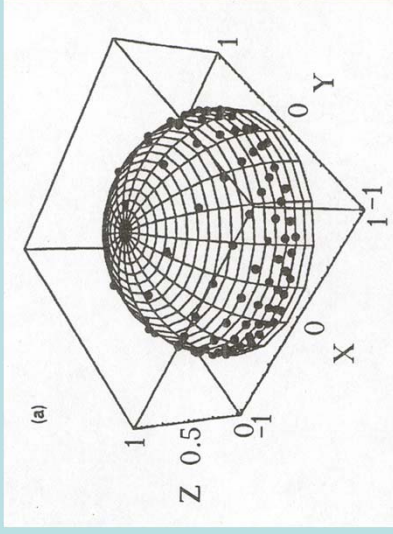
**Conclusions:** a) If event number **Ne** << **Ns** ⇒ highly redundant optimal sites;  
b) Only networks for **Ne** >> **Ns** are of interest; c) Yet: The distribution of the **Nos**  
optimal station sites depends on the **event distribution within the network!**



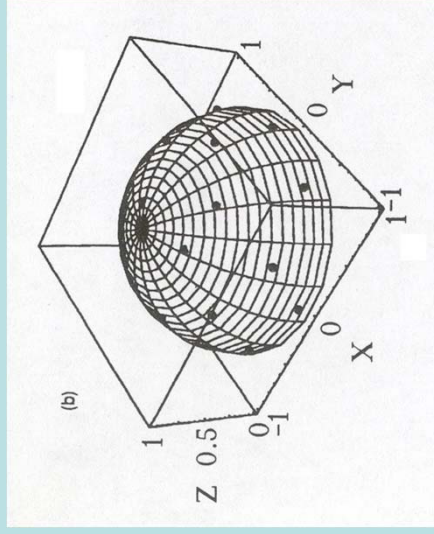
## 2.3 SA optimal network for determination of source mechanism

(Plot examples from Hardt and Scherbaum (1994) were color-coded for eased recognition)

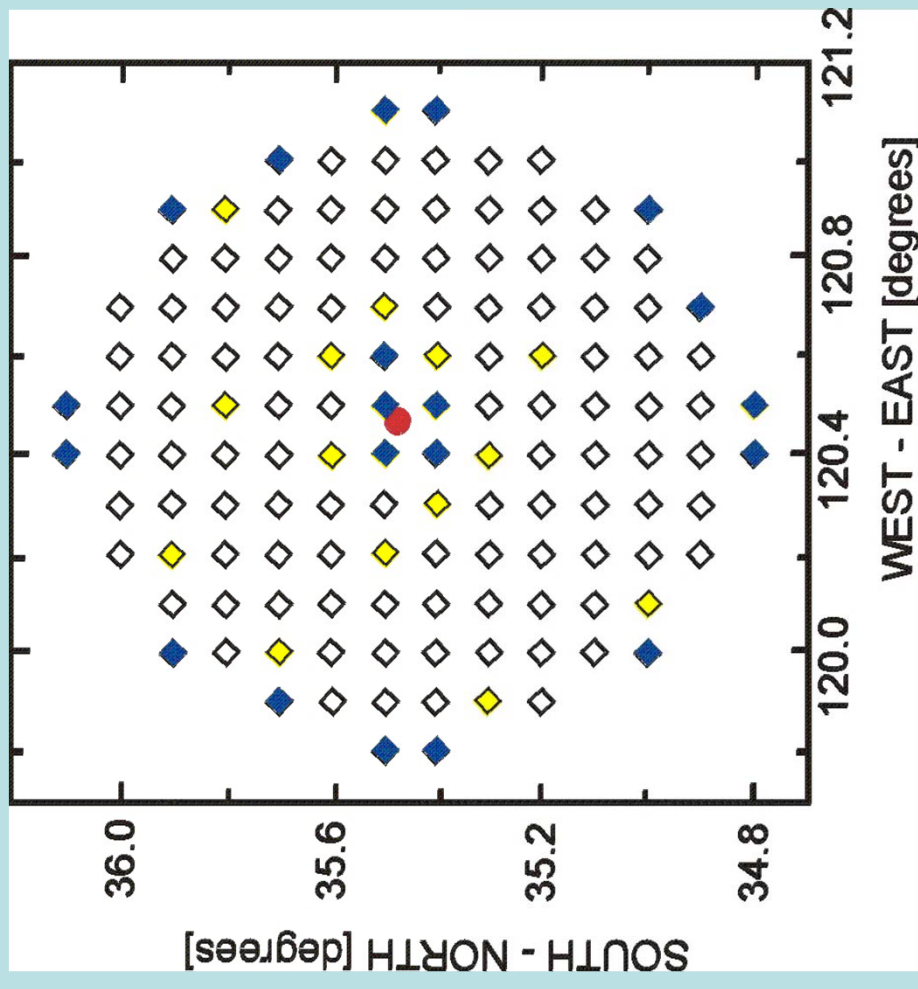
- Ray penetration points on the focal hemisphere of a **single event** in the network centre to:



all 132 assumed possible stations;



20 optimum stations, which record seismic rays that homogeneously sampled the focal sphere.



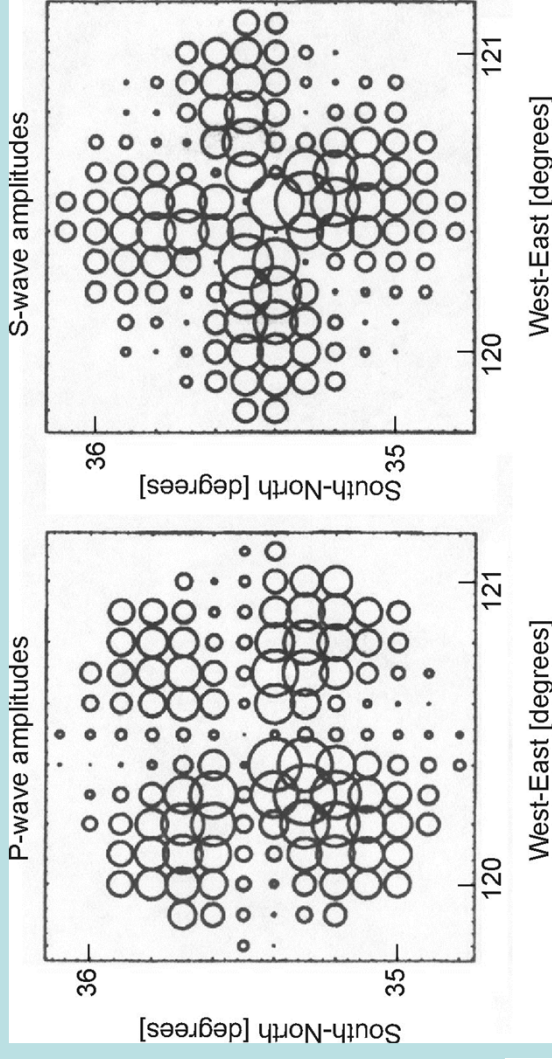
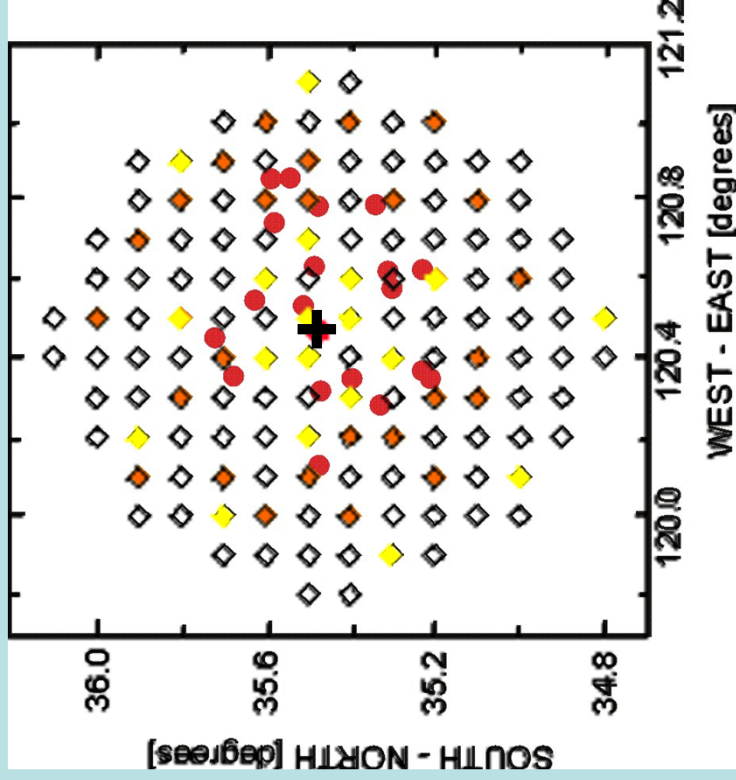
Optimum 20 station network for **source mechanism** determination of a single event in the network center.

Overlay: Optimum 20 station network for **location** of this event

**Conclusion: Optimal networks for location are not optimal for source mechanism!**

## However:

- P and S waves have different radiation pattern
- Amplitude distribution within the network varies also with the
  - spatial orientation of rupture
  - source kinematics/directivity
  - source depth and velocity structure



Azimuthal **amplitude radiation pattern** for P- and S-waves from a pure vertically dipping and/or N-S/E-W trending **strike-slip fault**.

**Note:** The pattern varies with type of **source mechanism** and the specifics of rupture kinematics (slip direction, rupture speed → **source directivity**).

## Moreover:

- Most events occur outside of the network center!

Optimum **30 station network** for source mechanism

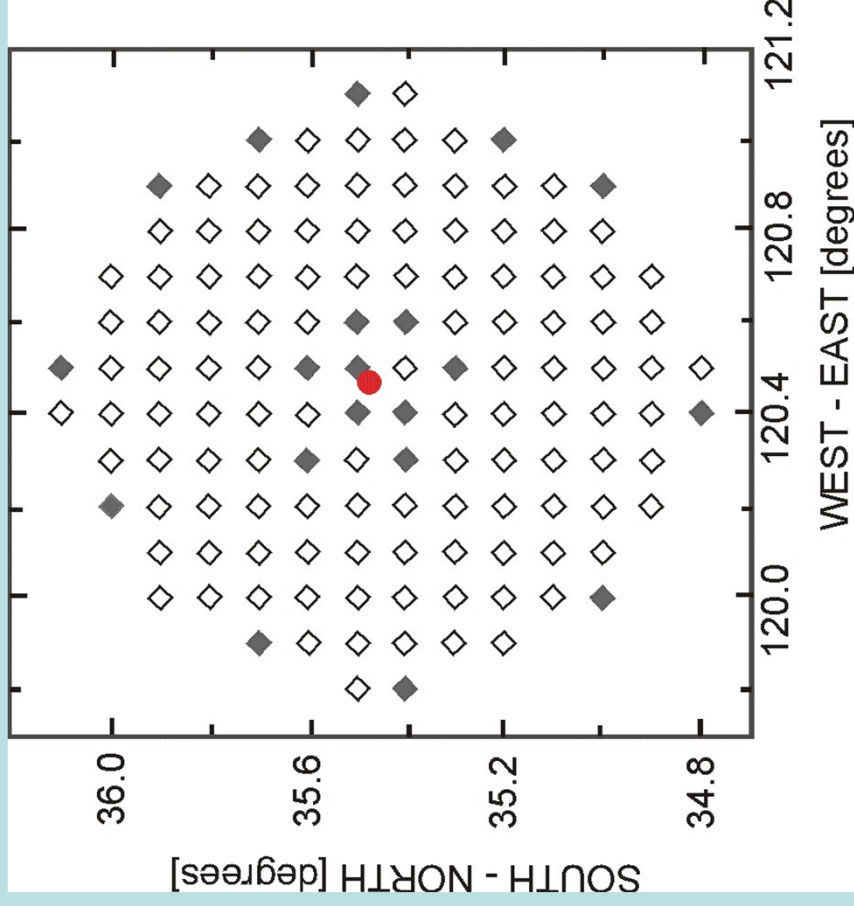
determination of **20 distributed events** •

**Overlay:** Optimum **20 station network** for source mechanism of one EQ + in network center

**Conclusion:** There is no optimal station distribution for the determination of source mechanism from distributed events with different source mechanisms!

## 2.4 Combined SA optimal network for location and source mechanism

(Plot example taken from Hardt and Scherbaum, 1994; color-coded for eased recognition)



Optimum **20 station network** for both event location and source mechanism of a **single event** • in the network center.

- This distribution is similar to a location optimal network for a single event in the network center.

**However, the solution depends on:**

- the weighting factor given to hypocenter location and source mechanism reliability, respectively;
- the velocity structure (especially with respect to the clustering near the network center for improved source mechanism;

**and**

- optimizing for many events distributed within the net changes again the configuration of “optimal sites”.



## 2.5 SA optimal network configuration for tomography

(3-D velocity and attenuation structure)

- Quality measure for tomography is the **model resolution matrix MRM** (Menke, 1989)
- Optimum resolution is obtained if the **MRM** is **diagonal**
- As **OF** one uses a **measure of diagonality** of the **MRM**, scaled between
  - 0 – for poor resolution
  - 1 – for optimum resolution
- The calculation of the MRM is rather time consuming
- **For an optimal net of 30 stations for 20 events**  
Hardt & Scherbaum (1994) calculated that this network would have  
**only < 30% optimality for tomography!**

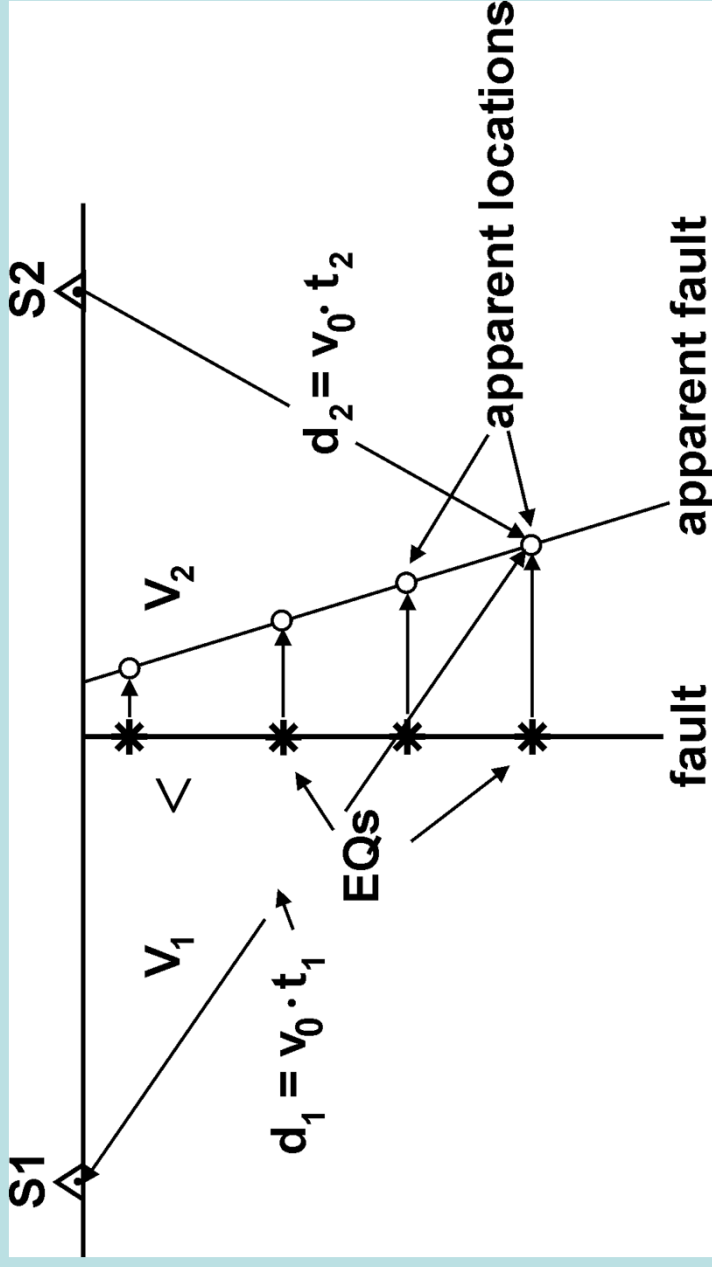
## 2.6 Conclusions from applying SA for network optimization

- Optimized networks for **LOCATION**  $\neq$  **SOURCE MECHANISM**  $\neq$  **TOMOGRAPHY**  
 $\neq$  ???

- Hardly one knows prior to an EQ the list of **potential and accessible** sites in its future aftershock area. The location and size of this area strongly depends on the unpredictable accurate location and size of the future rupture.
- **In reality** the number of deployed stations **Ns**  $\ll$  **Ne** (number of aftershocks) and in the wake of a disaster one can hardly access predetermined “ideal locations”.
- **Aftershocks are not at all randomly distributed** within a network (but see also qualifying **Comment 7**) yet will often occur :
  - near the borders or even outside of it;
  - within an  $\pm$  elongated source volume, which is **closely related to the main rupture** and its orientation in space, and to adjacent fault systems activated due to stress redistribution in the surroundings of the main rupture.
- **Therefore**, it is more promising to **optimize location after a tomographic study** of the 3D velocity inhomogeneities in the aftershock area has been made.

## Principle of mislocation due to lateral velocity inhomogeneities

(Figure 13 of IS 11.1)

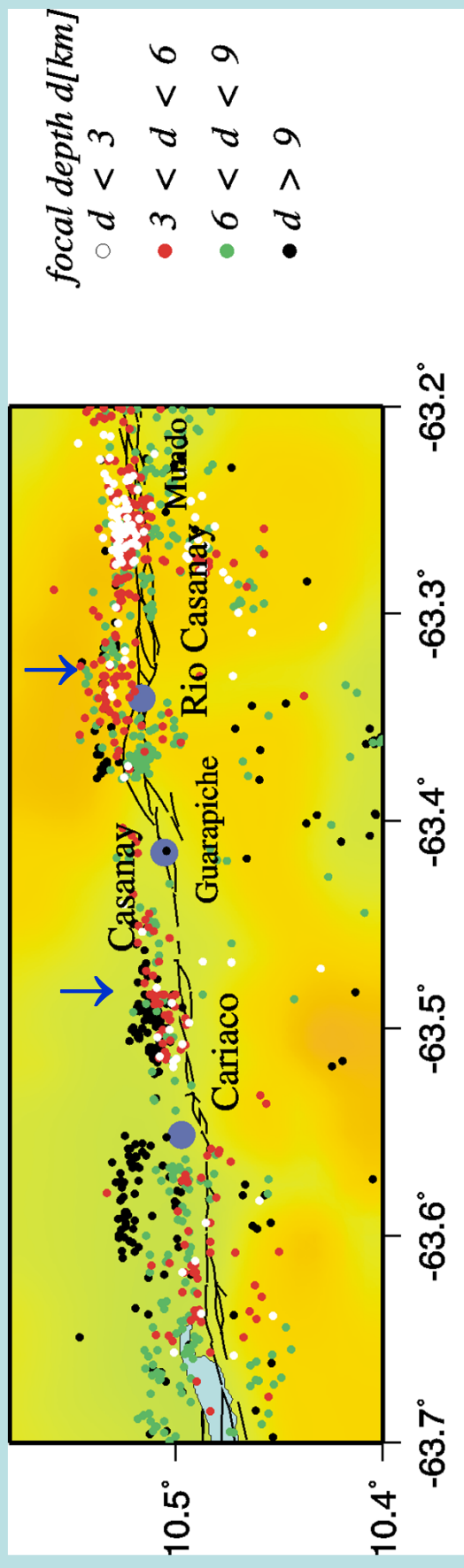


**Cartoon** illustrating systematic mislocation of earthquakes along a fault with strong lateral velocity contrast.  $v_0$  is the assumed model velocity with  $v_2 > v_0 > v_1$ .

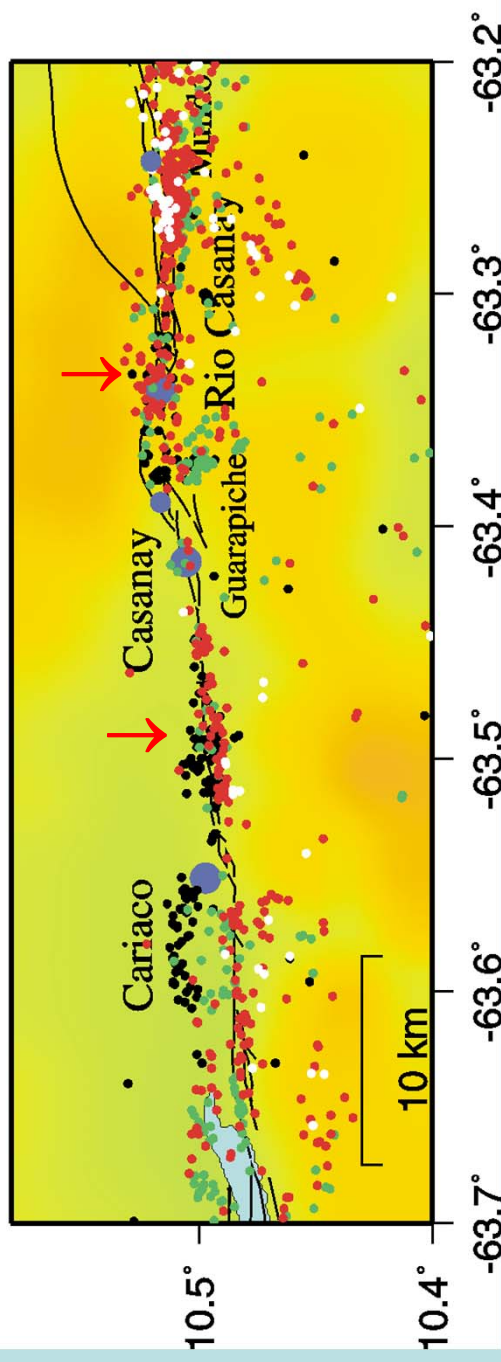
**Note:** Events located in an inhomogeneous medium by using a laterally homogeneous velocity model are always shifted into the direction of higher velocities, the more the less homogeneous the azimuthal distribution of stations used in the location procedure.

## REAL CASE STUDY

### Epicenter maps of aftershocks of the 1997 Ms6.8 Cariaco earthquake, NE Venezuela.

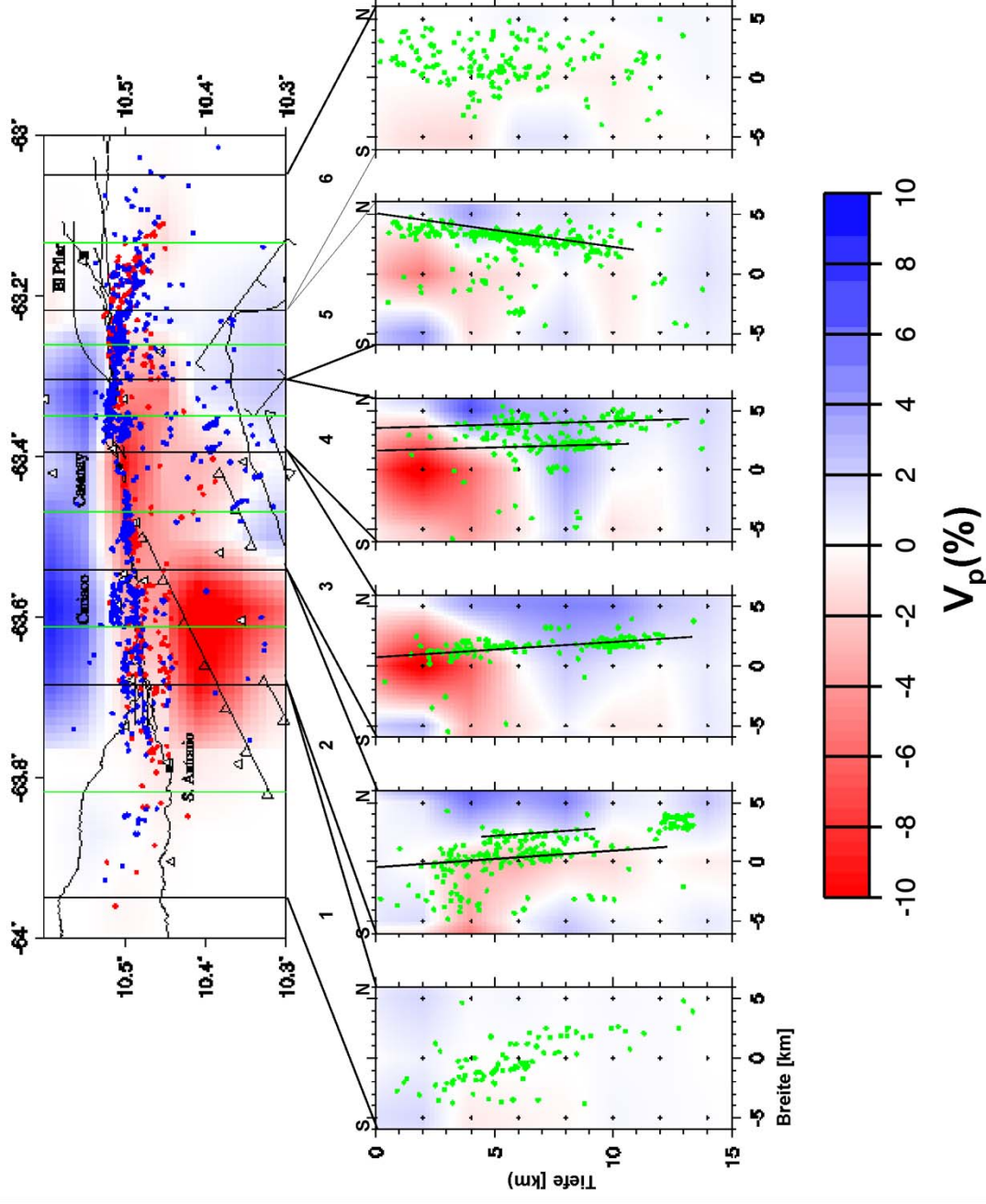


Top: Results from HYPO71 locations based on a 1-D velocity-depth model.



Note the shift  
of epicenters  
from North ↓  
*before*  
towards South ↓  
*after*  
**3-D relocation**

Bottom: Relocation of the aftershocks on the basis of a 3-D model derived from a **tomographic study** →  
of the aftershock region (Figure 11 of IS 11.1, courtesy of M. Baumbach, H. Grosser and A. Rietbrock, 2001).



**Note** the significantly higher P-wave velocities towards North !

**Note** the excellent linear alignment of most aftershocks with depth along the Cariaco fault system in the center area of good tomographic control and resolution and the wide scatter of hypocenter locations outside this area of good network coverage. (where deviations of velocity from the average model can not be resolved).

**3-D distribution of the P-wave velocity** in the focal region of the 1997 Cariaco earthquake as derived from a tomographic study. The map view shows the velocity distribution in the layer between 2 km and 4 km depth. Red and blue dots mark the epicenters of the aftershocks. The red ones were chosen because of their suitability for the tomography.

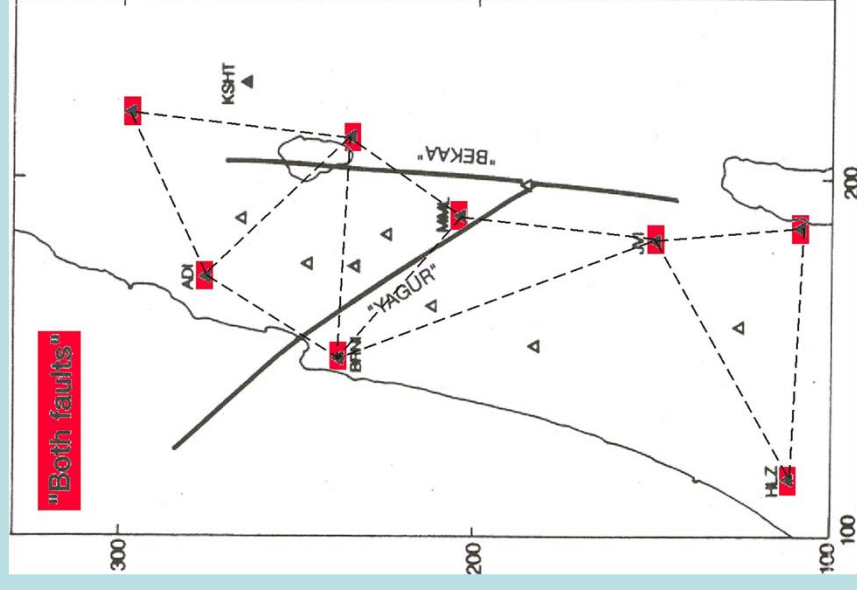
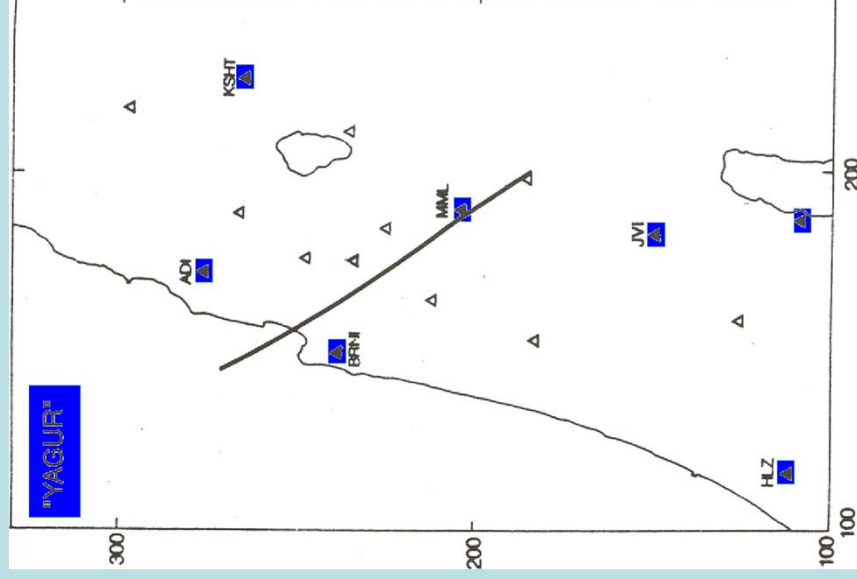
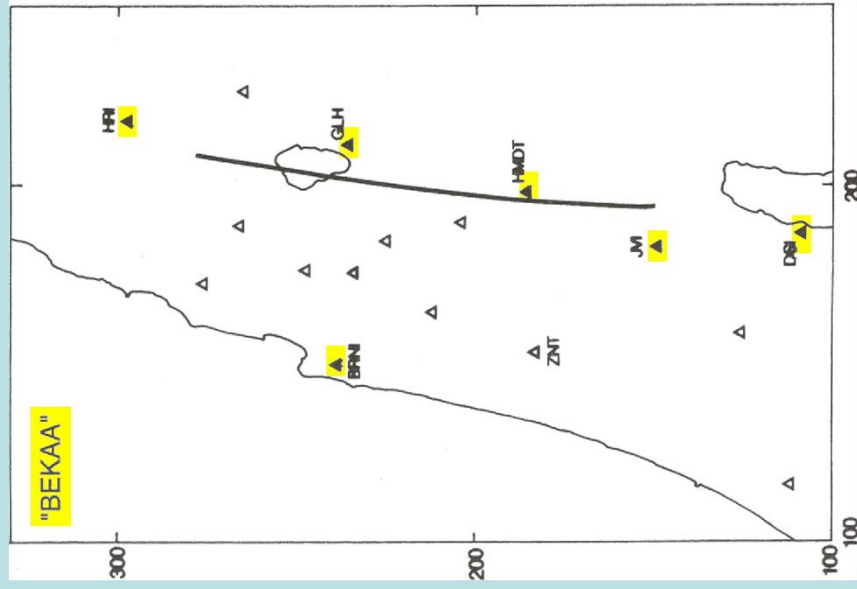
**Vertical cross sections** showing the depth distribution of the aftershocks (green dots) together with the deviations of the P-wave velocity from the average reference model. The depth range and the lateral changes of fault dip are obvious (Figure 12 of IS 11.1, courtesy of M. Baumbach, H. Grosser and A. Rietbrock, 2001).



### 3. Network optimization with respect to main seismogenic faults

(Case study by Steinberg et al., 1995)

- Tasks:**
- Reduction of an existing network from 17 stations to 6-8 station sites
  - Optimal 6-8 stations for monitoring the two main faults “Bekaa” and “Yagur”



Optimal six-station network (out of 17 existing stations) for monitoring the “Bekaa” fault (**yellow**) and “Yagur” fault (**blue**), as well as of an optimal eight-station network monitoring both fault systems (**red**) (according to Steinberg et al., 1995)

## 4. Detectability and EQ location accuracy modeling of seismic networks

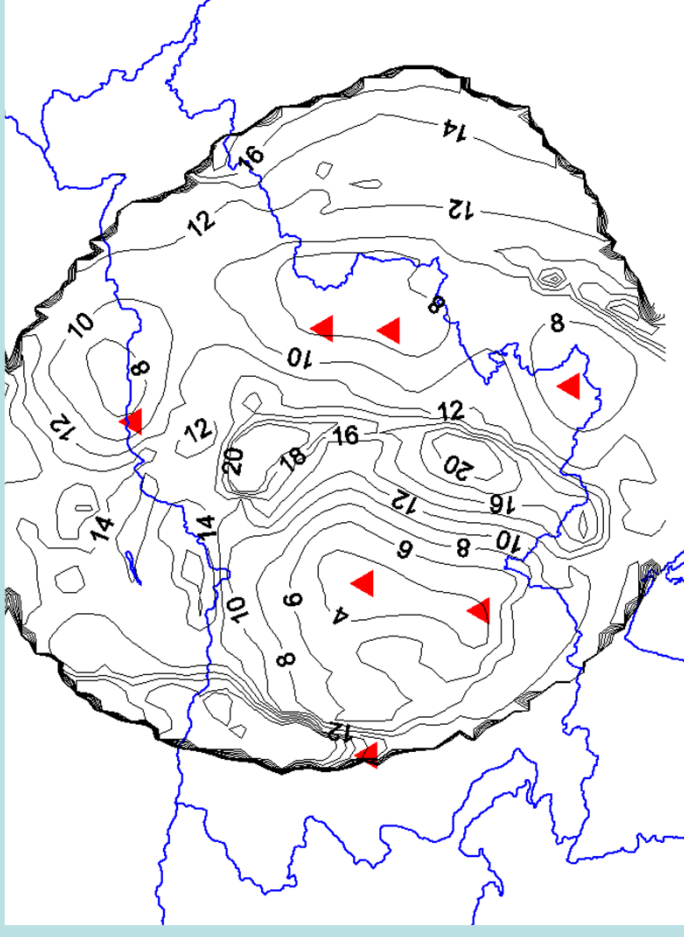
according to Živčić and Ravnik (see IS 7.4)

### Allows to calculate accuracy of event location:

- within and nearby outside of network;
- for different hypocenter depth and EQ magnitudes (i.e.,  $\neq$  SNR and accuracy of time picking; see also **Comment 8**).
- to recalculate network performance when configuration and/or number of station is changed.

### Assumptions:

- Station coordinates are exactly known;
- RMS value of noise in the used frequency range is known;
- instrument response flat between 1 - 10 Hz;
- P- and S- arrival times picked with known accuracy;
- P and S velocity model known (with some uncertainty);
- small local network which allows to use the flat Earth approximation.



Result of model calculations for the Stareslo network in Slovenia for  $M_l = 1.0$  earthquake.

Red triangles: station positions

Blue lines: borders of Slovenia,

Thin black lines: isolines of **hypocenter location error in km**;

Thick black line outer boundary: outer limit of the network's location capability for earthquakes of  $M_l = 1.0$ .

The **computer program LOK** by Živčić and Ravnik is described in more detail in IS 7.4 and can be downloaded from there.

## 5. Summary conclusions and recommendations for network optimization

1. **Theoretically optimal network geometries** obtained under idealized assumptions or restricting conditions **may not be optimal under real conditions** of

- spatially distributed,  $\pm$  irregular seismicity patterns;
- significant variations in velocity structure;
- strongly varying SNR conditions even at “optimal sites” (see **Comment 8!**);
- unaccessibility or other limitations of “optimal sites” due to lacking infrastructure and power supply, rough topography, bad site geology, data-link problems etc.

2. Station distributions should provide a **good azimuthal coverage** with gaps  $\leq 120^\circ$  and **nearly equidistant** spacing forming  $\pm$  equal-sided station triangles.

3. **Good depth estimates** require at least **one station near epicentre** (at  $D < 2$  h), or **stations both in the Pg and Pn distance range, or well identified depth phases.**

**THEREFORE: priority should be given to:**

- **Meeting conditions 2. and 3.** together with comparably **good detectability thresholds** (SNR) at the various stations/subnets as far as possible;

- **later object and/or problem-oriented optimization of existing networks** by addition or deletion of stations based on better knowledge of **seismicity patterns, velocity inhomogeneities, SNR, chief aim of study, financing** etc. (**see Final Comments**).



# Chapter 9

## Seismic Arrays

(Version December 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch9)

Johannes Schweitzer, Jan Fyen, Svein Mykkeltveit,  
Steven J. Gibbons, Myrto Pirli, Daniela Kühn, and Tormod Kværna<sup>1)</sup>

<sup>1)</sup> All: NORSAR, Gunnar Randers vei 15, P.O. Box 53, N-2027 Kjeller, Norway; contact via:  
Phone: +47-63805900, E-mail: [johannes.schweitzer@norsar.no](mailto:johannes.schweitzer@norsar.no)

|                                                                                   | page |
|-----------------------------------------------------------------------------------|------|
| <b>9.1 Outline</b>                                                                | 2    |
| <b>9.2 Introduction</b>                                                           | 3    |
| <b>9.3 Examples of seismic arrays</b>                                             | 5    |
| <b>9.4 Beamforming</b>                                                            | 9    |
| 9.4.1 Geometrical parameters                                                      | 11   |
| 9.4.2 Apparent velocity and slowness                                              | 13   |
| 9.4.3 Plane-wave time delays for sites in the same horizontal plane               | 15   |
| 9.4.4 Plane-wave time delays when including site elevations                       | 16   |
| 9.4.5 Beamforming formulae                                                        | 18   |
| 9.4.6 Examples of beamforming                                                     | 19   |
| <b>9.5 Data analysis with beamforming</b>                                         | 21   |
| 9.5.1 The n-th root process                                                       | 21   |
| 9.5.2 Weighted stack methods                                                      | 21   |
| 9.5.3 The double beam technique                                                   | 22   |
| 9.5.4 Time delay corrections                                                      | 22   |
| <b>9.6 Beamforming and detection processing</b>                                   | 23   |
| <b>9.7 Array transfer function</b>                                                | 28   |
| <b>9.8 Slowness estimation using seismic arrays</b>                               | 31   |
| 9.8.1 Slowness estimation by f-k analysis                                         | 32   |
| 9.8.2 Beampacking (time domain wavenumber analysis)                               | 36   |
| 9.5.3 Plane wave fitting - slowness estimation by time picks                      | 37   |
| 9.8.4 The VESPA algorithm                                                         | 38   |
| 9.8.5 The correlation method used at the UKAEA arrays                             | 40   |
| 9.8.6 Tracking earthquake source propagation with seismic arrays                  | 40   |
| 9.8.7 Slowness corrections                                                        | 41   |
| <b>9.9 Detection of repeating seismic events using waveform cross-correlation</b> | 43   |
| <b>9.10 Ambient noise array analysis</b>                                          | 45   |
| 9.10.1 Overview                                                                   | 45   |
| 9.10.2 Single station/array measurements                                          | 45   |
| 9.10.3 Array geometry for ambient vibration measurements                          | 46   |
| 9.10.4 Estimation of phase velocity curves                                        | 46   |
| 9.10.5 Inversion of dispersion curves to obtain shear wave velocity profiles      | 48   |
| 9.10.6 Difficulties                                                               | 48   |
| <b>9.11 Array design for the purpose of maximizing the SNR gain</b>               | 50   |
| 9.11.1 The gain formula                                                           | 50   |
| 9.11.2 Collection of correlation data during site surveys                         | 51   |

|             |                                                                  |           |
|-------------|------------------------------------------------------------------|-----------|
| 9.11.3      | Correlation curves derived from experimental data                | 52        |
| <b>9.12</b> | <b>Considerations when planning a new seismic array</b>          | <b>54</b> |
| 9.12.1      | Basics                                                           | 54        |
| 9.12.2      | Example: A possible design strategy for a 9-element array        | 55        |
| <b>9.13</b> | <b>Routine processing of array data at NORSAR</b>                | <b>56</b> |
| 9.13.1      | Introduction                                                     | 56        |
| 9.13.2      | Detection Processing – DP                                        | 59        |
| 9.13.3      | Signal Attribute Processing – SAP                                | 60        |
| 9.13.4      | Event Processing – EP                                            | 63        |
| 9.13.5      | Event location with a network of arrays                          | 65        |
| 9.13.6      | Teleseismic event location                                       | 65        |
| <b>9.14</b> | <b>Seismic arrays in Earthquake Early Warning Systems (EEWS)</b> | <b>66</b> |
| 9.14.1      | Introduction                                                     | 66        |
| 9.14.2      | Examples for array contributions to fast event locations         | 66        |
| 9.14.2.1    | Single array results                                             | 66        |
| 9.14.2.2    | The Fast Earthquake Information Service (FEIS) algorithm         | 67        |
| 9.14.2.3    | NORSAR’s Event Warning System (NEWS)                             | 67        |
| 9.14.3      | Usage of seismic arrays to monitor an EEWS relevant site         | 68        |
| 9.14.3.1    | The single array case                                            | 69        |
| 9.14.3.2    | A single array and a sparse regional network                     | 69        |
| 9.14.3.3    | Multiple array configuration                                     | 69        |
|             | <b>Acknowledgments</b>                                           | <b>70</b> |
|             | <b>References</b>                                                | <b>70</b> |

## 9.1 Outline

When the Manual of Seismological Observatory Practice (Willmore, 1979) was published, only a small number of seismic arrays with open data access were in operation. At the time, array seismology was limited to a very small number of specialists and, as a result, the chapter on array seismology in that issue of the Manual comprised just two pages, including one figure. Over the last three decades, many new seismic arrays have been deployed globally and, due to improvements in digital data acquisition systems and digital signal processing, it has become much easier to handle the large amount of data generated by seismic arrays. Array data have also become more easily accessible, and some array data are now even available in near real-time via SEEDlink servers. Consequently, array observations have now become more common in the seismological community. This chapter on array seismology explains the principles of seismic arrays and how array data can be used to analyze different types of seismic signals.

In the following sections, we define the term “seismic array” and show examples of seismic arrays installed around the world. We then describe the theoretical basics of the processing of seismic data observed with an array, continue with the explanation of tools for the analysis of transient signals and ambient noise observed with seismic arrays and describe some considerations in the optimization of seismic array configurations, in order to provide guidance for new array installations. We explain how seismic, in particular local and regional, events are located at the NORSAR Data Processing Center using only single array observations and, finally, discuss the application of seismic arrays to earthquake early warning systems.

## 9.2 Introduction

“The Conference of Experts to study the methods of detecting violations of a possible agreement on the suspension of nuclear tests” held in 1958 in Geneva under the auspices of the United Nations, was followed by several initiatives for improving the quality of seismic stations worldwide. At the same time, the idea of installing arrays of sensors to improve the Signal-to-Noise Ratio (SNR) of seismic onsets was adopted from radio astronomy, radar, acoustics, and sonar. In the 1960s, it was demonstrated that seismic arrays could facilitate detection and characterization of seismic signals that was superior to that possible using single three-component stations. Today, many of the seismic stations of the International Monitoring System (IMS) for the Comprehensive Nuclear-Test-Ban Treaty Organization (CTBTO) (see e.g., Dahlman et al., 2009) are arrays.

Seismic arrays were used to investigate the nature and source regions of microseisms (e.g., Capon, 1969b; Lacoss et al., 1969; Cessaro, 1994; Friedrich et al., 1998; Essen et al., 2003; Schulte-Pelkum et al., 2003). Since the 1980 (see e.g., Furumoto et al., 1990; Chouet, 1996; Almendros, 1997), seismic arrays have also become a tool for locating and tracking volcanic tremor and for analyzing complex seismic wave-field properties in volcanic areas (see also Section 13.4.4 in Chapter 13). Not only usual seismometers but also strong motion instruments (often sensitive to ground accelerations) have been installed in array configurations to study near field effects of earthquakes (see e.g., Spudich and Cranswick (1984); Abrahamson et al., (1987); Goldstein and Archuleta (1991b); Chiu et al., (1994); Huang (2001); or more recently Halldorson et al., 2009). For borehole installations of strong-motion sensors in array configurations see Section 7.4.6 of Chapter 7. Another possible application of seismic arrays was shown by Frankel et al. (1991), who used a temporary, very small aperture array to track aftershock activities after the 1989 Loma Prieta earthquake.

A seismic array is a set of seismometers deployed so that characteristics of the seismic wavefield at a specified reference point, within or close to the array, can be inferred by analyzing the waveforms recorded at the different sites. A seismic array differs from a local network of seismic stations mainly by the techniques used for data analysis. Thus, in principle, a network of seismic stations can be used as an array, and data from an array can be analyzed as data from a network. The size of an array is defined by its aperture, which is the largest horizontal distance between two sensors of the array. In practice, the geometry and the number of seismometer sites of an array are determined by the intended scientific purpose and economic limits. Details about array configurations can be found e.g., in Barber (1958; 1959), Haubrich (1968), Harjes and Henger (1973), or in Mykkeltveit et al. (1983; 1988).

Most array processing techniques require high signal coherency across the array, and this imposes signal frequency dependent constraints on the array geometry (spatial extent) and instrumentation. A homogeneous geology below all sites of the array is preferable since geological heterogeneity will diminish waveform semblance. Furthermore, proper analysis of array data depends on a stable, high precision relative timing of all array elements. This is required because most of the parametric information calculated using the array involves the measurement of (usually very small) time differences (phase shifts) between the seismic arrivals on the different sensors.

The superior signal detection capabilities of seismic arrays result from the use of so-called “beamforming” techniques whereby the signals at different sensors are delayed and stacked. The SNR is enhanced since the signals interfere constructively whereas the (random)

background noise is suppressed. Arrays can also provide estimates of the station-to-event azimuth (backazimuth, BAZ) and the apparent velocity of the seismic signals. These estimates are important both for event location purposes and for signal identification and classification, e.g., as P, S, local, regional, or teleseismic phases.

In principle, the slowness resolution of an array – i.e., how accurately the direction and apparent velocity of an incoming wavefront can be measured – improves with increasing the array aperture. However, the signal coherency diminishes with increasing sensor separation and so the spatial extent of an array is usually specified to provide an optimal trade-off between coherency and theoretical slowness resolution.

In this chapter we describe procedures for estimating the apparent wavefront velocity (inverse of the slowness or ray parameter), the angles of approach (backazimuth and incidence angle; cf. Figs. 9.8 and 9.9 in this Chapter) of a seismic signal as well as basic processing algorithms for signal detection, one-array regional phase association, and the preparation of an automatic event bulletin. Throughout the text, we use data processing examples from the large NORSAR array (NOA) in southern Norway and from the “regional” arrays NORES in southern Norway, ARCES in northern Norway and GERES in southeast Germany.

There have, historically, been few publications that provide tutorial details about basic array processing. There are numerous papers detailing advanced techniques and presenting results from observations, but the basics of beamforming and STA/LTA detection processing are mostly assumed to be known. The processing algorithms used are similar to many types of signal-processing applications and time-series analysis in fields such as radar technology (e.g., Barber, 1958; 1959) and in seismic prospecting. In seismic prospecting, “beamforming” is called “stacking”.

The basic processing techniques developed in the 1960s have survived, and are still in use. The *Seismic Array Design Handbook* (IBM, 1972) describes the processing algorithms for the arrays LASA and NORSAR (today NOA). References therein are mostly to reports prepared by J. Capon and R. T. Lacoss of Lincoln Laboratories. A description of many array methods and early array installations can be found in a proceedings volume (Beauchamp, 1975) of a NATO Advanced Study Institute conference in 1974 in Sandefjord, Norway. In 1990, a special issue of the *Bulletin of the Seismological Society of America* was published (Volume **80**, Number 6B) with contributions from a symposium titled “Regional Seismic Arrays and Nuclear Test Ban Verification”. This issue contains many papers on theoretical and applied array seismology. More recent reviews on array applications in seismology can be found in e.g., Douglas (2002), Gibbons et al. (2009) or Rost and Thomas (2002; 2009). Since the 1970s, NORSAR has published Semiannual Scientific Reports, and several issues of this series cover array data processing techniques and their applications. A special issue on “Array seismology in Europe: recent developments and applications” was published in 2011 in the *Journal of Seismology* (for an overview see Schweitzer and Krüger, 2011).

At the NORSAR Data Processing Center (NDPC) at Kjeller, Norway, data have been acquired for many years from different types of arrays: e.g., the large aperture NORSAR array (today NOA), the small aperture arrays NORES and ARCES and the very small aperture array on Spitsbergen (SPITS). We describe the general array-processing techniques for training purposes and for use as a reference for analysts new to the field of seismic array processing. Some algorithms are described in detail, whereas others have references to available

literature. It is assumed that the reader has a basic knowledge of time-series analysis such as bandpass filtering and Fourier transforms (e.g., Buttkus, 2000; Scherbaum, 2001).

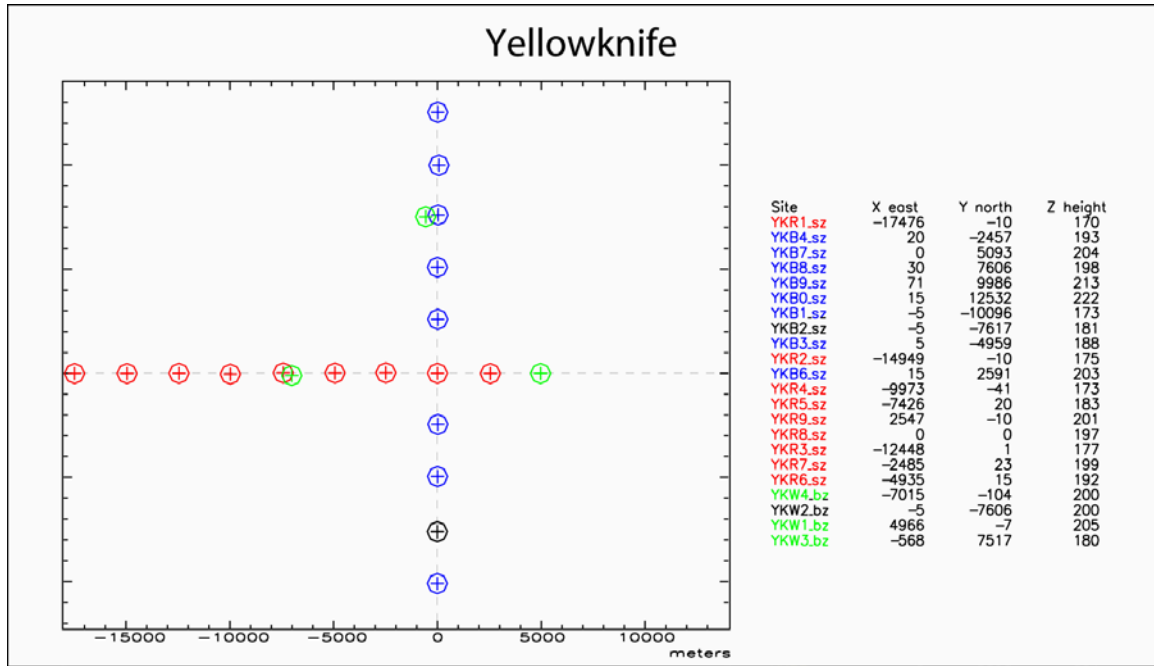
Communication technologies for data transmission (in near real time) between the arrays sites and data processing centers are now available and cost effective. The volume of data generated by an array of seismometers and the associated digital signal processing no longer represent insurmountable technical difficulties. However, low-threshold detection processing leads to large numbers of triggers, all of which require analysis. It is therefore of great importance to use techniques that are robust and easy to operate in an automatic, uninterrupted mode. Three main stages of automatic data processing are employed at the NDPC:

- Detection Processing (DP), which uses beamforming, filtering and STA/LTA detectors to define signal triggers;
- Signal Attribute Processing (SAP), which uses techniques such as frequency-wavenumber ( $f$ - $k$ ) analysis to estimate the slowness vector, and other techniques to estimate parameters such as the onset time, period, amplitude and polarization attributes for every trigger; and
- Event Processing (EP), which analyzes the attributes and sequence of triggers to associate seismic phase arrivals to define events.

A documentation of these processing steps can be found in Mykkeltveit and Bungum (1984) with results from a first version of a program (called RONAPP) for detecting and associating seismic signals from regional events using data from the small aperture, regional array NORES. Later, the NORSAR processing was re-coded to adapt to any array, several data formats and machine architectures. These algorithms are packaged into the programs DP for continuous detection processing, and EP for automatic signal attribute processing, event processing, and interactive special processing (Fyen, 1989; 2001a and b). These programs have been used for almost all examples herein. Section 9.13 shows and explains the output of this automatic data processing for some signals observed with the ARCES array as an example of routine array data analysis.

## 9.3 Examples of seismic arrays

To the best of our knowledge, the first experimental seismic array with more than four elements and openly available data was established in February 1961 by the United Kingdom Atomic Energy Agency (UKAEA) on Salisbury Plain (UK), followed in December 1961 by Pole Mountain (PMA, Wyoming, USA), in June 1962 by Eskdalemuir (EKA, Scotland, UK), and in December 1963 by Yellowknife (YKA, Canada). These types of arrays (the so-called UK-arrays) are orthogonal linear or L-shaped. Later, arrays of the same type were built in Australia (Warramunga), Brazil (Brasilia), and India (Gauribidanur). A detailed description of this type of arrays can be found in Keen et al. (1965), Birtill and Whiteway (1965), and Whiteway (1965; 1966). Fig. 9.1 shows the configuration of the Yellowknife array (Somers and Manchee, 1966; Manchee and Weichert, 1968; Weichert, 1975) as one example of this kind of medium-sized array, which is still in operation. The apertures of the UKAEA arrays vary between 10 and 25 km.



**Fig. 9.1** Configuration of the UKAEA - type Yellowknife array (YKA) in Northern Canada. The blue and the red sites have vertical short-period instruments, while three-component, broadband seismometers are installed at the green sites. The center seismometer of YKA has the geographic coordinates 62.4932°N, 114.6053°W.

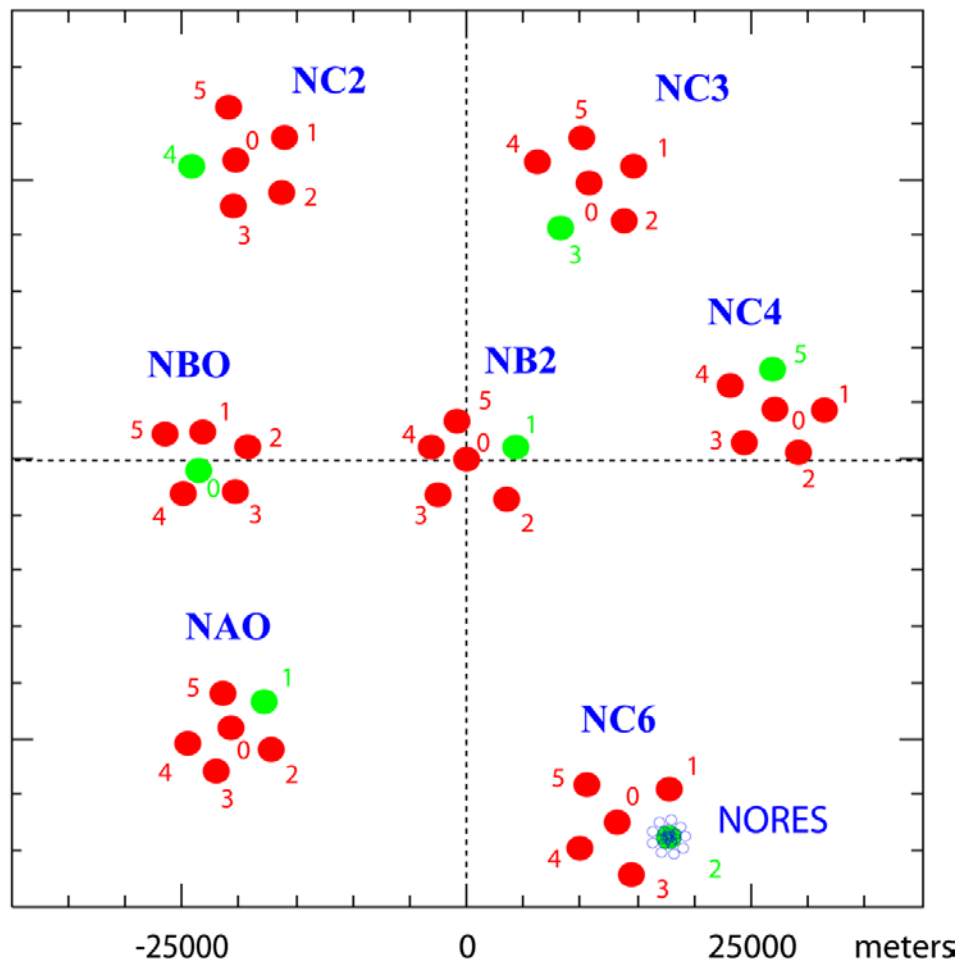
Since the 1950s, classified arrays deployed to monitor nuclear test activities at teleseismic distances had been built worldwide. Many of these arrays became known in the 1990s and are today part of the IMS as primary or auxiliary stations, e.g., AKASG (Malin, Ukraine), ASAR (Alice Springs, Australia), BRTR (Keskin, Turkey), CMAR (Chiang Mai, Thailand), ESDC (Sonseca, Spain), ILAR (Eielson, Alaska), and KURK (Kurchatov, Kazakhstan). Many of these arrays have quite diverse geometries and in some cases comprise different installations for short and long period signals (e.g., the Belbaşı and the Keskin array: Kuleli et al., 2001; Semin et al., 2011).

In the 1960s, arrays with very different apertures and geometries were tested, from small circular ones with apertures of a few kilometers to huge arrays with apertures of up to 200 km. The largest arrays were the Large Aperture Seismic Array (LASA: Frosch and Green, 1966) in Montana (USA), opened in 1965 and in operation until 1978 with 525 seismometer sites, and the original Norwegian Seismic Array (NORSAR: Bungum et al., 1971) in southern Norway, consisting of 132 sites over an aperture of approximately 100 km with altogether 198 seismometers, which became fully operational in spring 1971. The original NORSAR array was reduced in 1976 to seven subarrays (see Fig. 9.2) and was assigned the new code name NOA.

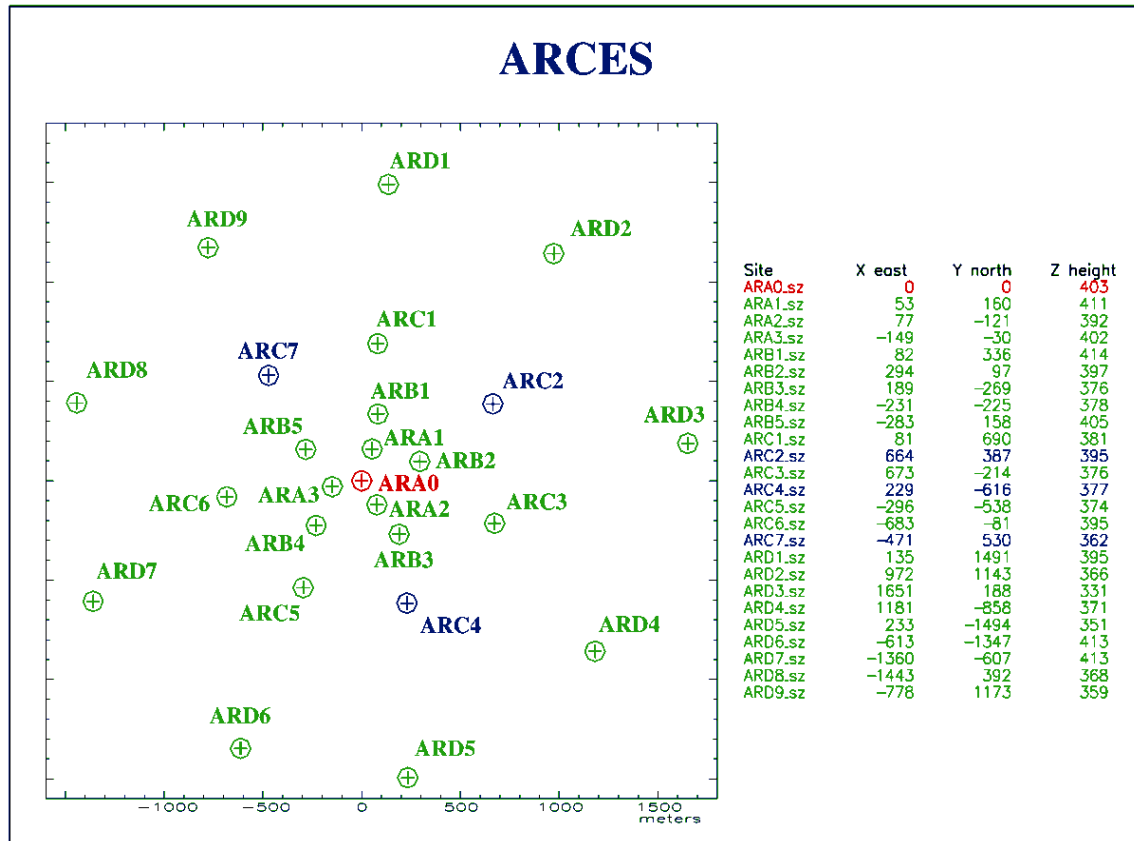
Figs. 9.2 and 9.3 show the configurations of the present day NORSAR (NOA) and NORES arrays and the layout of the seismometer sites for the ARCES array. The NORES and ARCES-type array design with sites located on concentric rings (each with an odd number of sites) spaced at log-periodic intervals is now used for most of the modern small aperture arrays; only the number of rings and the aperture (diameter of the outermost ring of sites) differ from installation to installation. For example, the SPITS array has only nine sites and

corresponds to the center site plus the A- and the B-ring of a NORES-type array; the FINES array in southern Finland consists of three rings with 16 sites altogether. The geometry of these regional, relatively small arrays was developed in the 1980s and 1990s. Today, the configuration of a SPITS-type 9-element array constitutes the standard for new arrays in the IMS, but in general with a larger aperture than SPITS.

## NOA and NORES



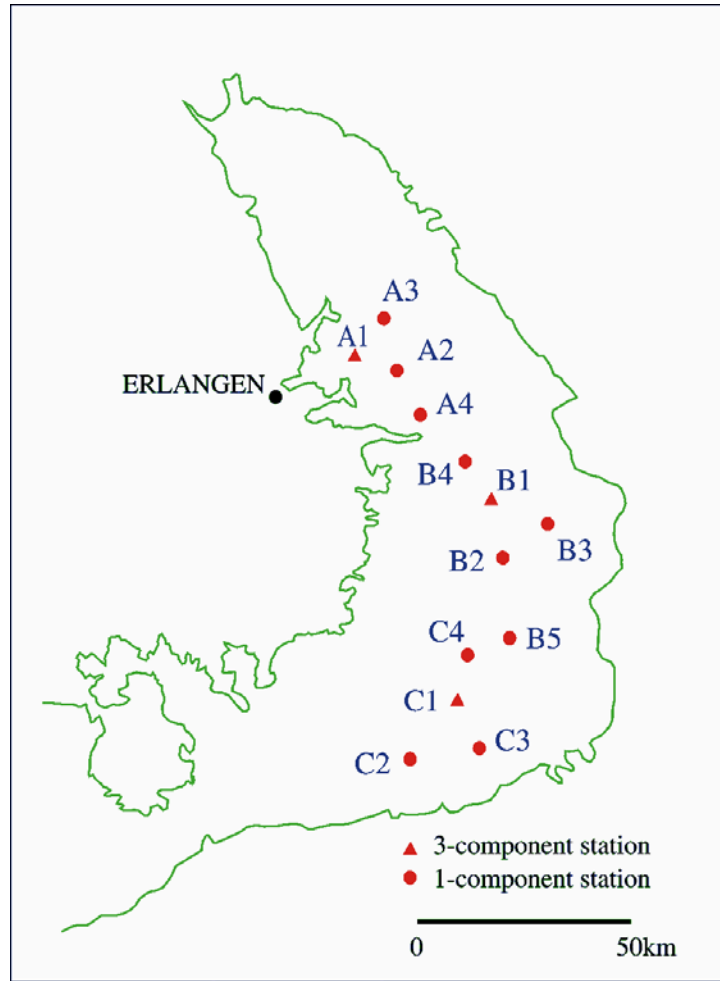
**Fig. 9.2** Configuration of the present day large aperture NORSAR array (NOA) and the small aperture array NORES. The NORES array is co-located with the NORSAR subarray NC6. The aperture of NOA has, since 1976, been about 60 km and the aperture of NORES is about 3 km (see the small blue symbols). Each seismometer site is marked with a circle. The present NOA configuration has 42 sites, whereas the NORES array has 25 sites. NOA has seven subarrays, each with six vertical seismometers installed in shallow boreholes. In addition, one site in each subarray (marked in green) has one three-component broadband seismometer. The geometry of NORES is identical to the geometry of ARCES shown in Fig. 9.3. The center seismometer of NOA in subarray NB2 has the geographic coordinates 61.03972°N, 11.21475°E. The center seismometer of NORES has the geographic coordinates 60.73527°N, 11.54143°E.



**Fig. 9.3** Configuration of the regional array ARCES, which is identical to the NORES array. Each vertical short-period seismometer site is marked with a circle and a cross. The ARCES array has 25 sites with vertical short-period seismometers. Four of these sites have in addition short-period horizontal seismometers. These short-period three-component sites are marked in blue or red. A broadband three-component seismometer is co-located with the short-period three-component seismometer at the center site (red). ARCES has one center instrument – ARA0 – and four rings: the A-ring with three sites and a radius of about 150 m, the B-ring with five sites and a radius of about 325 m, the C-ring with seven sites and a radius of about 700 m, and finally, the D-ring with nine sites and a radius of about 1500 m. The center seismometer of ARCES has the geographic coordinates 69.53486°N, 25.50578°E. The table on the right of the figure gives the relative coordinates between the single sites and the center site ARA0, and the elevation of all sites above sea level in meters.

The large LASA and NORSAR arrays and the UKAEA arrays had narrow-band short-period seismometers at all sites and additional long-period seismometers at selected sites in their original configurations, whereas the Gräfenberg Array (GRF) in Germany was planned and installed in the early 1970s as an array of broadband sensors. It has an aperture of about 100 km (Harjes and Seidl, 1978; Buttkus, 1986) and an irregular shape (Fig. 9.4) that follows the limestone plateau of the Franconian Jura.





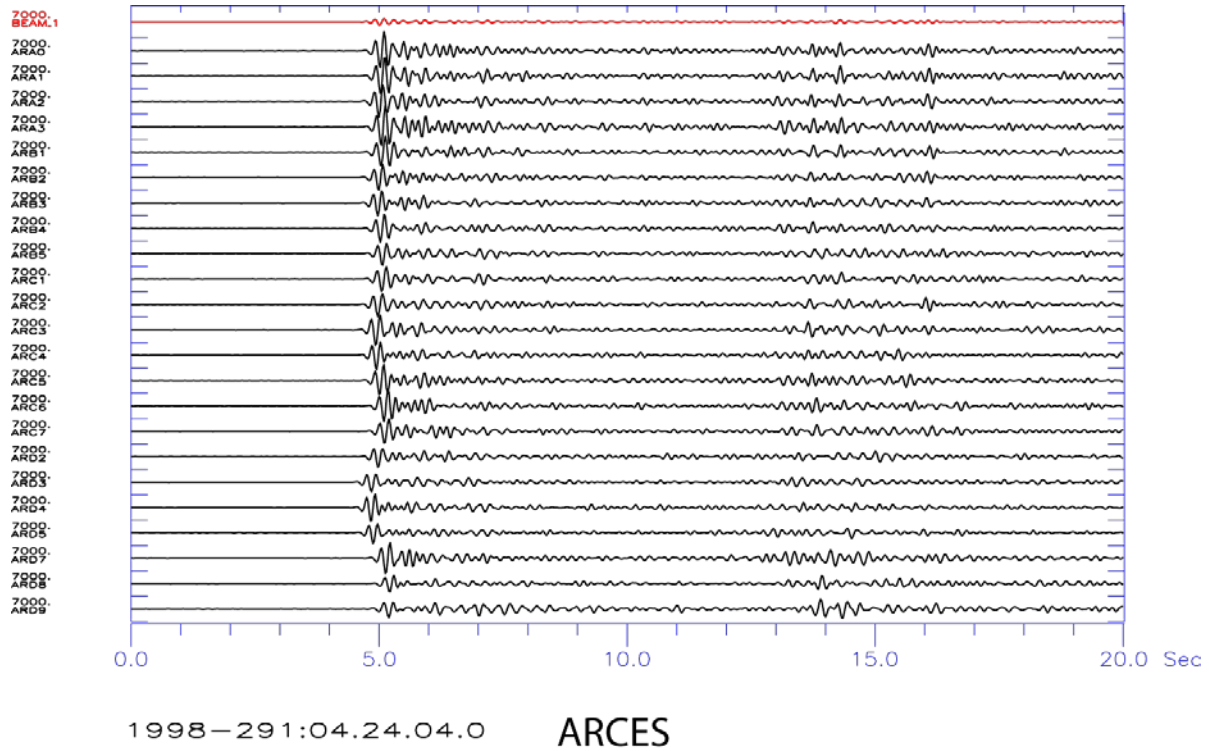
**Fig. 9.4** Configuration of the irregularly shaped Gräfenberg array (GRF). Since several years, all sites are equipped with three-component, broadband STS-2 seismometers. The green line follows the geological unit Franconian Jura, on which the array is located. The reference station GRA1 (at position A1 on the map) has the geographical coordinates 49.69197°N and 11.22200°E.

Another approach to arrays was developed in the 1990s. In parts of Japan and in the USA the networks of seismometer stations are so dense that data from all stations can be combined and used as array data. Examples of these are the so-called J-array, the Californian array and the ongoing USArray project with semi-temporary stations (Levander et al., 1999). All known array techniques can be applied to analyze data from these networks (see e.g., J-array Group, 1993 or Benz et al., 1994).

## 9.4 Beamforming

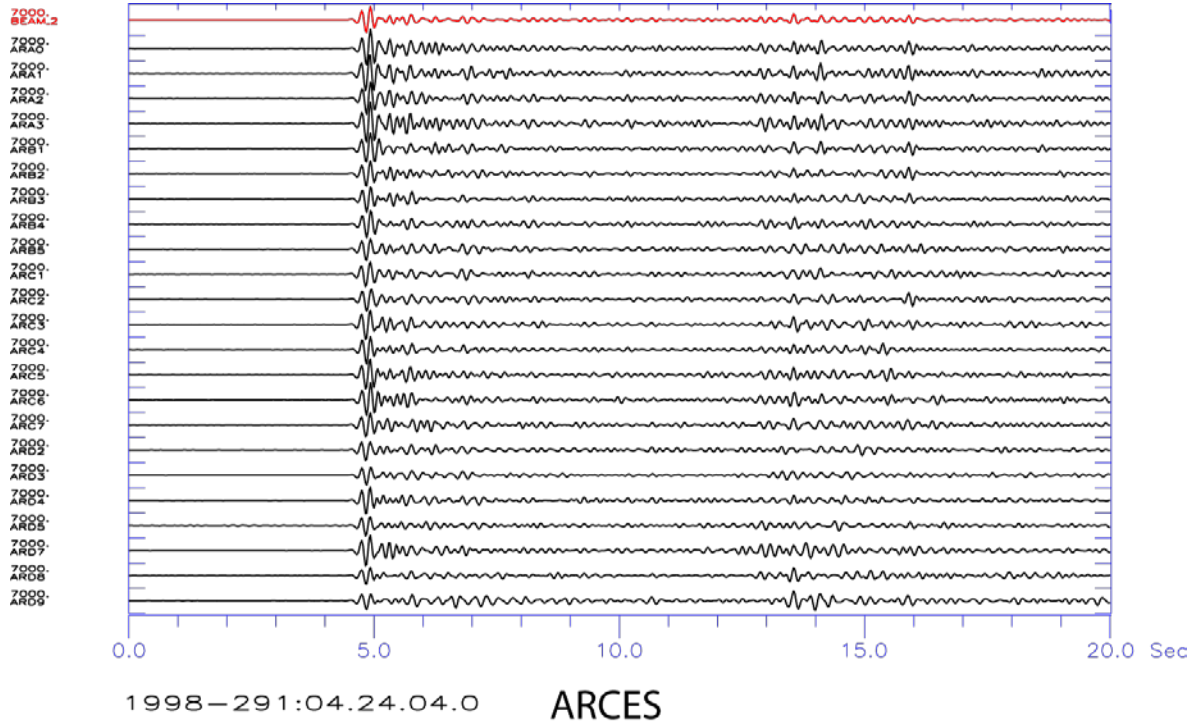
With an array we can improve the signal-to-noise ratio (SNR) of a seismic signal by summing the coherent part of seismic signals as observed at the different single array sites. Figs. 9.5 and 9.6 show P onsets of a regional event observed at the ARCES sites and, in addition, the summation trace (on top) of all single observations divided by the number of traces. In Fig 9.5 the data were summed without taking into account any signal delay times between the

different stations due to finite apparent velocity of wave propagation. Consequently, the P onset is suppressed by destructive interference. In Fig. 9.6 all traces were time-adjusted to provide alignment of the first P pulse before summation. Note the clear and short P pulse on the beam trace and the suppression of incoherent energy in the P coda and preceding noise. In the following, all calculations of summation traces always includes as a last processing step the division of the summation trace by the number of traces summed up. This is done to conserve the amplitude of seismic signals for any later signal analysis.



**Fig. 9.5** The figure shows P-phase onsets of a regional event observed with the vertical short-period seismometers of ARCES. The top trace is an array beam, and the remaining traces are single vertical short-period seismograms. All data were filtered with a Butterworth band pass filter between 4 and 8 Hz and are shown with a common amplification. All traces were summed to create the beam, which was then divided by the number (23) of traces, without any delay-time application.

This demonstrates that the most significant consideration in the summation (or beamforming) process is to find the best delay times, with which the single traces must be shifted before summation (“delay and sum”) in order to get the largest amplitudes due to coherent interference of the signals. The simplest way is just to pick the onset times of the signal on each trace and shift the traces with respect to the onset time at the reference site of the array. However, most onsets from weaker events have a much smaller SNR than in the example shown, and therefore onset times are often difficult to pick. With hundreds of onsets each day, this is not practical during routine operation. Therefore for routine processing of array data the best approach is to predefine many different beams and to calculate them automatically. A detector can then search for onsets on these beams.



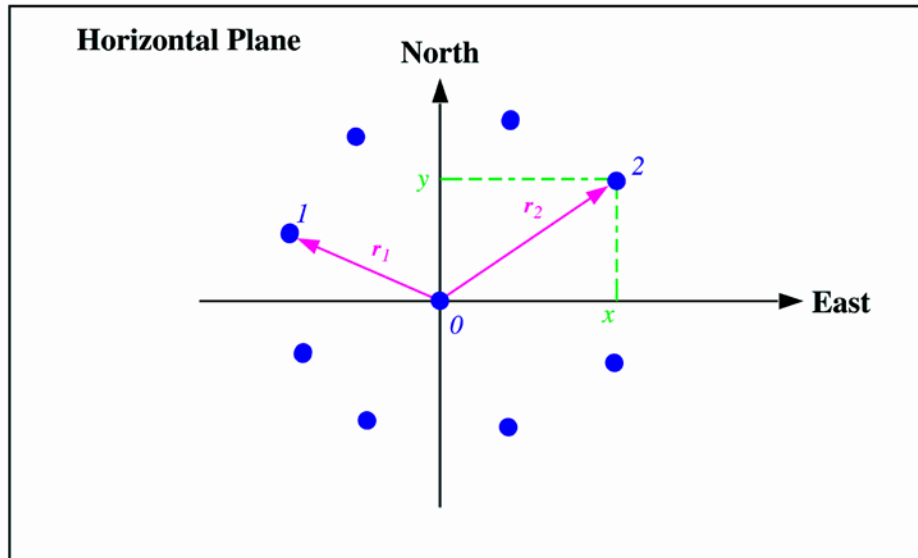
**Fig. 9.6** This figure shows P-phase onsets of a regional event observed with the vertical short-period seismometers of ARCES as in Fig. 9.5 but the single traces were first aligned, summed up and then divided by the number of traces (23). Note for this case the clear and relatively short pulse form of the first P onset on the beam and the suppression of incoherent energy in the P coda.

In the following sections, we explain how such delay times can be theoretically calculated for known seismic signals, using some basic equations and parameter definitions, and we give the formulas for a seismic beam.

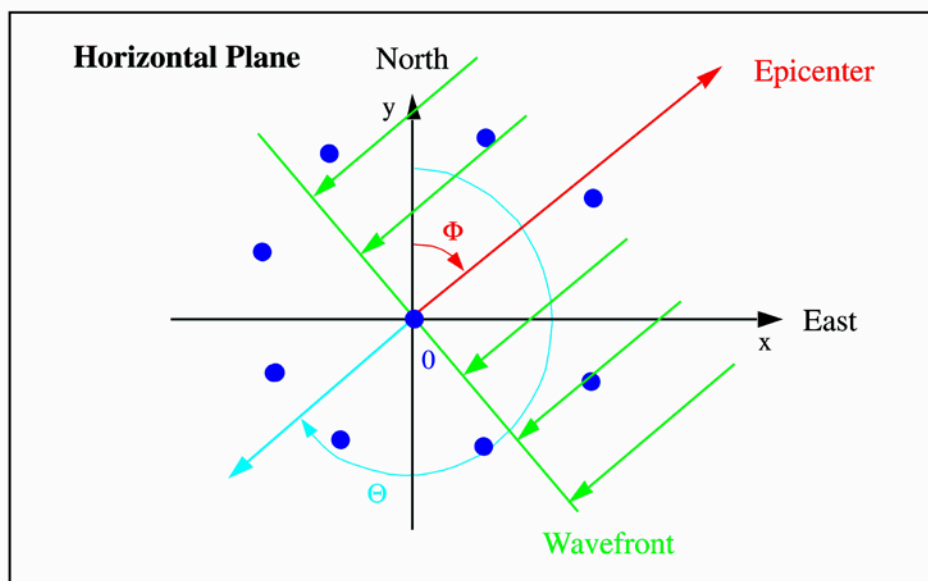
### 9.4.1 Geometrical parameters

As aforementioned, an array is defined as a set of seismometers with one seismometer site or any other location within or close to the array being assigned the role of a reference point. The relative distances from this reference point to all other array sites are used later in all array specific analysis algorithms (Fig. 9.7).

$\mathbf{r}_j$  Position vector of instrument  $j$  with a distance (absolute value)  $r_j$  from the defined reference point. We use bold characters for vectors and normal characters for scalars. The position is normally given relative to a (central) instrument at site  $O$ ,  $\mathbf{r}_j = f(x, y, z)$ , where  $(x, y, z)$  are the Cartesian coordinates measured in [km] with positive axes towards East ( $x$ ), towards North ( $y$ ), and in the vertical up above sea level ( $z$ ).



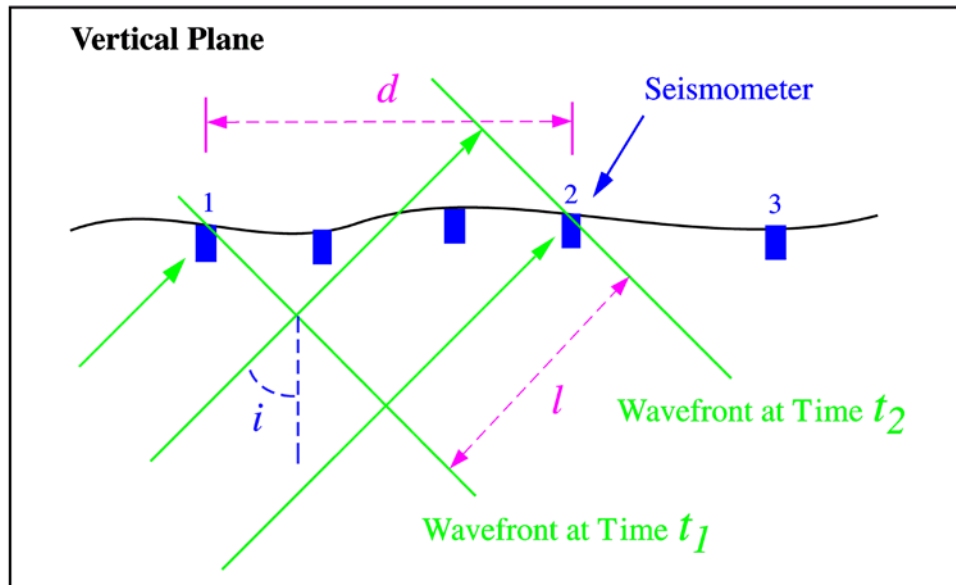
**Fig. 9.7** Illustration (horizontal plane) of an array of instruments (filled circles). The center instrument  $0$  is used as reference and origin for the relative coordinates  $x$ ,  $y$  (see also Fig. 9.3 for an example of such a circular array).



**Fig. 9.8** Definition of the angles  $\Theta$  (direction of wavefront propagation) and  $\Phi$  (direction to the epicenter = backazimuth); here, e.g., for a wavefront coming from north-east and crossing the array in a south-westerly direction.

For distances from the source much larger than about 10 wavelengths, a seismic wave approaches an array as a **wavefront** that is close to **planar**. The case when the (simplified) plane-wave approach is no longer applicable and a circular wavefront has to be assumed is discussed in Almendros et al. (1999). The directions of approach and propagation of the wavefront projected on to the horizontal plane are defined by the angles  $\Phi$  and  $\Theta$  (Fig. 9.8).

- $\Phi$  Backazimuth (often abbreviated as BAZ) = angle of wavefront approach, i.e., direction angle from a station to an epicenter, measured clockwise from the North to the direction towards the epicenter in  $[\circ]$ . In many cases this angle is for short called azimuth, but the recommendation is to use in all cases backazimuth to distinguish it from other azimuth measures.
- $\Theta$  Direction in which the wavefront propagates, also measured in  $[\circ]$  from the North, with  $\Theta = \Phi \pm 180^\circ$ .



**Fig. 9.9** Illustration (vertical plane) of a seismic plane wave crossing an array at an angle of incidence  $i$ .

In the vertical plane, the angle measured between the direction to the source and the vertical is called the angle of incidence  $i$  with  $i \leq 90^\circ$  (see Fig. 9.9). The seismic velocity below the array in the uppermost crust and the angle of incidence define the apparent propagation speed of the wavefront crossing the array. Be aware that the seismic velocity below the array is an effective mean velocity of the uppermost crustal layers depending on the dominant wavelength (frequency) of the seismic signal.

### 9.4.2 Apparent velocity and slowness

The upper crustal velocity together with the angle of incidence defines the apparent propagation speed of the wavefront at the observing instruments. This is **not** the physical propagation speed of the wavefront and is therefore called **apparent** velocity (see also Section 2.5.2 of Chapter 2). We start our consideration by defining the following quantities (for definition see Fig. 9.9):

- $d$  horizontal distances between different array sites in [km];
- $v_c$  effective, mean crustal velocity (for P or S waves, depending on the seismic phase) immediately below the array in [km/s];
- $i$  angle of incidence;

- $\mathbf{v}_{app}$  apparent velocity vector with the absolute value  $v_{app} = 1/s$ .  $\mathbf{v}_{app} = (v_{app,x}, v_{app,y}, v_{app,z})$ , where  $(v_{app,x}, v_{app,y}, v_{app,z})$  are the single apparent velocity components in [km/s] of the wavefront crossing an array;
- $v_{app}$  absolute value of the apparent velocity vector in [km/s] of a plane wave crossing an array (see Fig. 9.10);
- $v_{app,h}$  absolute value of the horizontal component of the apparent velocity
- $$v_{app,h} = \sqrt{v_{app,x}^2 + v_{app,y}^2}.$$

The inverse of the apparent velocity  $v_{app}$  is called slowness  $s$ , which is a constant for a specific ray. For local or regional applications the unit of slowness is [s/km]. For global applications it is more appropriate to use the unit [s/°] and slowness is then called ray parameter. The ray parameter of major seismic phases is usually tabulated for standard Earth models together with the travel times as a function of distance from the source. The following symbols are used:

- $\mathbf{s}$  slowness vector with absolute value  $s = 1/v_{app}$ .  $\mathbf{s} = (s_x, s_y, s_z)$ , where  $(s_x, s_y, s_z)$  are the single, inverse apparent velocity (= slowness) components in [s/km]. Note, because the vector  $\mathbf{s}$  is oriented in the propagation direction (in direction of  $\Theta = 225^\circ$ , see Fig. 9.8), a plane wave with backazimuth  $\Phi = 45^\circ$  would have negative values for both horizontal components with North and East directions being positive;
- $s$  absolute value of the slowness vector in [s/km] of a plane wave crossing an array;
- $p$  ray parameter  $p = s \cdot g$ , measured in [s/°], for a spherical Earth model with a radius of 6371 km, where  $g = \frac{\pi \cdot 6371 \text{ km}}{180^\circ} \cong 111.19 \text{ [km/°]}$ .

The relation between the parameters of a plane wave and the actual seismic signal is given by the wavenumber vector  $\mathbf{k}$ :

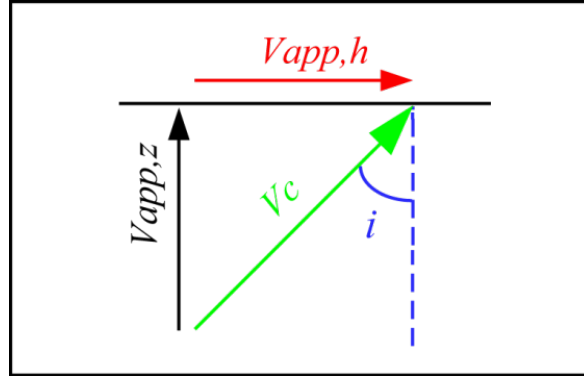
- $\mathbf{k}$  wavenumber vector defined as  $\mathbf{k} = \omega \cdot \mathbf{s}$  with the angular frequency  $\omega = 2 \cdot \pi \cdot f = 2 \cdot \pi / T$  measured in [1/s].  $T$  is the period and  $f$  the frequency of the seismic signal;
- $k$  absolute value of the wavenumber vector  $\mathbf{k}$  defined as  $k = \omega \cdot s = 2 \cdot \pi \cdot f \cdot s = 2 \cdot \pi / \lambda$ , measured in [1/km].  $\lambda$  is the wavelength of the signal and because of the analogy between  $\omega$  and  $k$ ,  $k$  is also called spatial frequency.

The time delay  $\tau_j$  is defined as the arrival-time difference of the wavefront between the seismometer at site  $j$  and the seismometer at the reference site. The unit of measurement is seconds with a positive delay meaning a later arrival with respect to the reference site.

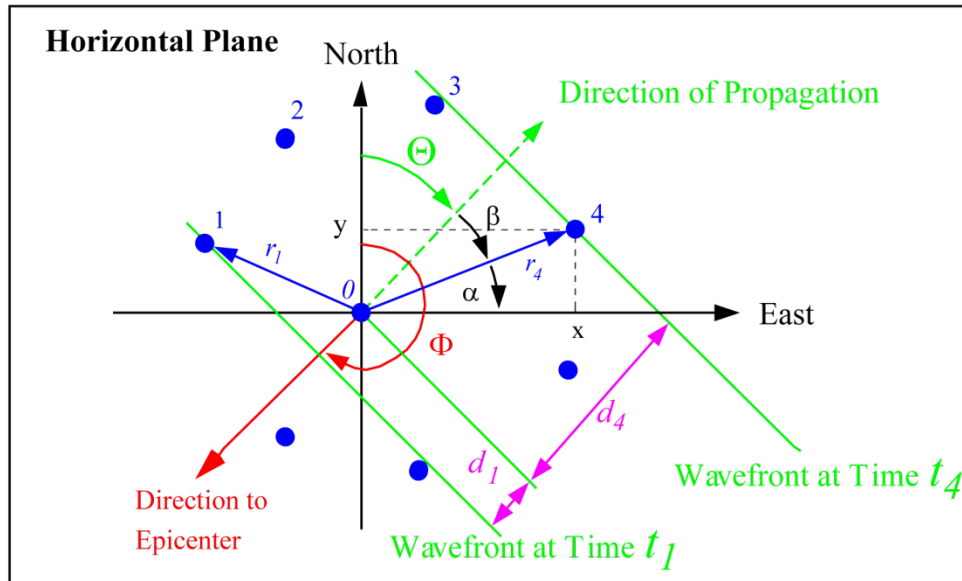
Assume a wavefront is propagating the distance  $l$  between time  $t_1$  and time  $t_2$  (Fig. 9.9). Then, if  $d$  is used for the horizontal distance between instrument 1 and 2 in [km], and if both instruments are assumed to be at the same elevation, we have:

$\tau_2 = (t_2 - t_1) = \frac{l}{v_c}$ , and the horizontal component of the apparent velocity  $v_{app,h}$  is then defined as a function of the incidence angle  $i$  (Fig. 9.10):

$$v_{app,h} = \frac{d}{(t_2 - t_1)} = \frac{v_c}{\sin i} \quad v_{app,z} = \frac{v_c}{\cos i} \quad (9.1)$$



**Fig. 9.10** A plane wave propagating in the direction of the green arrow with the velocity  $v_c$  reaches the Earth's surface. The decomposition of this velocity into a vertical ( $v_{app,z}$ ) and a horizontal component ( $v_{app,h}$ ) of the apparent velocity depends on the incidence angle  $i$ . The horizontal velocity component is only equal to the propagation velocity  $v_c$  for waves propagating parallel to the surface; in all other cases  $v_{app}$  is higher than  $v_c$ .



**Fig. 9.11** Illustration (horizontal plane) of a plane wave, coming from the Southwest (backazimuth  $\Phi$ ), crossing an array and propagating in a north-easterly direction  $\Theta$ .

### 9.4.3 Plane-wave time delays for sites in the same horizontal plane

In most cases, the elevation differences between single array sites are so small that travel-time differences due to elevation differences are negligible (Fig. 9.9). We can assume, therefore, that all sites are in the same horizontal plane. In this case, we cannot measure the vertical component of the wavefront propagation. The vertical apparent velocity component can then

be defined as  $v_{app,z} = \infty$ , and the corresponding slowness component becomes  $s_z = 0$ . From Fig. 9.11 we see that the time delay  $\tau_4$  [s] between the center site 0 and site 4 with the relative coordinates  $(x_4, y_4)$  is

$$\tau_4 = \frac{d_4}{v_{app,h}} = \frac{r_4 \cdot \cos \beta}{v_{app,h}}, \text{ with } r_4 = |\mathbf{r}_4|.$$

Now let us omit the subscript 4, and evaluate further:

$$\begin{aligned} \alpha + \beta + \Theta &= 90^\circ, \quad r \cdot \cos \beta = d, \quad r \cdot \cos \alpha = x, \quad d = \frac{x \cdot \cos \beta}{\cos \alpha} \\ d &= \frac{x \cdot \cos(90^\circ - \alpha - \Theta)}{\cos \alpha} = \frac{x \cdot (\sin \alpha \cdot \cos \Theta + \cos \alpha \cdot \sin \Theta)}{\cos \alpha} = x \cdot \frac{\sin \alpha}{\cos \alpha} \cdot \cos \Theta + x \cdot \sin \Theta \\ d &= x \cdot \frac{y}{x} \cdot \cos \Theta + x \cdot \sin \Theta = y \cdot \cos \Theta + x \cdot \sin \Theta \end{aligned}$$

With  $\Theta = \Phi \pm 180^\circ$  (Fig. 9.8), we get for the horizontal distance traveled by the plane wave  $d = -x \cdot \sin \Phi - y \cdot \cos \Phi$ .

Then, for any site  $j$  with the horizontal coordinates  $(x, y)$ , but without an elevation difference relative to the reference (center) site, we get the time delay  $\tau_j$ :

$$\tau_j = \frac{d_j}{v_{app,h}} = \frac{-x_j \cdot \sin \Phi - y_j \cdot \cos \Phi}{v_{app,h}} \quad (9.2)$$

These delay times can also be written in vector syntax with the position vector  $\mathbf{r}_j$  and the slowness vector  $\mathbf{s}$  as parameters. In this notation the delay times are defined as projection of the position vector onto the slowness vector:

$$\tau_j = \mathbf{r}_j \cdot \mathbf{s} \quad (9.3)$$

#### 9.4.4 Plane-wave time delays when including site elevations

In some cases, not all array sites are located on one horizontal plane. Then the calculation of the time delays becomes slightly more complicated. Site 2 (Fig. 9.12) has the relative coordinates  $(x_2, y_2, z_2)$  and we see that  $i + \gamma + \varphi = 90^\circ$ ,  $z_2 = r_2 \cdot \sin \varphi$ ,  $d_2 = r_2 \cdot \cos \varphi$ ,  $l = r_2 \cdot \cos \gamma$ , and  $\tau_2 = \frac{l}{v_c}$ .

$$l = \frac{z_2}{\sin \varphi} \cdot \cos \gamma = \frac{z_2}{\sin \varphi} \cdot \cos(90^\circ - i - \varphi) = \frac{z_2 \cdot (\sin i \cdot \cos \varphi + \cos i \cdot \sin \varphi)}{\sin \varphi}$$

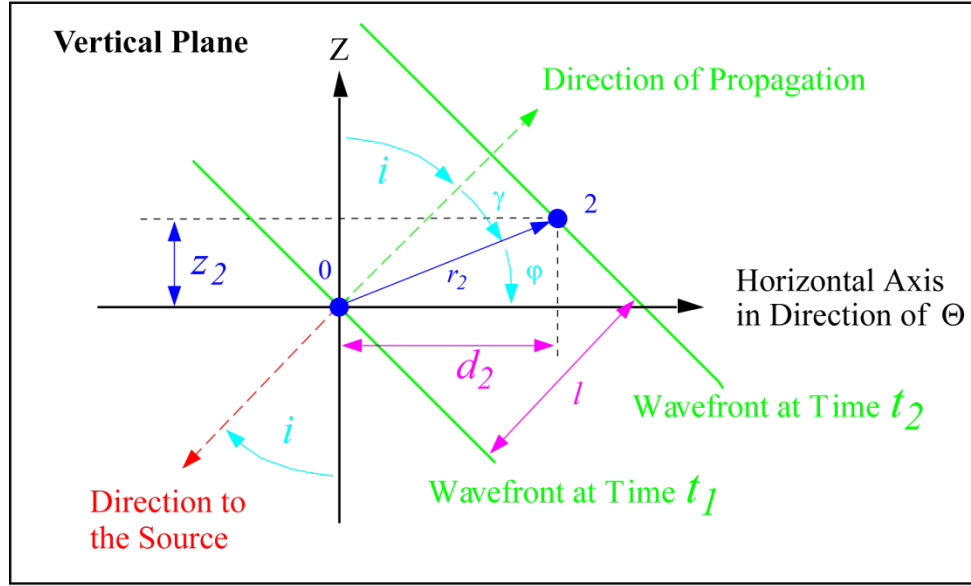


Omitting again the site number, we get:

$$l = z \cdot \frac{\cos \varphi}{\sin \varphi} \cdot \sin i + z \cdot \cos i = d \cdot \sin i + z \cdot \cos i$$

Using Eqs. (9.1) and (9.2), we get for the total time delay at site  $j$

$$\tau_j = \frac{-x_j \cdot \sin i \cdot \sin \Phi - y_j \cdot \sin i \cdot \cos \Phi + z_j \cdot \cos i}{v_c} = \frac{-x_j \cdot \sin \Phi - y_j \cdot \cos \Phi}{v_{app}} + \frac{z_j \cdot \cos i}{v_c} \quad (9.4)$$



**Fig. 9.12** Illustration (vertical plane) of a plane wave crossing an array at the angle of incidence  $i$ .

Now, the time delays  $\tau_j$  also depend on the local crustal velocities ( $v_c$ ) below the given site  $j$  and not just on the parameters of the wavefront ( $\Phi, v_{app}$ ). This is a clear disadvantage of arrays where single sites are not located on one horizontal plane, and if possible elevation variations should be avoided when planning a new array installation. Writing these time delays in vector notation will again result in Eq. (9.3), but note, the vectors are now three-dimensional.

$$\tau_j = \mathbf{r}_j \cdot \mathbf{s}$$

As rule of thumb one can use: elevation corrections should be applied whenever elevation related time delays between different array sites become equal or larger than about  $\frac{1}{4}$  of the dominant signal period.

### 9.4.5 Beamforming formulae

After deriving the delay times  $\tau_j$  for each station by solving Eq. (9.2) or Eq. (9.4) for a specific backazimuth and apparent velocity combination, we can define a “delay and sum” process to calculate the array beam. In the following we will use the shorter vector syntax of Eq. (9.3) to calculate time delays. The calculated delay times can be negative or positive. This depends on the relative position of the single sites with respect to the array’s reference point and to the backazimuth of the seismic signal. Negative delay times correspond to a delay and positive delay times correspond to a signal arriving earlier than at the reference site.

Let  $w_j(\mathbf{r}_j, t)$  be the digital sample of the seismogram from site  $j$  at time  $t$ , then the beam of the whole array is defined as

$$b(t) = \frac{1}{M} \sum_{j=1}^M w_j(t + \mathbf{r}_j \cdot \mathbf{s}) = \frac{1}{M} \sum_{j=1}^M w_j(t + \tau_j). \quad (9.5)$$

This operation of summing the recordings of the  $M$  instruments by applying the time delays  $\mathbf{r} \cdot \mathbf{s}$  and then dividing by the number of traces  $M$  is called beamforming.

Because we are using digital data, sampled with a defined sampling rate, we will always need an integer number of samples in programming Eq. (9.5), that is, the term  $t + \mathbf{r}_j \cdot \mathbf{s} = t + \tau_j$  needs to be converted to an integer sample number. However, to avoid aliasing effects by following the rules of digital signal processing, it is in most cases sufficient for beamforming to use the nearest integer sample, as long as the dominant frequency is lower than ~25% of the sampling rate. If better resolution is needed one has to oversample the data. An example of the positive effects of oversampling is the beampacking algorithm (see section 9.8.2).

If seismic waves were harmonic waves  $S(t)$  without noise, with identical site responses, and without attenuation, then a “delay and sum” with Eq. (9.5) would reproduce the signal  $S(t)$  accurately. The attenuation of seismic waves within an array is usually negligible, but large amplitude differences can sometimes be observed between data from different array sites due to differences in the crust directly below the sites (see Fig. 4.34 in Chapter 4). In such cases, it can be helpful to normalize the amplitudes or to weight the traces before beamforming.

Real data  $w(t)$  are, of course, the sum of background noise  $n(t)$  plus the signal of interest  $S(t)$ , i.e.,  $w(t) = S(t) + n(t)$ .

The actual noise conditions and signal amplitude differences will influence the quality of a beam. However, the noise is usually more incoherent than the signal, i.e., signal and noise will be influenced differently by beamforming. The improvement of the signal-to-noise ratio (SNR) due to the beamforming process can be estimated as follows:

Calculating the sum of  $M$  observations and including noise we get:

$$B(t) = \sum_{j=1}^M w_j(t + \mathbf{r}_j \cdot \mathbf{s}) = \sum_{j=1}^M (S_j(t + \mathbf{r}_j \cdot \mathbf{s}) + n_j(t + \mathbf{r}_j \cdot \mathbf{s})).$$

Assuming that the signal is coherent and not attenuated, this sum can be split and we get:

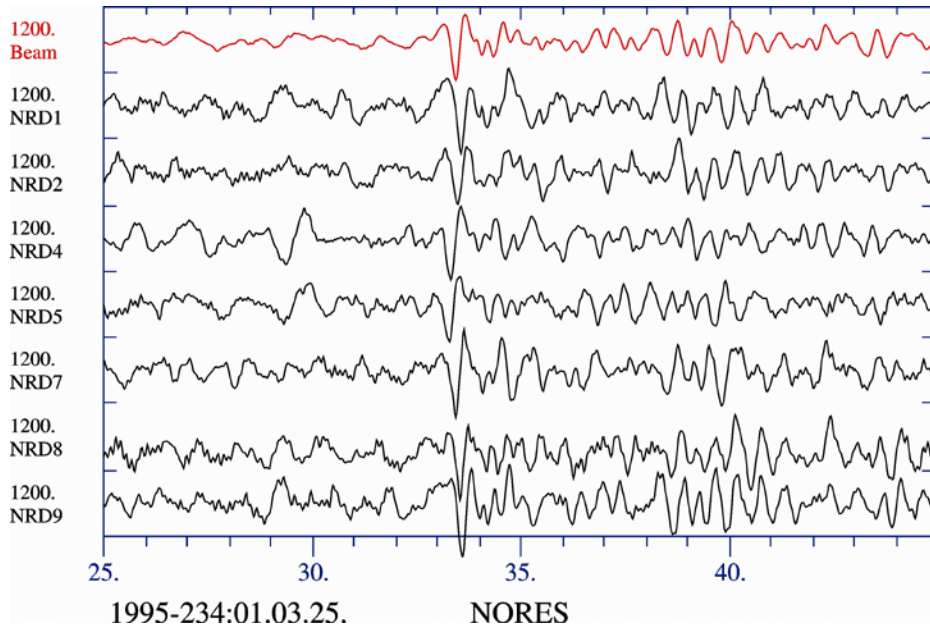
$$B(t) = M \cdot S(t) + \sum_{j=1}^M n_j(t + \mathbf{r}_j \cdot \mathbf{s}). \quad (9.6)$$

Now we assume that the noise  $n_j(\mathbf{r}_j, t)$  has a normal amplitude distribution, a zero mean value, and the same variance  $\sigma^2$  at all  $M$  sites. Then, for the variance of the noise after summation, we get  $\sigma_s^2 = M \cdot \sigma^2$  and the standard deviation of the noise in the beam trace will become  $\sqrt{M} \cdot \sigma$ . That means that the standard deviation of the noise will be multiplied only with a factor of  $\sqrt{M}$ , but the coherent signal with the factor  $M$  (Eq. (9.6)). So, the theoretical improvement of the signal-to-noise ratio (SNR) by the “delay and sum” process will be  $\sqrt{M}$  for an array containing  $M$  sites. The SNR improvement, also called array gain  $G$ , of an  $M$ -sensor array can then be written as

$$G^2 = M \text{ or } G = \sqrt{M}. \quad (9.7)$$

### 9.4.6 Examples of beamforming

In the following, two examples of beamforming are shown with data from the small aperture, short-period arrays NORES and GERES. With sampling rates of 40 Hz, these arrays are optimized to observe seismic phases at regional distances. A further advantage of these arrays is the relatively large number of 25 array elements, which can theoretically give an array gain (SNR improvement) of up to  $G = 5$ .



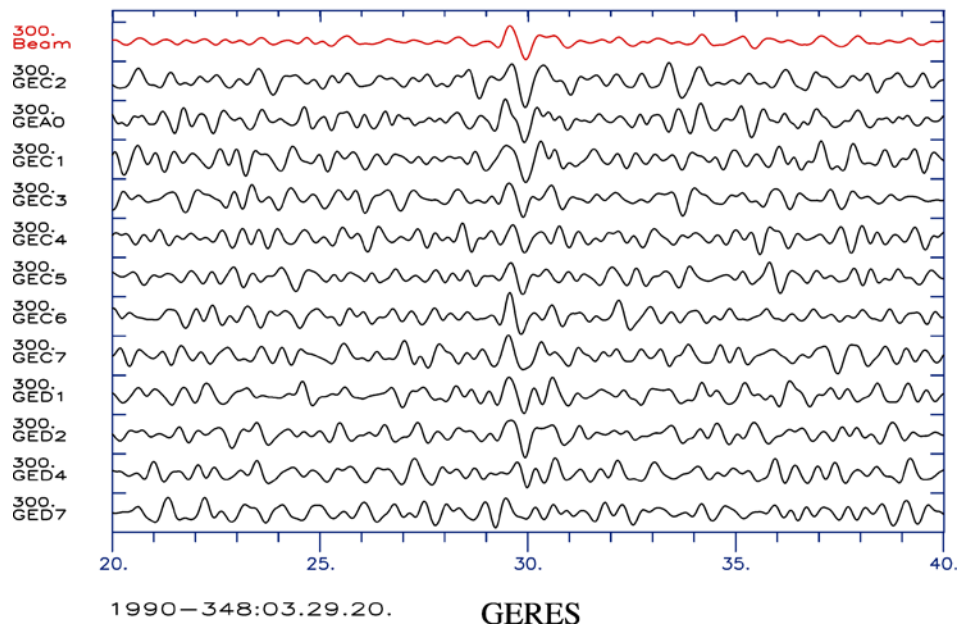
**Fig. 9.13** Selected unfiltered NORES channels recording an event in Greece, with the beam displayed in red at top. All traces have the same amplitude scale and the unit of the time scale is in [s].

In Fig. 9.13 (top trace), we display a beam calculated by using the known apparent velocity ( $v_{app} = 10.0$  km/s) and backazimuth (BAZ =  $158^\circ$ ) for the P-onset of an event in Greece

recorded at NORES at an epicentral distance of  $21.5^\circ$ . 21 vertical sensors of the array have been used, but only the data recorded with the NORES D-ring elements are displayed together with the beam. Note that the signal on the beam is very similar to the individual signals, but the noise changes both in frequency content and amplitude level. The beam is made by calculating time delays for the given slowness and applying Eq. (9.2), and the individual traces have been shifted with these delays. There is an obvious SNR improvement (e.g., between the trace NRD8 and the beam of 2.98) but clearly less than the theoretical optimum of 4.58 times. This can have several reasons, such as:

- Non-optimum calculation of the time delay due to deviation from the assumed planar wave due to velocity anomalies below the array and along the ray path;
- The influence of lateral heterogeneities on the pulse shape of the signal at the different array sites due to multipathing, scattering or signal damping;
- Destructive interference of the noise may not work as theoretically assumed because the mean noise level at the different sites is very different;
- The noise at the different sites is not statistically independent but may have coherent components (Braun and Schweitzer, 2008).

Some approaches to further increase the SNR during beamforming of real data are described under Section 9.5.



**Fig. 9.14** Geres beam (top trace) in red for a PcP onset observed at an epicentral distance of  $9.6^\circ$  from a deep focus event in the Tyrrhenian Sea. All traces are 1 – 3 Hz bandpass filtered and have the same amplitude scale and the unit of the time scale is in [s].

The next example in Fig. 9.14 demonstrates the ability of arrays to detect small signals that are difficult to detect with single stations. It shows the tiny onset of the core reflection PcP recorded at the Geres array from a deep focus event in the Tyrrhenian Sea ( $h = 275$  km) at an epicentral distance of  $9.6^\circ$ . For the “delay and sum” process, data from 20 sites of Geres were used (with only a subset of the single traces shown). Note that although the signal coherency is low, because of the large superimposing noise amplitudes, the noise suppression on the beam is clearly visible. In this case, the SNR improvement is also less than

theoretically expected (e.g., 3.9 between GEC3 and the beam trace instead of the theoretical value of 4.5) but the onset can now be analyzed.

## 9.5 Data analysis with beamforming

### 9.5.1 The n-th root process

A non-linear method to enhance the SNR during beamforming is the so-called n-th root process (Muirhead, 1968; Kanasewich et al., 1973; Muirhead and Ram Datt, 1976). Before summing up the single seismic traces, the n-th root is calculated for each trace retaining the sign information; Eq. (9.5) then becomes:

$$B_N(t) = \frac{1}{M} \sum_{j=1}^M |w_j(t + \tau_j)|^{1/N} \cdot \text{signum}\{w_j(t)\}, \quad (9.8)$$

where the value of the function  $\text{signum}\{w_j(t)\}$  is defined as -1 or +1, depending on the sign of the actual sample  $w_j(t)$ . After this summation, the beam has to be raised to the power of  $N$ , again retaining the sign information:

$$b_N(t) = |B_N(t)|^N \cdot \text{signum}\{B_N(t)\} \quad (9.9)$$

$N$  is an integer ( $N = 2, 3, 4, \dots$ ) that has to be chosen by the analyst. The n-th root process weights the coherency of a signal higher than the amplitudes, which results in a distortion of the waveforms: the larger  $N$ , the less the original waveform of the signal is preserved. However, the suppression of uncorrelated noise is better than with linear beamforming.

### 9.5.2 Weighted stack methods

Schimmel and Paulssen (1997) introduced another non-linear stacking technique to enhance signals through reduction of incoherent noise, which shows a smaller waveform distortion than the n-th root process. In their method, the linear beam is weighted with the mean value of the so-called instantaneous phase of the actual signal. The phase term itself follows a power law, which can be defined by the analyst. With this phase-weighted stack all phase-incoherent signals will be suppressed and small coherent signals will be relatively enhanced.

Instead of the instantaneous phase, Kennett (2000) proposed the use of the semblance of the signal as a weighting function. He applied this approach not only on one (vertical) component of the observed wave field but also jointly on all three components taking into account the cross-semblance between the three components of ground movement. He achieved a resolution similar to what is achieved with the method of Schimmel and Paulssen (1997).

An easily implementable weighted stack method would be to weight the amplitudes of the single sites of an array with the SNR of the signal at this site before beamforming, but this does not directly exploit the coherency of the signals across the array. All described stacking methods can increase the slowness resolution of vespagrams (see Section 9.8.4).

### 9.5.3 The double beam technique

Spudich and Bostwick (1987) applied the principal of reciprocity and used a cluster of earthquakes as a source array to analyze coherent signals in the seismic coda. This idea was consequently expanded by Krüger et al. (1993) who analyzed seismic array data from well-known source locations (i.e., mostly explosion sources) with the so-called “double beam method”. Here the principle of reciprocity is used for source and receiver arrays to further increase the resolution and the SNR for small amplitude signals by combining both arrays in a single analysis.

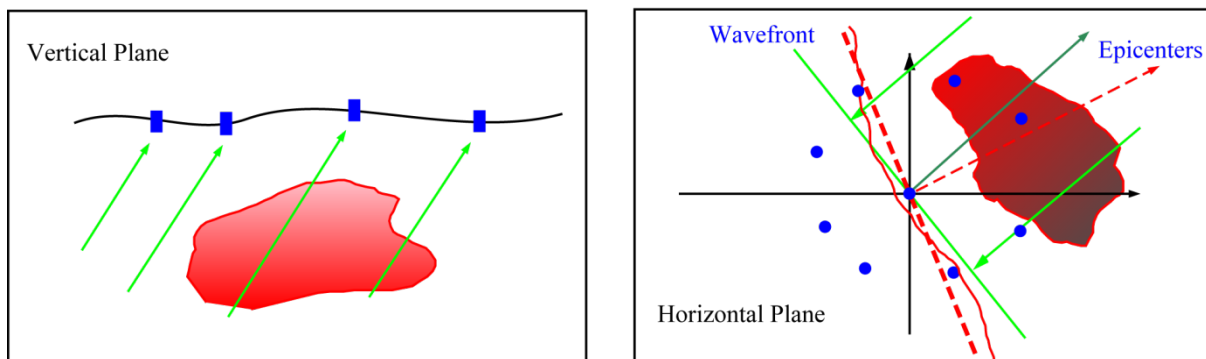
### 9.5.4 Time delay corrections

Calculating time delays using  $\tau_i = \mathbf{r}_i \cdot \mathbf{s}$  is a simplification, often ignoring both elevation differences between the instruments and the fact that seismic waves are not perfect plane waves when crossing an array. To compensate for this, we have to introduce a correction  $\Delta\tau_i$ . Calculating time delays as done in Section 9.4.4 will compensate for deviations due to elevation differences (see Fig. 9.15). However, deviations from a plane wave may also be due to lateral heterogeneities below the array and if known, time corrections may improve the conformance between observed and theoretically predicted time delays.

For the large NORSAR array (NOA) e.g., elevation corrections are not used for beamforming. Instead, a database with empirical time delay corrections  $\Delta\tau_{i,l}$  was established that corrects for both elevation differences and inhomogeneities, and this database is still in use (Berteussen, 1974). So, for all beamforming, including each point in the beampacking process (see Section 9.8.2), the delays  $\Delta\tau_i$  of Eq. (9.3) are corrected according to Eq. (9.10). Note that the corrections depend on the slowness  $\mathbf{s}_l$  of the signal.

$$\tau_i = \mathbf{r}_i \cdot \mathbf{s}_l + \Delta\tau_{i,l}. \quad (9.10)$$

**Plane Wave Deviations**



**Fig. 9.15** This figure illustrates how lateral heterogeneities below an array (reddish areas) can disturb the propagation of seismic waves. Instead of a plane wave (green line on the right figure) a curved wavefront (red line) is passing the array and then (wrongly) interpreted as a plane wave (broken red line). The theoretically expected plane wave would be the green line.

A method to determine velocity heterogeneities by inverting such deviations of observed onset times from the theoretical plane wave was developed at NORSAR, the so-called ACH method (Aki, Christoffersson, and Husebye, 1977; Christoffersson and Husebye, 2011).

## 9.6 Beamforming and detection processing

A major task in processing seismic data is that of detecting possible signals in the data samples collected at the seismometers. A “signal” is defined to be distinct from the background noise due to its amplitude, different shape, and/or frequency content; in other words, the variance of the time series is increased when a signal is present. Statistically, we can form two hypotheses: the observation is noise or the observation is a signal plus noise. As aforementioned, the seismic signal observed at different sites of an array should be more coherent than random noise. If we assume that the time series recorded are independent measurements of a zero-mean Gaussian random variable, then it can be shown that the hypothesis of the recording being noise can be tested by measuring the power within a time window. If this power exceeds a preset threshold, then the hypothesis is false, i.e., the recording is signal plus noise. In practice, the threshold cannot be calculated precisely and may vary with time as is true for the background noise. An approximation to such a detector in seismology is to estimate the signal power over a long time interval (LTA = Long Term Average), and also over a short time interval (STA = Short Term (power) Average). The ratio STA/LTA, which is usually called signal-to-noise ratio (SNR), is compared with a preset threshold. If the SNR is larger than this threshold, the status of detection is set to “true” and we are speaking about a detected seismic signal.

This kind of an STA/LTA detector, proposed by Freiburger (1963), was installed and tested to our knowledge for the first time at LASA (van der Kulk et al., 1965), and later at Yellowknife (Weichert et al., 1967) and at NORSAR (Bungum et al., 1971). For complementary details on the STA/LTA trigger algorithm and its parameter settings in general, see IS 8.1. To calculate the STA, we can use a sum of the absolute values rather than the squared values (power) for computational efficiency; the difference in performance is minimal and the results are slightly more robust. The definition of the short-term average (STA) of a seismic trace  $w(t)$  is:

$$STA(t) = \frac{1}{L} \cdot \sum_{j=0}^{L-1} |w(t-j)| \quad L = \text{sampling rate} \cdot \text{STA window length}, \quad (9.11)$$

the recursive definition of the long-term average (LTA) is:

$$LTA(t) = 2^{-\zeta} \cdot STA(t-\varepsilon) + (1 - 2^{-\zeta}) \cdot LTA(t-1), \quad (9.12)$$

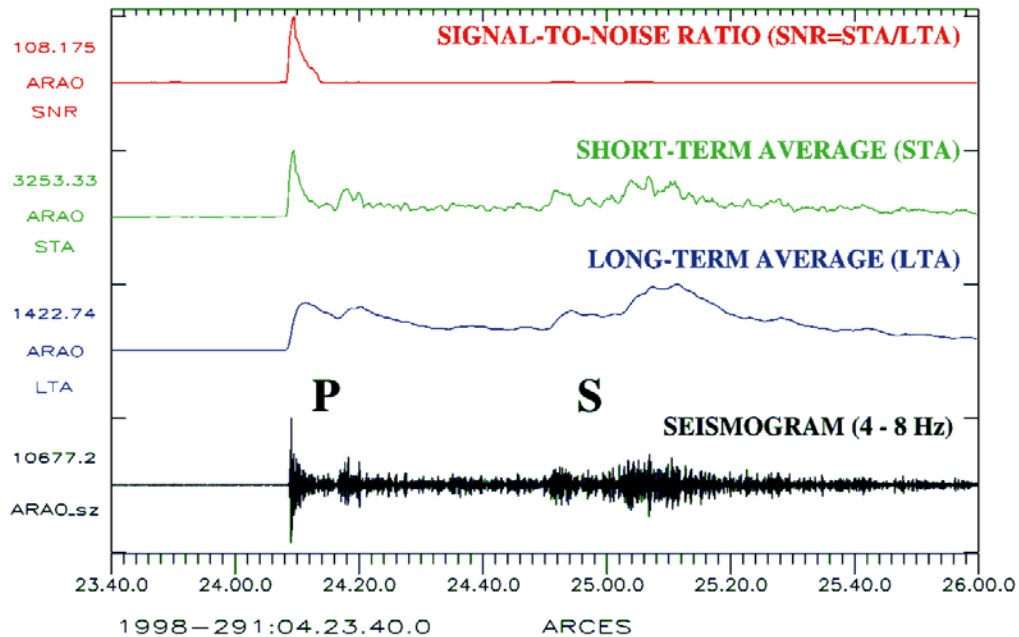
where  $\varepsilon$  is a time delay, typically a few seconds, and  $\zeta$  is a steering parameter for the LTA update rate. The parameter  $\varepsilon$  is needed to prevent a too early influence of the often-emergent signals on the LTA. In the case of a larger signal, the LTA may stay too long at a relatively high level and we will therefore have problems detecting smaller phases shortly after this large signal. Therefore, the LTA update is forced to lower the LTA values again by the exponent  $\zeta$ .

Then the signal-to-noise ratio (SNR) is defined as:

$$SNR(t) = STA(t)/LTA(t). \quad (9.13)$$

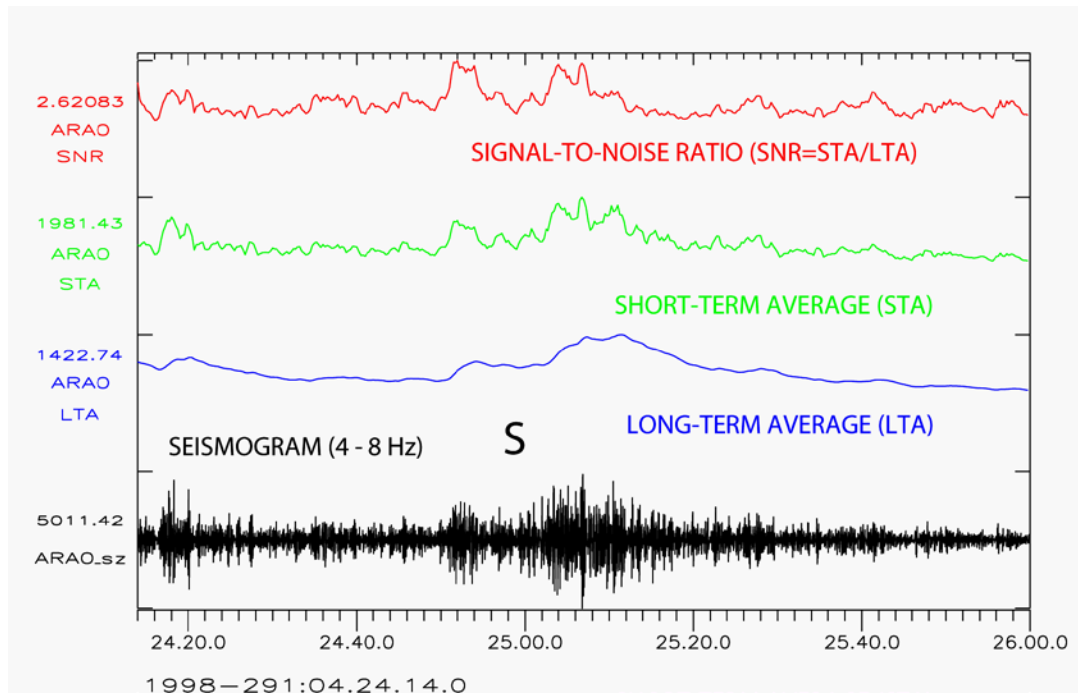
The STA/LTA operator may be used on any type of seismic signals or computed traces. That means, the input time series  $w(t)$  may be raw data, a beam, filtered data or a filtered beam.  $L$  is the number of points of the time series  $w(t)$  to be integrated. The recursive formulation for calculating the LTA means that the linear power estimate of the noise is based mainly on the last minute's noise situation, which is a very stable estimate. The influence of older noise conditions on the actual LTA value and a weighting of the newest STA value can be defined by the parameter  $\zeta$ , for which, e.g., at NORSAR, a value of about 6.0 is used for STA update rates of about 0.2 seconds. It is also advisable to implement a delay of  $\varepsilon = 3 - 5$  seconds for updating the LTA as compared to the STA. A simpler implementation is to estimate the LTA according to Eq. (9.11), but using an integration length that is 100 or 200 times longer for the LTA than for the STA. However, when detecting signals with frequencies above 1 Hz, it is also recommended that the LTA should not be updated while the SNR is above the detection threshold. This feature is easier to implement by using Eq. (9.12). Further details about the SNR and its improvement can be found in Section 4.4 of Chapter 4.

Figs. 9.16 and 9.17 demonstrate how the STA/LTA detector works for a single seismogram. The direct P onset of this regional event is sharp and detected clearly (Fig. 9.16). However, the P coda increases the background noise for later phases and the SNR of these phases becomes very small (Fig. 9.17). In this case, the advantages of using an array to detect seismic signals can be demonstrated easily. The apparent velocities of the P onsets and the S onsets are so different that calculating the corresponding S beam will decrease the P-phase energy and amplify the S-phase energy (Fig. 9.18).

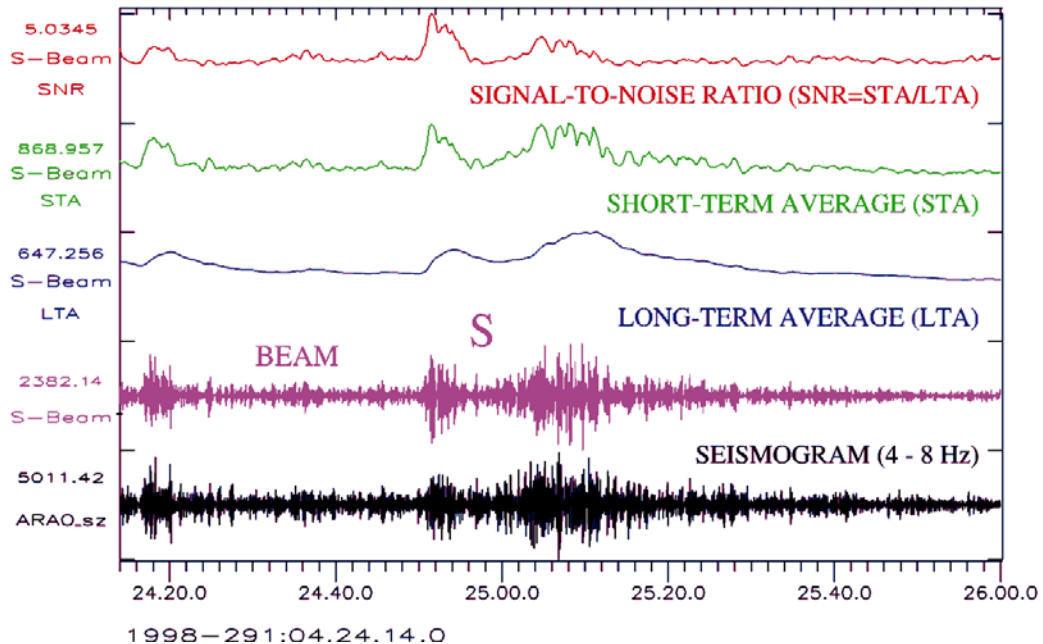


**Fig. 9.16** LTA, STA and STA/LTA (= SNR) traces for a seismogram of a regional event observed at the ARCES reference site ARA0 (bottom). The seismogram was bandpass filtered between 4 and 8 Hz. Note the sharp onset for the P phase with an SNR of 108.175.



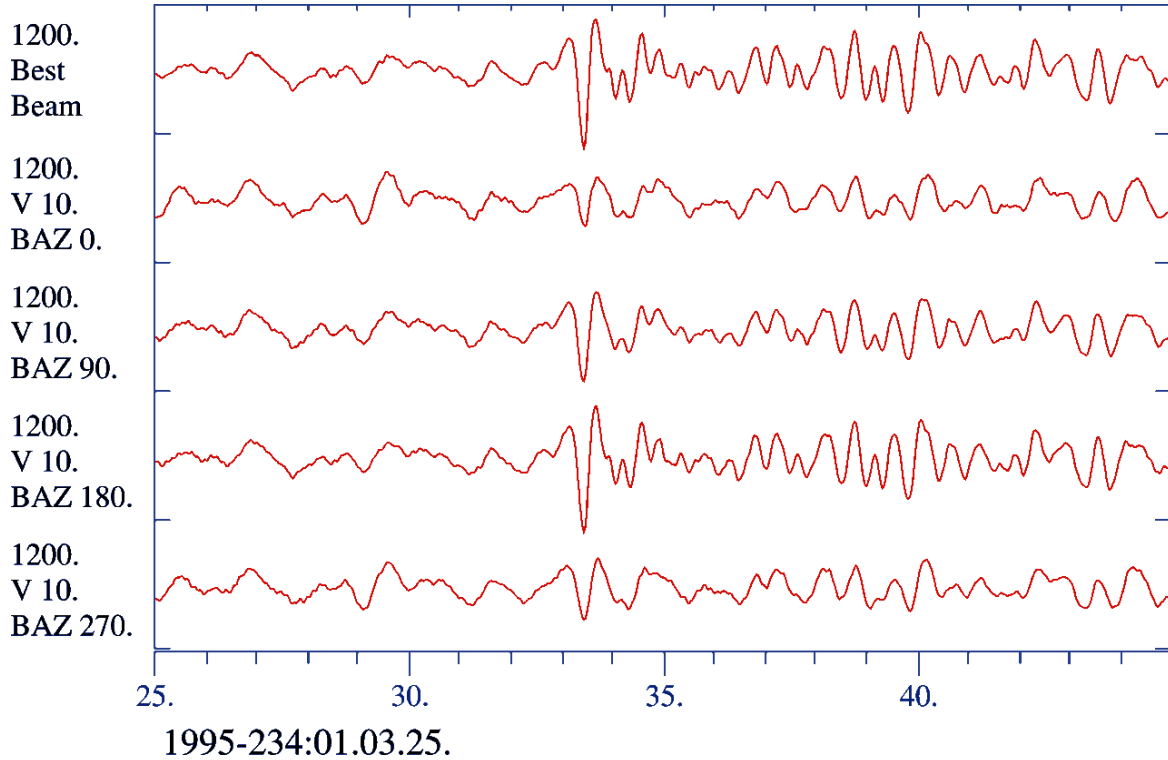


**Fig. 9.17** As Fig. 9.16, but only for the time window after the P onset. Note that due to the P coda the noise and consequently the LTA is increased. Therefore the SNR of the S-phase onsets becomes relatively small on this single vertical trace.



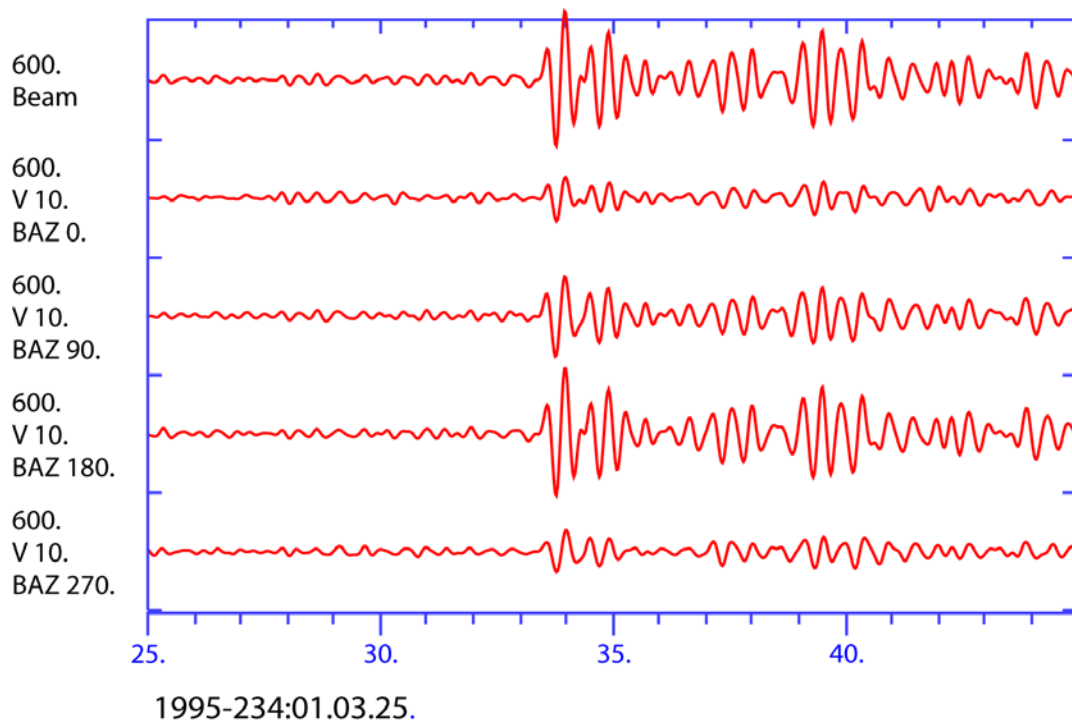
**Fig. 9.18** As Fig. 9.17, but with LTA, STA and SNR calculated for a beam optimized for the first S onset. The array beam is shown as the second trace from the bottom. Compare the relative amplitudes of the P-coda on the array beam and on the single station seismogram at ARA0, which is shown at the bottom (note the different scaling factors to the left of the traces).

In Fig. 9.19 we display again the Greek event from Fig. 9.13 with the “best” beam ( $v_{app} = 10.0$  km/s, backazimuth  $158^\circ$ ) on top, together with beams using the same apparent velocity of 10.0 km/s but different backazimuths ( $0.0^\circ$ ,  $90.0^\circ$ ,  $180.0^\circ$  and  $270.0^\circ$ ). Note the difference in amplitudes of the beams for signal and noise. Because the “best” backazimuth of  $158^\circ$  is relatively close to  $180.0^\circ$ , the top trace and the second trace from the bottom differ only slightly. Thus, Fig. 9.19 demonstrates **resolution limits** for small aperture arrays like NORES. To find the “best” beam is, in principle, a matter of forming beams with different slowness vectors and comparing the amplitudes or the power of the beams, and then finding which  $v_{app}$ -backazimuth combination gives the highest signal energy on the beam.

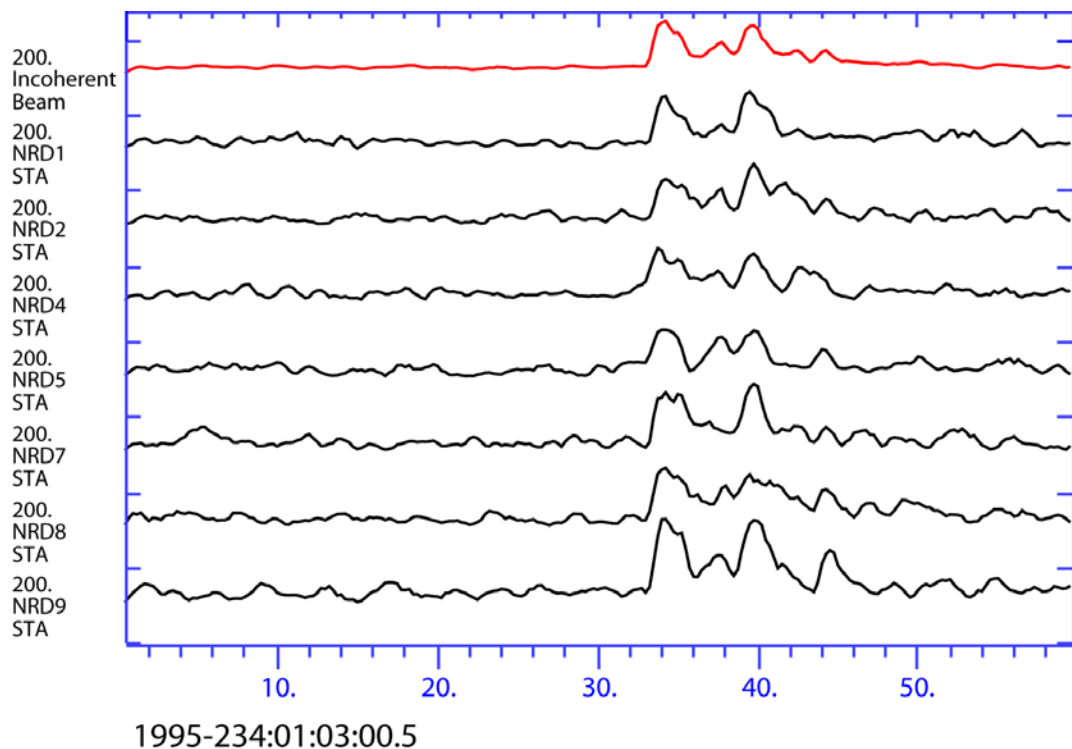


**Fig. 9.19** NORES beams for the same event as in Fig. 9.13 with different slownesses. At top we plot again the best beam of Fig. 9.13. The other four traces are NORES beams of the same data now calculated by applying a constant apparent velocity of 10 km/s and different backazimuths (BAZ). The traces are unfiltered and are plotted with an equal amplitude scale.

In Fig. 9.20, the same beams as in Fig. 9.19 are shown, but now filtered using a Butterworth 3rd order bandpass filter between 2 and 4 Hz. When beamforming using Eq. (9.5), we can either filter all the individual traces first and then beamform, or we can beamform first, and then filter the beam, which is faster by a factor of the number of sites minus one. Both procedures should theoretically give the same result because both beamforming and filtering are linear processes. However, locally increased noise levels at single sites can make it useful to filter the single traces first to avoid leakage of low frequency energy when using tools containing Fourier analysis applications.



**Fig. 9.20** This figure shows the same beams as in Fig. 9.19 but filtered with a Butterworth bandpass filter between 2 and 4 Hz. All traces have an equal amplitude scale.



**Fig. 9.21** Illustration of an incoherent beam (see text) which is shown on top in red. The other traces are STA time series. The selected NORES channels have been pre-filtered with a Butterworth bandpass filter between 2 and 4 Hz. All traces have an equal amplitude scale.

Fig. 9.21 (top trace) shows an incoherent beam, made by first filtering the raw data, then computing STA time series of each trace and afterwards, summing up the STA traces. The STA traces can be time shifted using time delays for a given slowness vector, but for detection purposes, when using a small aperture array, this is not necessary since the time shifts will be very small compared to the envelope of the signal. An incoherent beam will reduce the noise variance and can be used to detect signals that are incoherent across an array. Such incoherent signals are typically of higher frequency. When calculating incoherent beams, filtering must (!) be the first step.

## 9.7 Array transfer function

The array transfer function describes sensitivity and resolution of an array for seismic signals with different frequency contents and slownesses. When digitizing the output from a seismometer, we are sampling a seismic signal in the time domain, and to avoid aliasing effects, we need to apply an anti-aliasing filter. Similarly, when observing the wavefront of a seismic signal using an array, we obtain a spatial sampling of the ground movement. With an array, or a dense network, we are able to observe the wavenumber  $k = 2\pi/\lambda = 2\pi \cdot f \cdot s$  of this wave defined by its wavelength  $\lambda$  (or frequency  $f$ ) and its slowness  $s$ . While analog-to-digital conversion may give aliasing effects in the time domain, the spatial sampling may give aliasing effects in the wavenumber domain. Therefore the wavelength range of seismic signals, which can be investigated, and the sensitivity at different wavelengths must be estimated for a given array.

A large volume of literature exists on the theory of array characteristics, e.g., Barber (1958; 1959), Somers and Manchee (1966), Haubrich (1968), Doornbos and Husebye (1972), Harjes and Henger (1973), Harjes and Seidl (1978), Mykkeltveit et al. (1983; 1988), Johnson and Dudgeon (1993; 2002), Harjes (1990), and Wang (2002), but array theory can also be found in seismological textbooks (e.g., Capon, 1973; Aki and Richards, 1980; Bullen and Bolt, 1985; Buttkus, 2000; Kennett, 2002). We will describe in the following how the array transfer function can be estimated.

Assuming a noise and attenuation free plane-wave signal, the difference between a signal  $w$  at the reference site  $A$  and the signal  $w_n$  at any other sensor  $A_n$  is only the travel time between the arrivals at the sensors. As we know from Section 9.4, a plane wave is defined by its propagation direction and its apparent velocity, or in short by its slowness vector  $\mathbf{s}_o$ . Thus we can write:

$$w_n(t) = w(t - \mathbf{r}_n \cdot \mathbf{s}_o) .$$

Following Eq. (9.5) the best beam of an array with  $M$  sensors for a seismic signal for the slowness  $\mathbf{s}_o$  is defined as

$$b(t) = \frac{1}{M} \sum_{j=1}^M w_j(t + \mathbf{r}_j \cdot \mathbf{s}_o) = w(t) . \quad (9.14)$$

The seismic signal at sensor  $A_n$  of a plane wave for any other slowness  $\mathbf{s}$  can be written as  $w_n(t) = w(t - \mathbf{r}_n \cdot \mathbf{s})$  and the beam is given by

$$b(t) = \frac{1}{M} \sum_{j=1}^M w_j(t + \mathbf{r}_j \cdot \mathbf{s}). \quad (9.15)$$

If we calculate all time shifts for a signal with the (correct) slowness  $\mathbf{s}_o$  (Eq. (9.14)) with respect to any other slowness  $\mathbf{s}$  (Eq. (9.15)), we get the difference for the signal at site  $A_n$   $w(t + \mathbf{r}_n \cdot \mathbf{s}_o - \mathbf{r}_n \cdot \mathbf{s}) = w(t + \mathbf{r}_n \cdot (\mathbf{s}_o - \mathbf{s}))$ . The calculated beam can be written as

$$b(t) = \frac{1}{M} \sum_{j=1}^M w_j(t + \mathbf{r}_j \cdot (\mathbf{s}_o - \mathbf{s})). \quad (9.16)$$

This beam is now a function of the difference between the two slowness values  $(\mathbf{s}_o - \mathbf{s})$  and the geometry of the array  $\mathbf{r}_j$ . If the correct slowness is used, the beam calculated with Eq. (9.16) will be identical to the original signal  $w(t)$ . The seismic energy of this beam can be calculated by integrating over the squared amplitudes:

$$E(t) = \int_{-\infty}^{\infty} b^2(t) dt = \int_{-\infty}^{\infty} \left[ \frac{1}{M} \sum_{j=1}^M w_j(t + \mathbf{r}_j \cdot (\mathbf{s}_o - \mathbf{s})) \right]^2 dt. \quad (9.17)$$

This equation can be written in the frequency domain by applying Parseval's theorem and then the shifting theorem:

$$E(\omega, \mathbf{s}_o - \mathbf{s}) = \frac{1}{2\pi} \int_{-\infty}^{\infty} |\bar{w}(\omega)|^2 \cdot \left| \frac{1}{M} \sum_{j=1}^M e^{i\omega \mathbf{r}_j \cdot (\mathbf{s}_o - \mathbf{s})} \right|^2 d\omega \quad (9.18)$$

with  $\bar{w}(\omega)$  being the Fourier transform of the seismogram  $w(t)$ . Using the definition of the wavenumber vector  $\mathbf{k} = \omega \cdot \mathbf{s}$ , we can also write  $\mathbf{k}_o = \omega \cdot \mathbf{s}_o$ :

$$E(\omega, \mathbf{k}_o - \mathbf{k}) = \frac{1}{2\pi} \int_{-\infty}^{\infty} |\bar{w}(\omega)|^2 \cdot |C(\mathbf{k}_o - \mathbf{k})|^2 d\omega, \text{ where} \quad (9.19)$$

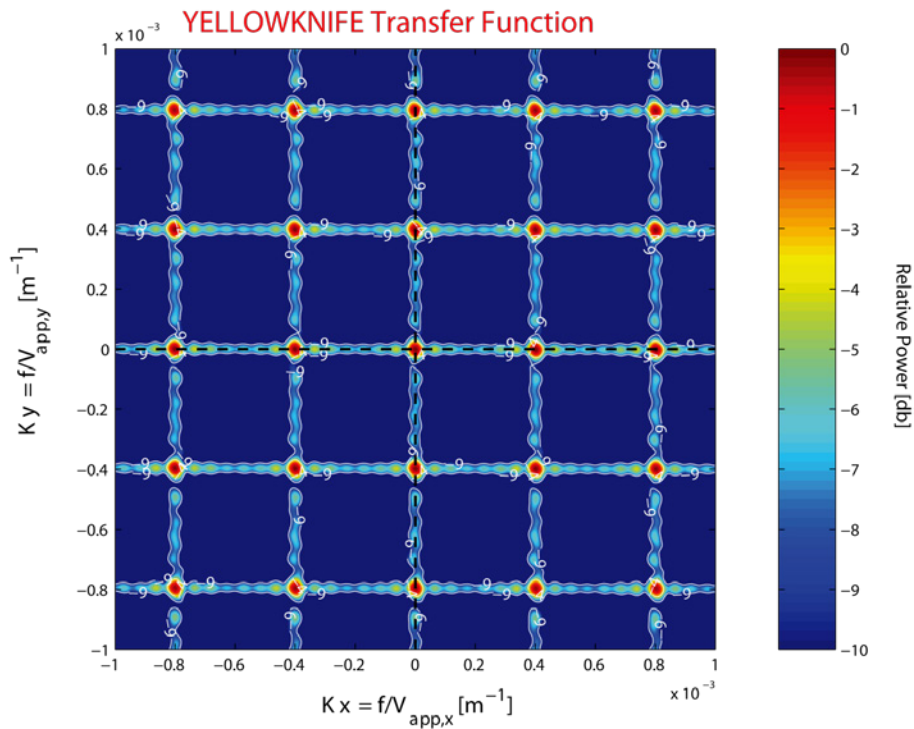
$$|C(\mathbf{k}_o - \mathbf{k})|^2 = \left| \frac{1}{M} \sum_{j=1}^M e^{i\omega \mathbf{r}_j \cdot (\mathbf{k}_o - \mathbf{k})} \right|^2. \quad (9.20)$$

Eq. (9.18) or Eq. (9.19) defines the energy of an array beam for a plane wave with the slowness  $\mathbf{s}_o$  (or  $\mathbf{k}_o$ ), but calculating the applied time shifts for a slowness  $\mathbf{s}$  (or  $\mathbf{k}$ ). If the difference between  $\mathbf{s}_o$  and  $\mathbf{s}$  (or  $\mathbf{k}_o$  and  $\mathbf{k}$ ) changes, the resulting beam has a different energy. However, this dependency is not a function of the actual signals observed at the single sites, but only a function of the array geometry weighted with the slowness difference  $\mathbf{r}_n \cdot (\mathbf{k}_o - \mathbf{k})$  (Eq. (9.19)). If the slowness difference is zero, the factor  $|C(\mathbf{k}_o - \mathbf{k})|^2$  becomes 1.0 and the array is optimally tuned for this slowness. All other energy propagating with a different slowness will be (partly) suppressed. Therefore, Eq. (9.20) is called the transfer function of an array. This function is not only dependent on the slowness of the seismic phase observed with the array, but also on the frequency, i.e., a function of the wavenumber  $k$  of the observed signal, and of the array geometry.

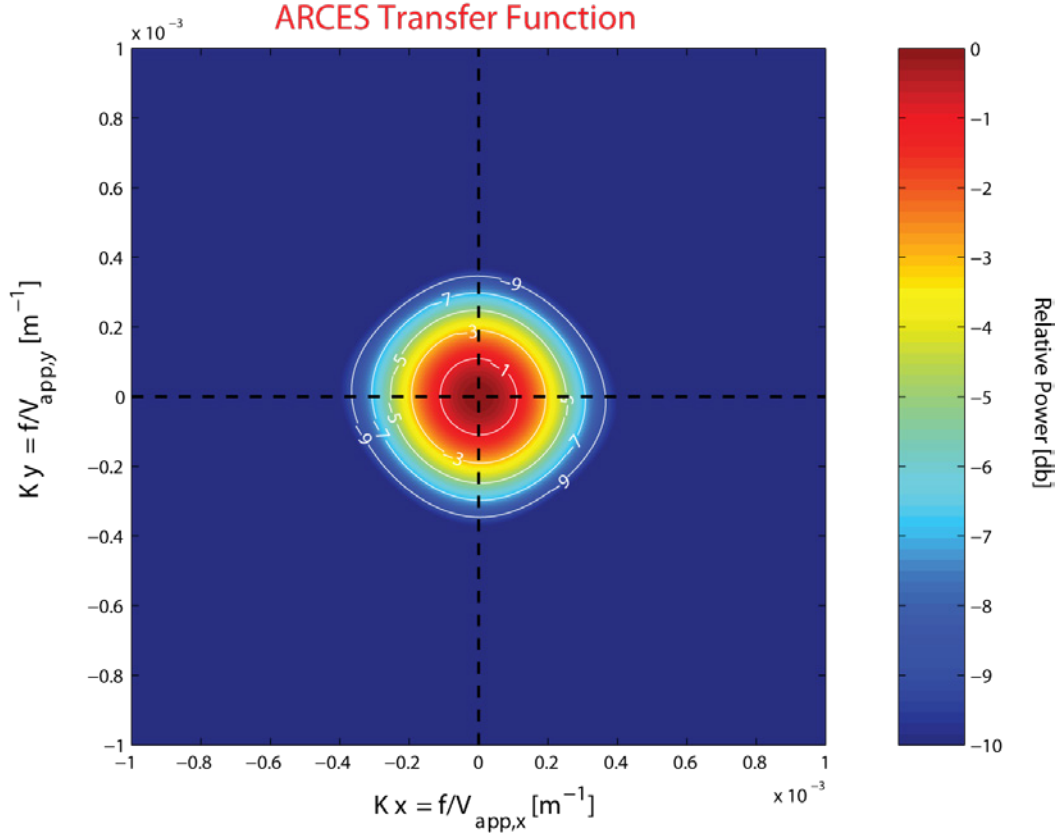
There exists literature about general criteria used to evaluate array transfer functions for a given array geometry (see e.g., Harjes and Henger, 1973; Johnson and Dudgeon 1993; 2002). Some general rules about transfer characteristics of arrays can be formulated as follows:

- 1) The aperture of an array defines the resolution of the array for small wavenumbers. The larger the aperture  $a$ , the smaller are the wavenumbers that can be measured with the array. The upper limit for the longest wavelength  $\lambda$  that can be resolved by array techniques is approximately similar to the aperture  $a$  of the array. The array behaves like a single station for signals with  $\lambda \gg a$ .
- 2) The number of sites controls the quality of the array as a wavenumber filter, i.e., its ability to suppress seismic energy crossing simultaneously with a different slowness the array than the one on which the array is steered. (see point 4).
- 3) The distances between the seismometers define the position of the side lobes of the array transfer function and the largest resolvable wavenumber: the smaller the mean distance, the smaller the wavelength of a resolvable seismic phase will be (for a given seismic velocity).
- 4) The geometry of the array defines the azimuth dependence of the aforementioned points.

Some of these points are demonstrated by two examples of array transfer functions. Fig. 9.22 shows the array transfer function of the cross-shaped Yellowknife array (YKA, see Fig. 9.1) and Fig. 9.23 shows the array transfer function of the circular, small aperture array ARCES.



**Fig. 9.22** This figure shows the array transfer function of the cross-shaped Yellowknife array (see Fig. 9.1) as relative power (color coded and with white isolines) of the array response normalized with its maximum.



**Fig. 9.23** This figure illustrates the array transfer function of the circular ARCES array (see Fig. 9.3). The figure shows relative power of the array response normalized with its maximum. White isolines were plotted at -1, -3, -5, -7, and -9 db below the maximum of the array response.

The ARCES geometry (see Fig. 9.3) gives a perfect azimuthal resolution, and the side lobes of the transfer function are far away from the main lobe. However, because of the small aperture, this array cannot distinguish between waves with small wavenumber differences, as can be seen in the relatively wide main lobe of the transfer function. In contrast, in the case of Yellowknife, the main lobe is very narrow because of the much larger aperture of the array. This results in a higher resolution in measuring apparent velocities. But the array shows resolution differences for different azimuths, which are caused by its geometry. The many side lobes of the transfer function are the effect of the larger distances between the single array sites.

Details on array design for the purpose of maximizing the gain achievable by beamforming can be found in Section 9.11.

## 9.8 Slowness estimation using seismic arrays

In the following sections we will introduce the “f-k analysis” and “beampacking” methods. Simply speaking, it is all a matter of forming beams with different slowness vectors and comparing the amplitudes or the power of the beams, and then finding out which  $v_{app}$  – backazimuth combination gives the highest energy on the beam, i.e., to find out which beam

is the “best” beam. In f-k analysis the process is done in the frequency domain rather than in the time domain. Both methods are nowadays the theoretical background for many different approaches not only in global and regional array seismology but also in local velocity structure and site effect investigations (see also Chapter 13, Section 13.4.4 and Chapter 14, Section 14.3.2.1).

### 9.8.1 Slowness estimation by f-k analysis

Frequency-wavenumber analysis – “f-k analysis” – is used as a reference tool in array processing for estimating slowness. A description of the method may be found, e.g., in Capon (1969a) or as a larger review in Capon (1973). This method has been further developed to include wide-band analysis, maximum-likelihood estimation techniques, and three-component data (Kværna and Doornbos, 1986; Kværna and Ringdal, 1986; Ødegaard et al., 1990). Theoretical aspects and information on the development of the methodology are also found in several textbooks on seismology and geophysics (Capon, 1973; Aki and Richards, 1980; Buttkus, 2000; Kennett, 2002), as well as in numerous original articles or review papers since the 1960s (e.g., Burg, 1964; Lacoss, 1965; Linville and Laster, 1966; Iyer, 1968; Liaw and McEvelly, 1979; Rost and Thomas, 2002).

The methodology exploits the deterministic, non-periodic character of seismic wave propagation to calculate the frequency-wavenumber spectrum of the signals by applying the multidimensional Fourier transform. Under these assumptions, a monochromatic, plane wave  $w(x,t)$  will propagate along the  $x$  direction according to equation:

$$w(x,t) = Ae^{i2\pi(f_0t - k_0x)}, \quad (9.21)$$

where  $A$  is the amplitude,  $f_0$  the frequency and  $k_0$  the wavenumber in the  $x$  direction. Application of the two-dimensional Fourier transform provides:

$$W(k_x, f) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} Ae^{-i2\pi[t(f-f_0) - x(k_x-k_0)]} dx dt. \quad (9.22)$$

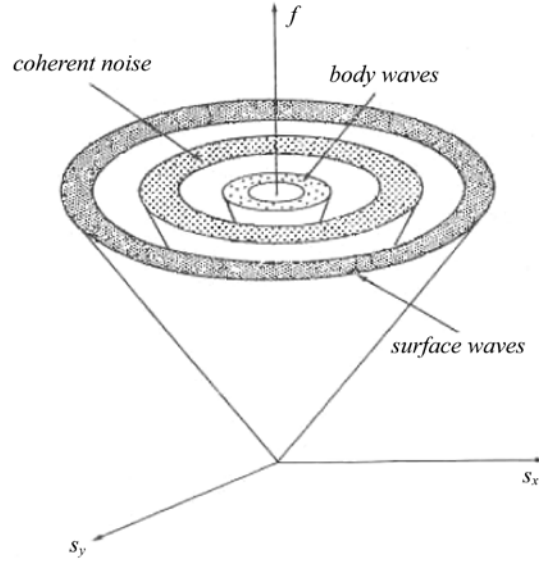
Using the delta function properties, Eq. (9.22) can be rewritten as follows:

$$W(k_x, f) = A\delta(f - f_0)\delta(k_x - k_0). \quad (9.23)$$

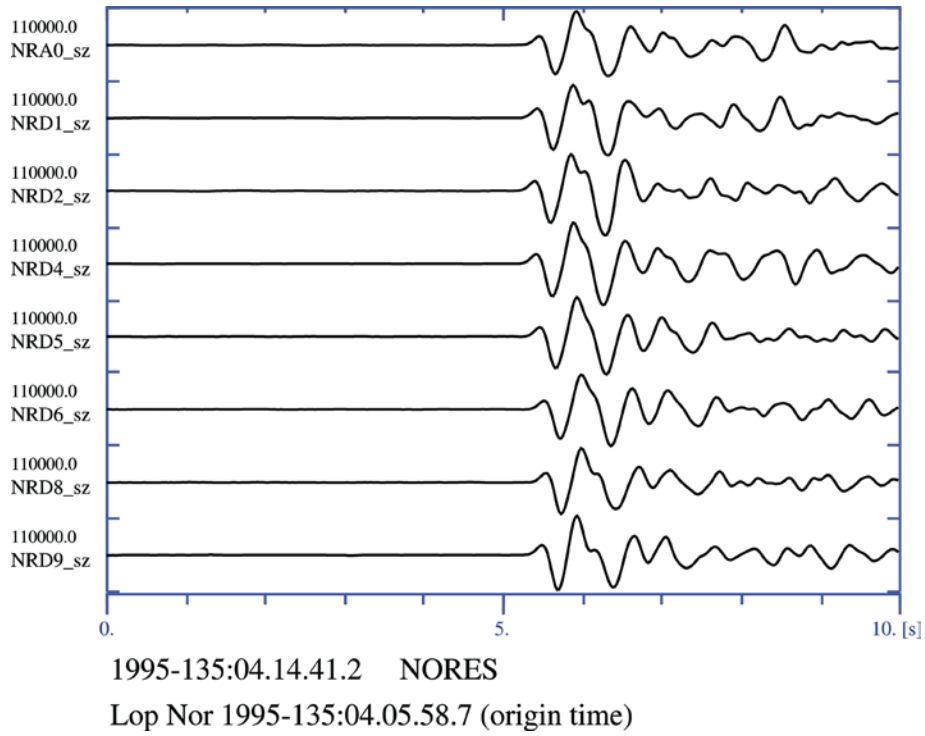
Eq. (9.23) suggests the possibility to map a plane, monochromatic wave in the frequency-wavenumber domain as a point, with coordinates  $(f, k_x) = (f_0, k_0)$ , which are nothing else than the frequency and wavenumber of the wave (e.g., Buttkus, 2000). Since the velocity of this wave is related to its wavenumber through the expression  $v_0 = \frac{f_0}{k_0}$ , it is easy to observe that

the wave’s velocity is the slope of the line connecting the point  $(f_0, k_0)$  in the frequency-wavenumber domain with the origin of the applied coordinate system  $f, k_x$ .





**Fig. 9.24** Frequency-slowness distribution of different types of waves: surface waves, body waves and coherent noise (slight modification of Fig. 7.3 from Buttkus, 2000, p. 107).



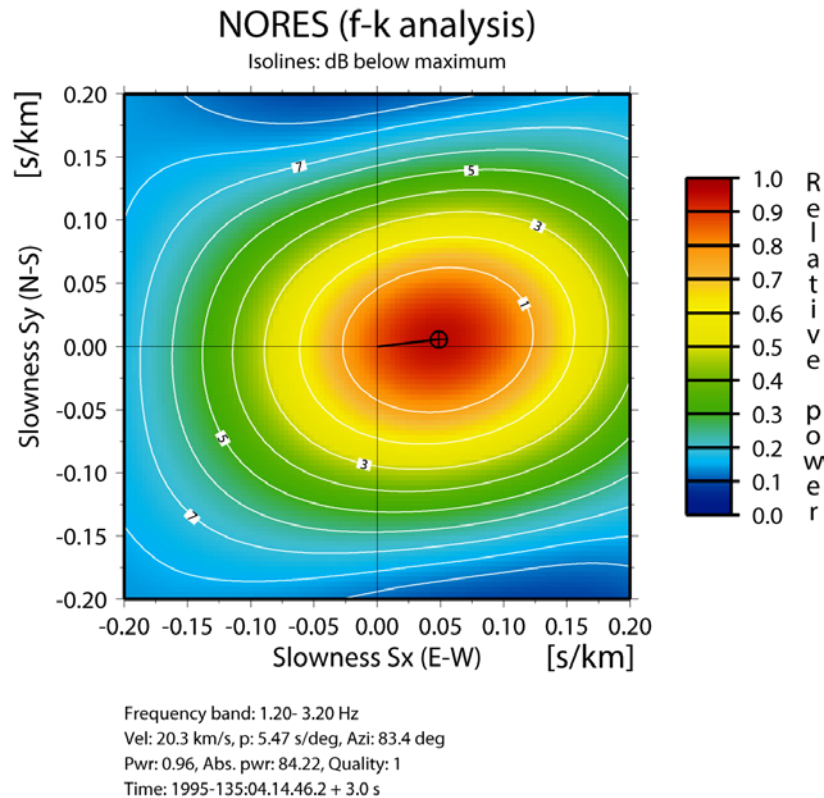
**Fig. 9.25** NORES recordings (raw data) of the Lop Nor explosion on May 15, 1995. Traces from the center site A0 and the D-ring instruments are shown to the same scale. The time scale is in [s].

Transferring our observations onto the  $k_x k_y$  plane, the angle between the  $k_y$  axis and the direction to the point  $(f_0, k_0)$  is the azimuth of the wave under discussion. Thus, measuring waves in two-dimensions, which we place on the Earth's surface, and assuming that  $k_x, k_y$  are

the wavenumber along the  $x$  and  $y$  direction respectively, provides us through f-k analysis with  $f$ ,  $k_x$ ,  $k_y$  and thereby the slowness distribution of the different types of signals. Adding frequency as a third dimension results in the formation of a ‘cone’, as in Fig. 9.24, where the different types of waves (i.e., body and surface waves, coherent noise) occupy specific concentric areas which in most cases do not overlap. In reality, the spectrum of seismic signals in its entirety exhibits a much larger complexity than the simplistic approach shown in Fig. 9.24. This is the result of a large number of factors affecting seismic signals, such as the geometry of near-surface crustal structure, the inhomogeneity of the crust, topography, etc., to mention only a few.

Practically, f-k analysis is performed in the frequency domain and represents in principle beamforming in the frequency domain for a number of different slowness values. A time shift in the time domain is equivalent to a phase shift in the frequency domain, since time is the analogue to angular frequency and space the analogue to wavenumber. Normally, at NORSAR we use slowness values between -0.4 and 0.4 s/km equally spaced over 51 by 51 points. For every one of these points the beam power is evaluated, giving an equally spaced grid of 2601 points with power information.

Such a power grid is displayed in Fig. 9.26 with the slowness ranging from -0.2 to 0.2 s/km; the unfiltered data are shown in Fig. 9.25. The power is displayed by isolines of dB down from the maximum power. The maximum power in the grid and the corresponding slowness vector define the slowness of the plane wave.



**Fig. 9.26** Result from wide-band f-k analysis of NORES data from a 3 second window around the signal shown in Fig. 9.25. The isolines are in dB from the maximum peak and the color-coded relative power is a measure of signal coherency.

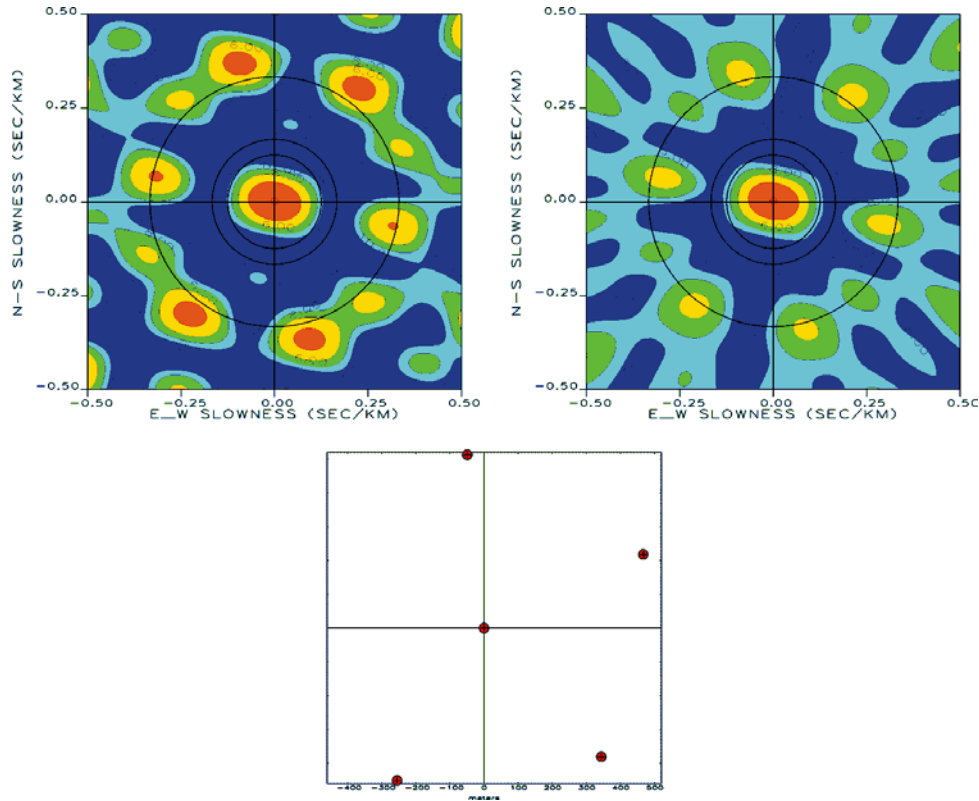
In the f-k plot in Fig. 9.26 color-coding is used to represent the relative power of the multichannel signal for 51 by 51 points in slowness space. Because the f-k analysis is a frequency-domain method, one has to define a frequency range of interest. In our case the data were analyzed in the frequency range between 1.2 and 3.2 Hz. The peak level is found at an apparent velocity of 20.3 km/s and a backazimuth of 83.4°. The normalized relative peak power is 0.96. This measure tells us how coherent the signal is between the different sites and that a beam formed with the corresponding slowness will give a signal power that is 0.96 times the average power of the individual sensors. This means that the estimated beam signal will have practically no signal loss for this slowness and in this filter band as compared to individual sensors. However, such beam loss can be taken into account when calculating station magnitudes by adding empirically derived corrections. The equivalent beam total power in Fig. 9.26 is 84.22 dB. The isolines tell us that using any different (i.e., wrong) slowness will give a signal loss of maximum 10 dB, which would correspond to an underestimation of magnitude by 0.5 magnitude units, if uncorrected.

The example in Fig. 9.26 is in a way a cross-section of the cone of Fig. 9.24 for the frequency interval 1.2 to 3.2 Hz, “focused” on the characteristics of the seismic phase selected for analysis. As already mentioned in Section 9.7, the shape of the horizontal slowness distribution is defined by the position of the elements of the recording array and the distances between them.

An uncertainty of the estimated apparent velocity and backazimuth can be derived from the size of the observed power maximum in the f-k plot at a given dB level below the maximum, the SNR of the signal, and the power difference between the maximum and any existing secondary maximum that may be present in the plot.

Fig. 9.27 displays the f-k analysis results for a 1.5 s long window of a signal arriving vertically at the array displayed at the bottom, for a narrow (left) and a wide (right) frequency range. The strong side-lobes appearing in the case when the technique is applied to a narrow frequency band ( $6.0 \pm 0.1$  Hz) are eliminated by the application over a wider range (1.9 – 10.0 Hz). Kværna and Doornbos (1986) reported on f-k analysis techniques using an integration over a wider frequency band (also called “wide-band” or “broadband” f-k analysis) rather than the single frequency-wavenumber analysis (e.g., Capon, 1969a; 1973) as applied by many authors. Fig. 9.27 demonstrates the increased stability of the wide-band f-k analysis (Kværna and Doornbos, 1986; Kværna and Ringdal, 1986) for data of adequately high SNR over the selected frequency range. The reason for this is that the positions of the side-lobes in the f-k space are frequency dependent. To the contrary, the main-lobe for different frequencies is always positioned at the same place. So, summing up f-k results for different frequencies increases the amplitude distance between the main and the side-lobes.

The result of an f-k analysis also depends on the method employed to determine the maximum power in the grid. There exists a variety of methods, such as the maximum likelihood method (Kelly and Levin, 1964; Levin, 1964; Capon et al., 1967), the maximum entropy method (Burg, 1967), the multiple signal classification (MUSIC) algorithm (Schmidt, 1986) and various adaptive algorithms for the estimation of spectral power density (e.g., Goldstein and Archuleta, 1991a; Goldstein and Chouet, 1994; Fujiwara, 1997).



**Fig. 9.27** Results from narrow- (upper left) and wide-band (upper right) f-k analysis of a 1.5 second window of a signal arriving vertically at the seismic array shown below. The narrow frequency band extends from 5.9 to 6.1 Hz, while wide-band f-k analysis was applied over the range 1.9 – 10.0 Hz. The isolines are in dB from the maximum peak and the color-coded relative power is a measure of signal coherency. The black circles represent apparent velocities of 8, 6 and 3 km/s, respectively.

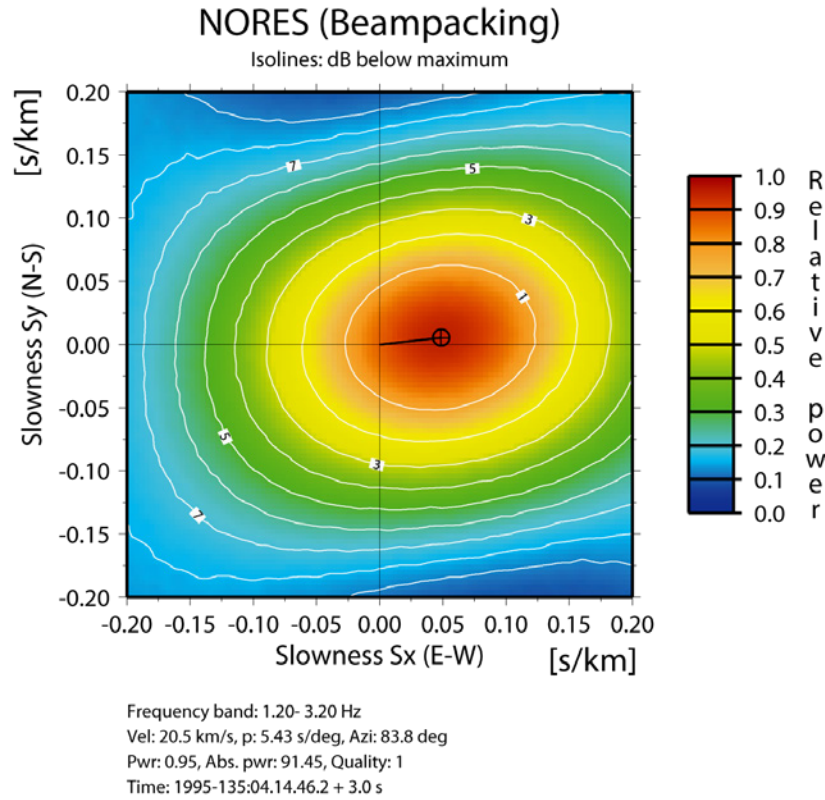
### 9.8.2 Beampacking (time domain wavenumber analysis)

An alternative to the technique described above is the so-called beampacking scheme, which had been developed at NORSAR to apply f-k analysis of regional phases to data of the large NORSAR array (NOA). This algorithm performs in the time domain beamforming over a predefined grid of slowness points and measures the power of the beam.

As an example see Fig. 9.28, where we used the same NORES data as for the f-k analysis in Fig. 9.26. All data were prefiltered with a Butterworth 1.2 – 3.2 Hz bandpass filter to make the results comparable with the f-k result in Fig. 9.26. To obtain a similar resolution as for the f-k analysis, the time domain wavenumber analysis requires a relatively high sample rate of the data. Therefore, we oversampled the data in this example 5 times by interpolation, i.e., we changed the sample rate from 40 to 200 Hz.

One can see from the beampacking process that we get practically the same slowness estimate as for the f-k analysis in the frequency domain (Fig. 9.26). In the time domain case, the relative power is the signal power of the beam for the peak slowness divided by the average sensor power in the same time window. The total power of 91.45 dB in Fig. 9.28 is the maximum beam power.

Compared to the f-k process used, the resulting total power is now 6 dB higher, which is due to the measurement method, and not a real gain. However, the beampacking process results in a slightly (about 10%) narrower peak for the maximum power as compared with f-k analysis.



**Fig. 9.28** Result from beampacking of the NORES data in Fig. 9.27 in an equispaced slowness grid. The data were prefiltered in the band 1.2 – 3.2 Hz and were resampled to 200 Hz. The isolines represent power of each beam within the 3-second window analyzed.

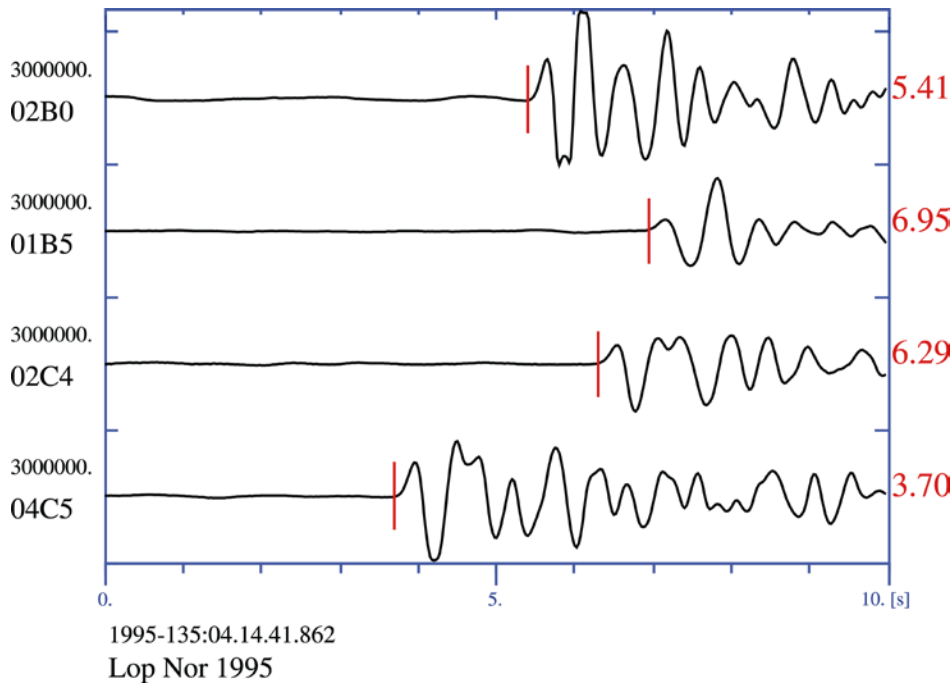
### 9.8.3 Plane wave fitting – slowness estimation by time picks

Yet another way of estimating slowness is to pick carefully times of the first onset or any other common distinguishable part of the same phase (same cycle) for all instruments in an array (see Fig. 9.29). Estimating the ray parameter and BAZ of a teleseismic P onset by fitting an assumed plane wave to a set of onset times was first applied by Abt (1907). Today, we may, e.g., use Eq. (9.24) to estimate the slowness vector  $\mathbf{s}$  by a least squares fit to the observations.

Let  $t_i$  be the arrival time picked at site  $i$ , and  $t_{ref}$  be the arrival time at the reference site, then  $\tau_i = t_i - t_{ref}$  is the observed time delay at site  $i$ . We observe the plane wave at  $M$  sites. With  $M \geq 3$ , we can estimate the horizontal components ( $s_x, s_y$ ) of the slowness vector  $\mathbf{s}$  by using e.g., a least squares fitting algorithm. If  $M \geq 4$ , the vertical component of the slowness vector ( $s_z$ ) can also be resolved. The uncertainties of the estimated parameters can be calculated simultaneously by solving the equation system of Eq. (9.24).

$$\sum_{i=1}^M (\tau_i - \mathbf{r}_i \cdot \mathbf{s})^2 = \min . \quad (9.24)$$

This method requires interactive analyst work. However, to obtain automatic time picks and thereby provide a slowness estimate automatically, techniques like cross-correlation (matched filtering) or just picking of peak amplitude within a time window (for phases that have an impulsive onset and last two or three cycles) may be used (see e.g., Del Pezzo and Giudicepietro, 2002).



**Fig. 9.29** NOA recording of the Lop Nor explosion of May 15, 1995. Vertical traces (sz) from the sites 02B0, 01B5, 02C4, and 04C5 of the NOA array (see also Fig. 9.2) are shown at the same amplitude scale. Note the large time delays as compared to the smaller NORES array in Fig. 9.27. The figure illustrates a simple time pick procedure of the individual onsets. A plane wave fit to these 4 onset-time measurements gives a backazimuth of  $77.5^\circ$  (theoretically expected  $76.1^\circ$ ) and an apparent velocity of 16.3 km/s, which is close to the theoretical value of 14.5 km/s corresponding to an epicentral distance of  $61^\circ$ .

The so-called Progressive Multichannel Cross Correlation (P.M.C.C) program package can apply inter-sensor correlation functions between single array traces to estimate BAZ and slowness of seismic signals in a continuous mode (see Cansi et al., 1993; Cansi, 1995). However, because of the amount of required computations, this method is most effective for arrays with a smaller number of sites or for subarray configurations.

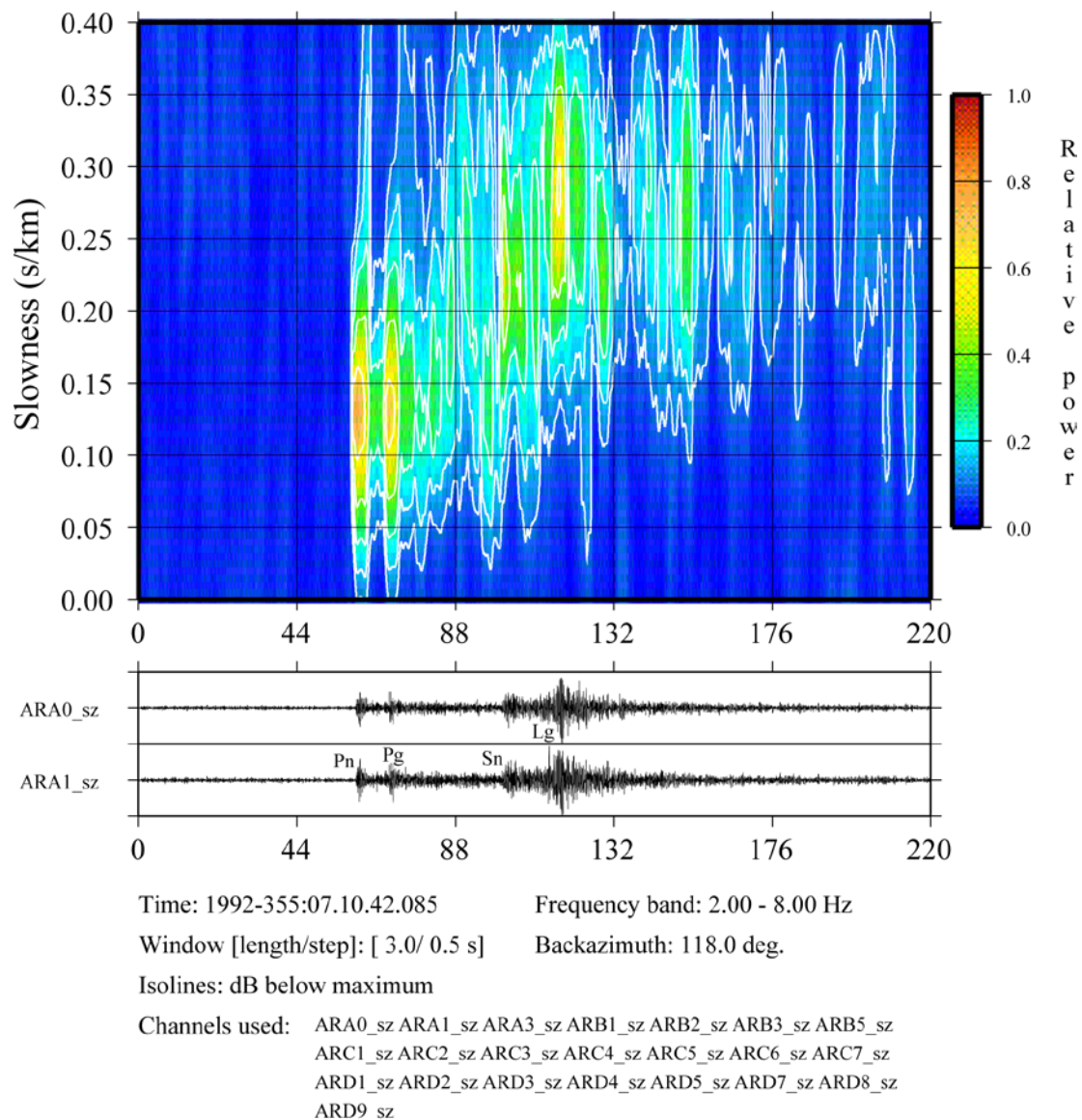
#### 9.8.4 The VESPA algorithm

A method to separate signals propagating with different apparent velocities is the VElocity SPectrum Analysis (VESPA). The principal idea of this method is to estimate the seismic energy reaching an array with different slownesses and to plot the beam energy along the time



axis. The usual way to display a vespagram is to calculate the observed energy for specific beams and to construct an isoline plot of the observed energy for the different slowness values. The original VESPA process was defined for plotting the observed energy from a specific azimuth for different apparent velocities versus time (Davies et al., 1971).

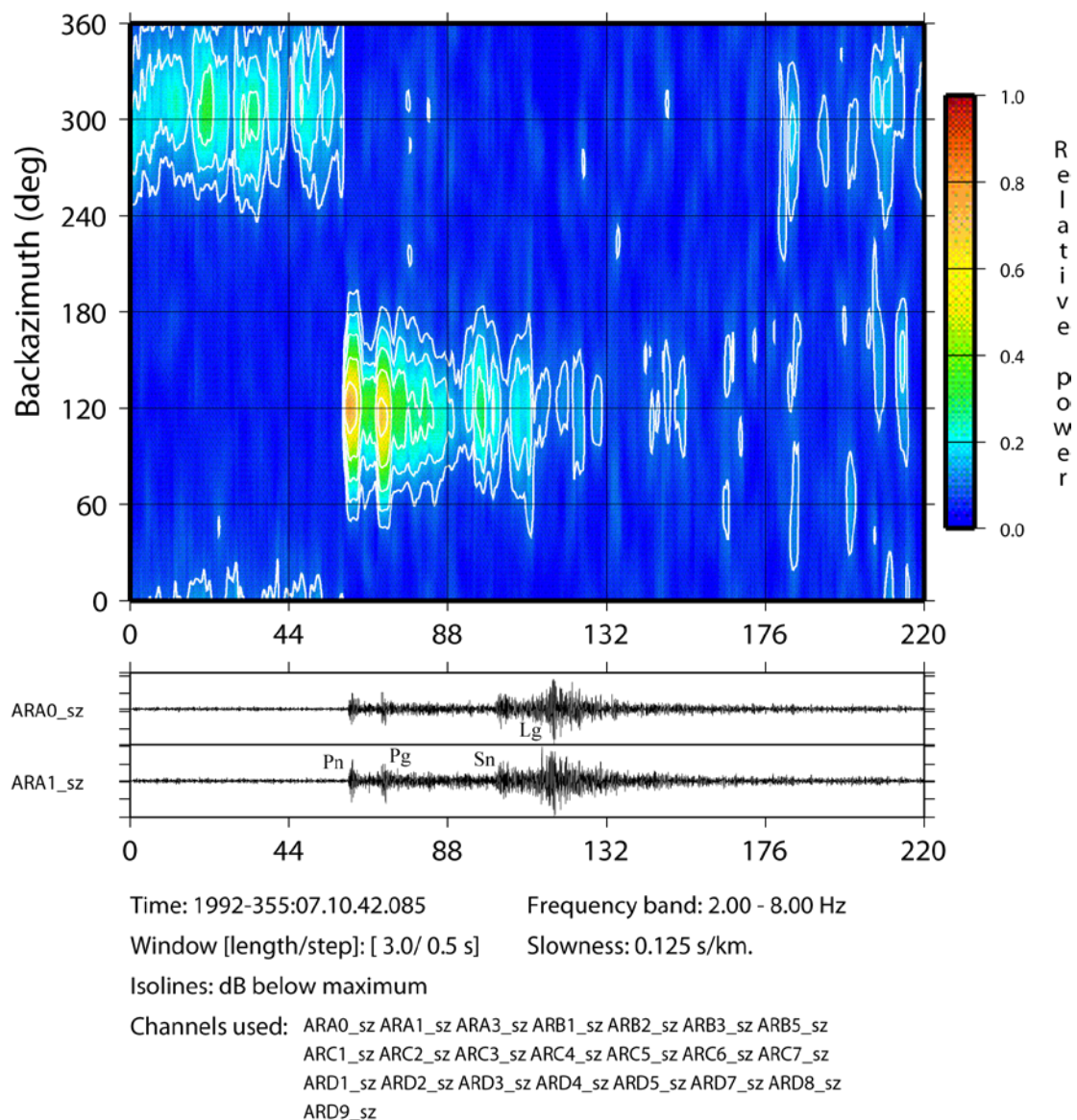
Fig. 9.30 shows as an example of a vespagram for a mine blast in the Khibiny Massif (Kola Peninsula) observed with the ARCES array. The underground blasting of about 190 tons of explosives occurred on December 21, 1992 at 07:10 (latitude 67.67°, longitude 33.73°) at about 3.55° epicentral distance from ARCES. All beams were calculated with the theoretical backazimuth of 118°, and the seismograms were bandpass filtered between 2 and 8 Hz. Fig 9.30 (middle) shows two of the filtered seismograms used to calculate the vespagram.



**Fig. 9.30** Vespagram for a mining explosion (December 21, 1992; 07:10 h; latitude 67.67° N, longitude 33.73° E) in the Khibiny Massif observed at ARCES. Shown is the observed seismic energy for different apparent velocities (slownesses) and a constant backazimuth of 118°.

The energy for the different slowness values was calculated for 3 second-long sliding windows moved forward in 0.5 s long steps. The observed energy was normalized with the maximum value and the isolines were plotted as contour lines in [dB] below this maximum. Note that the first two P onsets both have a slowness of about 0.125 s/km equivalent to an apparent velocity of 8 km/s; these are the Pn phase and a superposition of onsets from several crustal phases. The S phases are clearly separated from the P phases in the slowness space; Sn with a slowness of about 0.225 s/km (corresponding to an apparent velocity of about 4.44 km/s) and the dominating Lg phase with a slowness of about 0.28 s/km or an apparent velocity of about 3.57 km/s.

Later, the vespagram concept was expanded by plotting the observed energy from different azimuths using a specific apparent velocity. Fig. 9.31 shows an example for such a plot for the same event in the Khibiny Massif as for Fig. 9.30.



**Fig. 9.31** As Fig. 9.30 but the energy is now calculated for a constant apparent velocity of 8 km/s (i.e., a slowness of 0.125 s/km) and different backazimuths.



Instead of a constant backazimuth, a constant apparent velocity of 8 km/s was used to calculate the beam energy from all azimuth directions. Note also that the noise preceding the first P onset contains energy with apparent velocities around 8 km/s, but this noise approaches the array from a different backazimuth ( $310^\circ$ ), and the crustal P phases show a slight shift in the backazimuth direction relative to the first mantle P phase ( $P_n$ ).

### **9.8.5 The correlation method used at the UKAEA arrays**

As discussed in Section 9.7, the array transfer function of the UKAEA array YKA shows strong side lobes along a rectangular grid (see Fig. 9.21). This effect can be observed at all orthogonal linear or L-shaped arrays (Birtill and Whiteway, 1965). To improve the lower resolution along these principal axes, a correlation method has been introduced. In a first step, theoretical beams are calculated separately for each of the two seismometer lines; the geometrical crossing point of the two linear subarrays is used as the common reference point for both beams. If the actual signal has the same slowness as the slowness used to calculate the two beams, the signal must be in phase on both beams. In a second step, the calculation of the cross-correlation between the two beams tests this condition. The correlator trace is calculated for a short, moving time window. This non-linear process is very sensitive to small phase differences and improves the resolution of such arrays especially along the principal axes of their transfer functions (for further details see Whiteway, 1965; Birtill and Whiteway, 1965; Weichert et al., 1967).

### **9.8.6 Tracking earthquake source propagation with seismic arrays**

Usually earthquakes are modeled as point sources. The hypocenter of an earthquake is defined as the point where the initial rupture starts (see IS 11.1). However, for many studies the geometry of the rupture and its dynamics are of importance. Spudich and Cranswick (1984) published a study, in which they used data from a small linear accelerometer array to investigate the dynamics of the 1979 Imperial Valley, California earthquake. Similar studies were made by Goldstein and Archuleta (1991b) and Huang (2001), who used data from two very small aperture accelerometer arrays in Taiwan (SMART-1 (Abrahamson et al., 1987) and SMART-2 (Chiu et al., 1994)) to investigate rupture propagation of large Taiwan earthquakes. Common for these studies is that they all applied array analysis techniques to strong motion data from nearby strong earthquakes.

After the huge 2004 Sumatra earthquake, a slightly different approach was tested and implemented. The fault length of this event was more than 1000 km long and using data from the German network of broadband stations or the Japanese Hi-Net array, the rupture propagation of this earthquake could be monitored at teleseismic distances by applying array techniques (Krüger and Ohrnberger, 2005a and b; Ishii et al., 2005a and b). These authors measured BAZ and slowness in a moving window for times after the direct first P onset. Usually, one can observe the different seismic phases (see e.g., the VESPA algorithm in Section 9.8.4) arriving at the array. In the case of very large arrays, however, the seismograms during these time windows are dominated by the seismic energy radiated during the ongoing source process itself. The authors could show that the observed systematic change of BAZ and slowness can only be explained by the continuous seismic energy radiating rupture along the seismic fault. Later, several authors applied this technique to investigate rupture dynamics

of large earthquakes and today even an automated version for continuous monitoring of megathrust zones has been tested (Roessler et al., 2010).

### 9.8.7 Slowness corrections

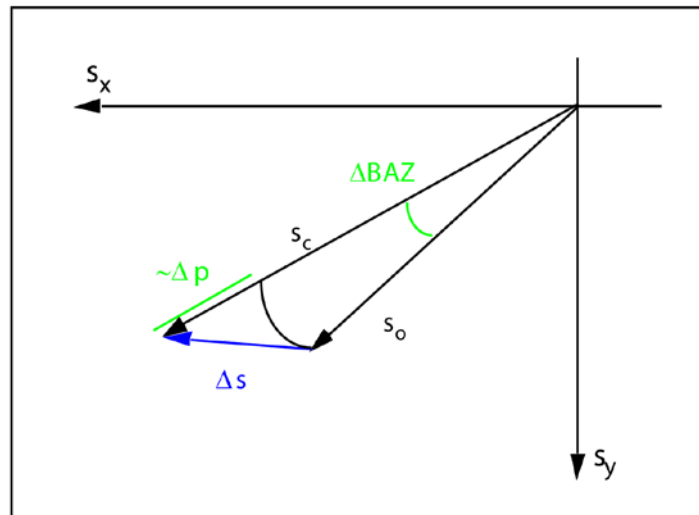
When observing the backazimuth of an approaching wave, we find deviations from the expected backazimuth. In addition, the observed ray parameter will also be different from the theoretical one. This observation is valid for any seismic station. If the deviation is systematic and consistent for a given source location (or small region), we can correct for this deviation. If the predicted slowness is  $s_c$  and the observed slowness is  $s_o$  (Fig. 9.32), then the slowness deviation is

$$\Delta s = s_o - s_c. \quad (9.25)$$

It is also common to use the ray parameter  $p$  [ $s/^\circ$ ] and the backazimuth  $BAZ$  [ $^\circ$ ] as slowness vector components and to express the residuals as:

$$\Delta p = p_o - p_c \text{ and } \Delta BAZ = BAZ_o - BAZ_c. \quad (9.26)$$

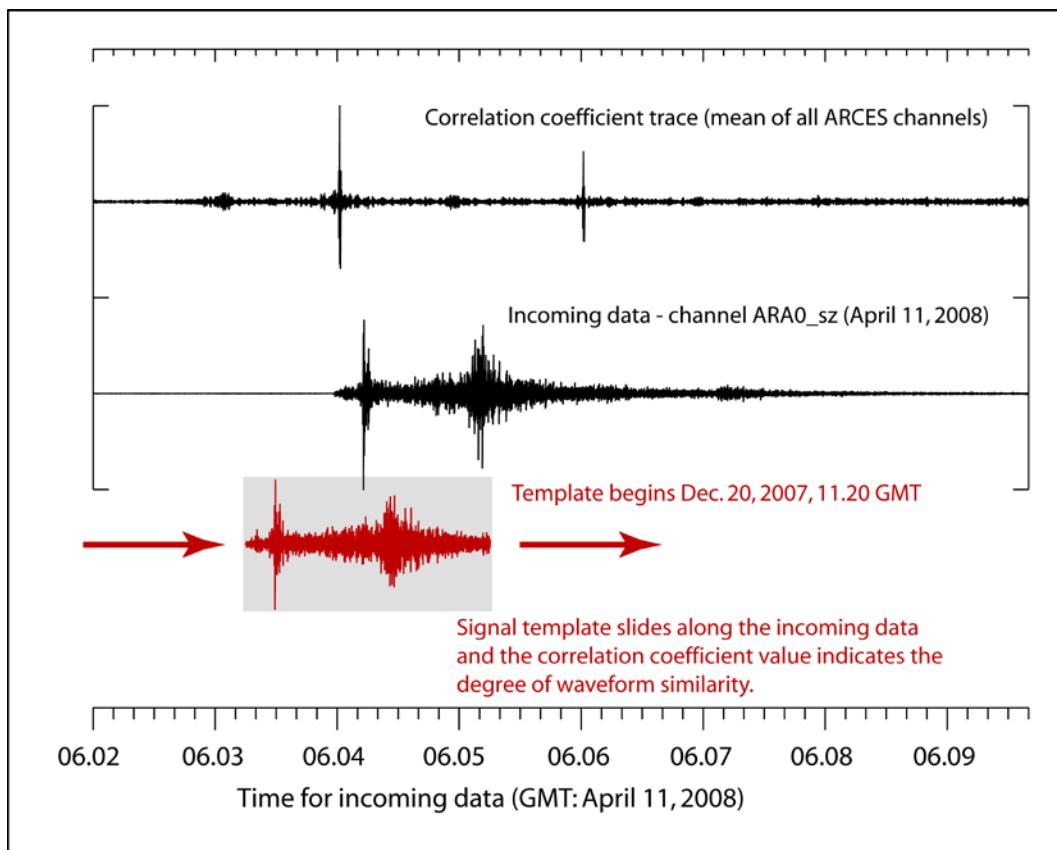
However, every array has to be calibrated with its own corrections. Numerous studies have been performed to obtain slowness corrections for different seismic arrays (see e.g., Berteussen, 1976; Krüger and Weber, 1992; Bondár et al., 1999; Kuleli et al., 2001; Schweitzer, 2001; and more recent examples of such studies are, e.g., Hao and Zheng, 2009; Tibuleac and Stroujkova, 2009; or Shen and Ritter, 2010). Usually, the derivation of slowness corrections for the whole slowness range observable with one array needs a large amount of corresponding reference data and therefore takes some time.



**Fig. 9.32** Slowness vector deviation in the horizontal plane. The vector  $s_c$  denotes the theoretical slowness. The vector  $s_o$  denotes the observed slowness. The vector  $\Delta s$  denotes the slowness residual, also referred to as the mislocation vector. The length of the slowness vector measured in [ $s/^\circ$ ] is the ray parameter  $p$ , and the angle between North and the slowness vector, measured clockwise, is the backazimuth  $BAZ$ .

## 9.9 Detection of repeating seismic events using waveform cross-correlation

In the same way that signals generated by a seismic event will resemble each other on two sensors with a sufficiently small separation, signals generated by two different seismic events will show similarity on a given sensor provided that the spatial separation of the event source locations is sufficiently small and that the event mechanisms are similar (see e.g., Geller and Mueller, 1980). This is to say that the signal observed at any seismic station is like a fingerprint for events within a given limited geographical region, as long as the seismic wave radiation pattern does not change. The degree of similarity between two segments of waveform can be measured by calculating their cross-correlation coefficient (CC). Given a signal template from a previously observed event, subsequent occurrences of that signal can be detected by calculating a continuous cross-correlation trace on the incoming data stream at the same site.

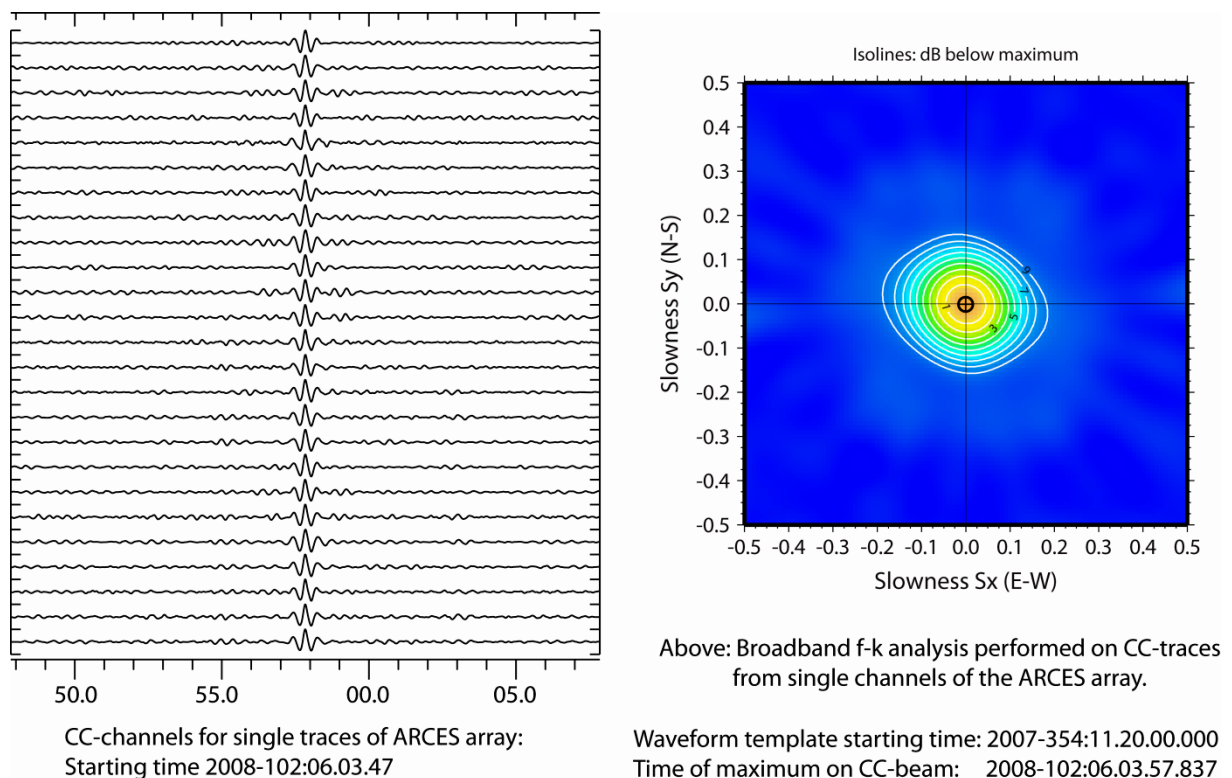


**Fig. 9.33** Illustration of detection by waveform correlation on the ARCES array of an earthquake in April 2008 using a template from an event in 2007 as a master signal (from Bungum et al., 2010). All waveforms were bandpass filtered between 2 and 8 Hz prior to performing the cross-correlation.

The principle is illustrated in Fig. 9.33. A signal template is extracted on the ARCES array for a shallow earthquake in the Steigen region of northern Norway at an epicentral distance of approximately 460 km. This template is correlated with successive segments of data and the occurrences of new earthquakes close to the site of the master event are detected by identifying times, at which the correlation coefficient is sufficiently greater than the

background level. A second peak is observed in the correlation coefficient trace approximately two minutes after the first. This indicates the presence of an aftershock whose signal is buried in the coda of the first event. This processing is done for all available array traces. If a seismic signal is present, all correlation coefficient traces should have a maximum at the same time and one can stack the correlation coefficient traces without any time delay application. The SNR on the summed correlation coefficient trace will be increased and even tiny signals hidden in the noise or other signals can be detected.

Gibbons and Ringdal (2006) describe a number of applications of the method and point out that, especially on small-aperture arrays, false alarms can occur due to a coincidental waveform similarity over a short time window. Such false detections can frequently be identified by examining the alignment of the correlation coefficient traces on the single sensors. In the case of small-aperture arrays, Gibbons and Ringdal (2006) demonstrate that the correlation traces emulate a wavefront that propagates over the array with an apparent slowness vector which is approximately equal to the difference between the dominant slowness vectors of the master and detected events. If performing wide-band f-k analysis on the cross-correlation (CC) traces indicates a non-zero slowness vector then the detected signal cannot arrive from the same direction as the master event signal and the detection can be classified automatically as a false alarm and ignored. Fig. 9.34 displays the alignment of single channel CC-traces at ARCES for the Steigen event in Fig. 9.33 together with the slowness grid from the f-k analysis.



**Fig. 9.34** Verification of the validity of a cross-correlation detection by performing f-k analysis on the individual correlation coefficient traces. A non-zero slowness vector would indicate that the signals from the master and detected events do not come from the same direction.

## 9.10 Ambient noise array analysis

### 9.10.1. Overview

During the last decades, ambient seismic vibration measurements have gained in importance, especially in regions covered with soft sediments (e.g., Aki, 1957; Horike, 1985; Tokimatsu, 1997; Okada, 2003; Foti et al., 2011). Advantages of the method are its low cost as well as the non-destructivity of measurements and thus its applicability even in densely populated areas. Whereas the presence of strong seismic noise is obstructive for active seismic investigations, it constitutes a continuous energy source for ambient seismic vibration measurements illuminating the subsurface.

One widespread application of ambient vibration measurements are microzonation studies (single station measurements analyzed with the H/V method, see e.g., Tokimatsu, 1997). Later on, the H/V method emerged as a geophysical tool to assess soil and sedimentary thickness, the analyzed depth ranging from tens of meters to more than 1000 m (e.g., Ibs von Seht and Wohlenberger, 1999; Delgado et al., 2000; Parolai et al., 2001). The H/V spectrum has even been employed to gain shear wave velocity profiles (Fäh et al., 2003; Parolai et al., 2002; Arai and Tokimatsu, 2004), but usually, ambient vibration array measurements are used to estimate the shear wave velocity structure (e.g., Asten and Henstridge, 1984; Horike, 1985; Tokimatsu, 1997; Scherbaum et al., 2003). However, the trade-off between accumulated travel times in the sediment structure and depth of the main impedance contrasts led to the development of combined inversion strategies from both techniques (Scherbaum et al., 2003; Parolai et al., 2005; Picozzi et al., 2005). A set of useful guidelines for single station and array measurements of ambient seismic vibrations was developed within the framework of the EU-project SESAME (Site Effects uSing AMbient Excitations, see e.g., SESAME group, 2004; Jongmans et al., 2005). For more details about the investigation of site response in urban areas by using earthquake data or seismic noise see Chapter 14.

A detailed review of sources of ambient vibrations can be found in Section 4.3 in Chapter 4. A discussion of noise which is specifically related to ambient vibration analysis is given in Bonnefoy-Claudet et al. (2004). At low frequencies (below 1 Hz), ambient vibrations seem to mainly consist of fundamental Rayleigh waves mostly caused by meteorological phenomena. However, at higher frequencies the noise sources are mostly caused by human activities and there is no agreement yet between authors on the ratio between body and surface waves as well as the ratio between Love and Rayleigh waves and fundamental and higher modes.

### 9.10.2 Single station/array measurements

Single station H/V analysis techniques (also known as Nakamura's technique, see Nogoshi and Igarashi, 1971; Nakamura, 1989; Bard, 1999) analyze ambient vibrations measured on horizontal and vertical seismometer components to estimate the fundamental resonance period of high impedance sediment layers overlaid by lower impedance materials (e.g., sediment over bedrock). The fundamental period is derived from the observation of a peak in the H/V spectral ratio, although the physical explanation of this peak is still under discussion, since both reverberating SH body waves as well as the ellipticity characteristics of Rayleigh surface waves may account for its existence. However, it has been demonstrated that both hypotheses explain the observations equally well in case of large impedance contrasts and simple stratigraphy (Malischewsky and Scherbaum, 2004).

Ambient vibration array techniques are logistically more demanding, but allow for the direct estimation of wave propagation characteristics, i.e., apparent wave velocity and direction of the seismic wave front. Frequency-wavenumber (f-k) and autocorrelation techniques are used to obtain the apparent phase velocity within narrow frequency bands. The resulting frequency-dependent phase velocity curves are interpreted as dispersion curve branches of surface waves and are inverted for the physical properties of 1-D Earth models parameterized as stack of horizontally homogeneous layers. Since the dispersion curve inversion constitutes a highly non-linear and non-unique problem, it is of utmost importance to obtain an accurate dispersion curve measurement and a well-founded interpretation of observations. Considering the resolution capabilities of array settings with a low number of sensors, adaptive measurement strategies have to be used to capture the full available wavelength range of the surface waves included in the ambient vibration wavefield (Bonnefoy-Claudet et al., 2005). Starting with small inter-station distance and small aperture provides a good estimate of the shallow shear wave velocity which can be used to optimize the array configuration for redeployment with larger apertures and inter-station distances. Thus, the array configuration is iteratively adapted to the next larger target wavelength. A proposed work flow can be found, e.g., in Jongmans et al. (2005). Further, a priori information (e.g., shallow shear wave velocities derived from hammer blow refraction seismic, borehole stratigraphies) may help to reduce the non-uniqueness of the inverse problem.

### **9.10.3 Array geometry for ambient vibration measurements**

The optimal array geometry for ambient vibration measurements has been discussed by e.g., Asten and Henstridge (1984) and Kind (2002). Whereas small aperture arrays designed for monitoring natural or man-made seismic sources are optimized for (usually undispersed) broadband transient signal arrivals from relatively distant sources (see also Section 9.7), arrays used for ambient vibration measurements are supposed to record a mostly random wavefield caused by nearby superficial sources, which is analyzed within narrow frequency bands (Jongmans et al., 2005). The resolution of narrowband array responses is significantly reduced compared to wide-band array responses and aliasing peaks are fully developed due to the lack of superpositioning of beam patterns for different frequencies (Jongmans et al., 2005; see also Fig. 9.24). The resolution can be enhanced by using either a large number of sensors or choosing the array size according to the specific narrow wavelength range of interest, which means re-deployment of several array configurations for different frequency ranges as mentioned in Section 9.10.2. For ambient vibration measurements, Jongmans et al. (2005) recommend a circular array deployment with an odd number of stations (see also Section 9.12) since this configuration is known to have the best possible azimuthal suppression capability for a given number of sensors and allows for analysis using f-k as well as autocorrelation methods.

### **9.10.4 Estimation of phase velocity curves**

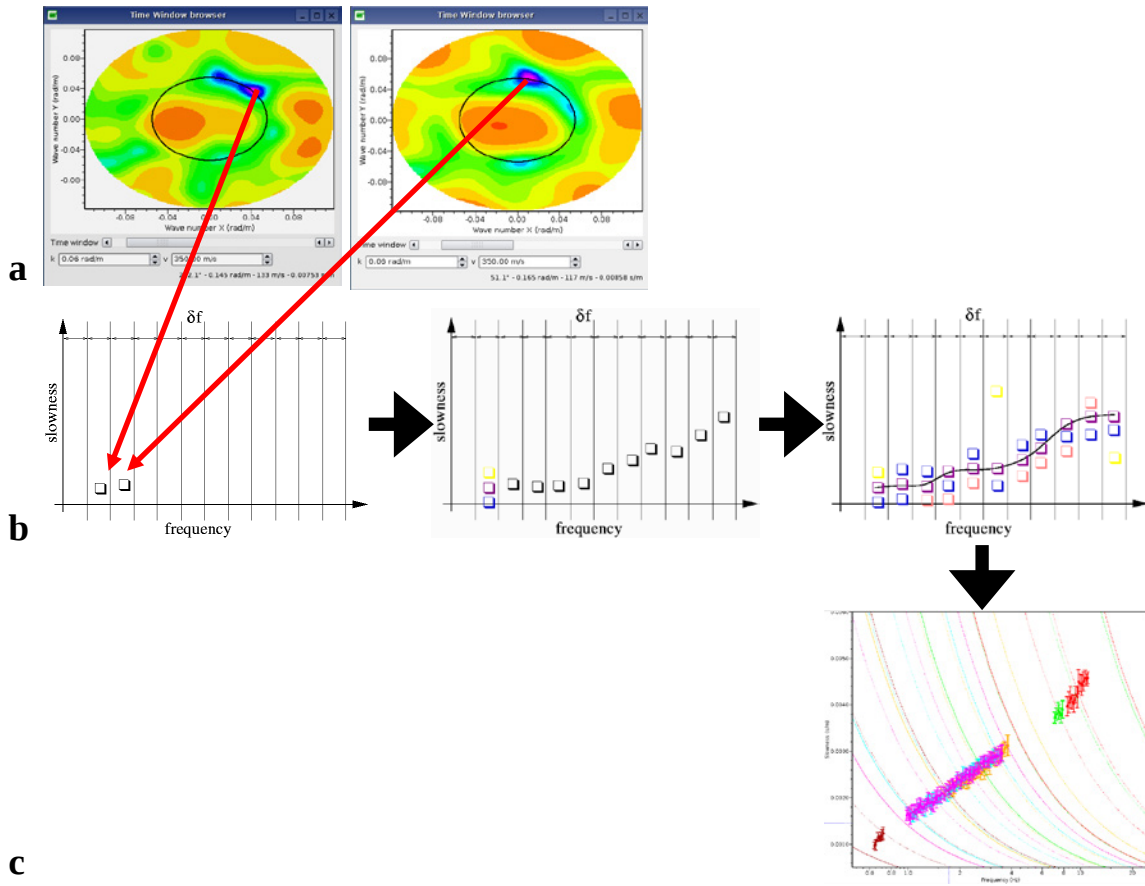
The ability to extract phase velocity curves from ambient vibration measurements within specific frequency bands is inherently coupled with the propagation characteristics at the site and the nature of ambient vibration sources on the one hand, and with experimental conditions and data analysis procedures on the other hand (Jongmans et al., 2005). In principle, either Rayleigh or Love wave phase velocity curves can be produced, but often, only Rayleigh-wave

dispersion curves are analyzed (see, e.g., Okada, 2003, for a general discussion about three-component ambient vibration analysis).

Data processing methods can be classified in two basic types: f-k methods and autocorrelation methods. The first class of methods comprises the conventional semblance based frequency-wavenumber method (CVFK) after Kværna and Ringdal (1986; see also Section 9.8), which is implemented using sliding time windows and narrow frequency bands around a center frequency (see, e.g., Dietrich et al., 2001). Capon (1969a) formulated a high-resolution f-k approach, which optimizes the windowing function based on the evaluation of the cross spectral matrix (CSM matrix) such that the wavenumber response approaches a delta-function in the f-k domain. However, the gain in slowness resolution is traded against a lack of time resolution caused by the estimation of the cross spectral matrix ("block averaging") and its inversion. Since the construction of the windowing function involves the signal itself, the response cannot be independently studied from the signal. Only a lower limit of slowness resolution can be derived from the beam pattern (Kind, 2002). Another f-k analysis method is the MUSIC algorithm (Multiple Signal Classification, Schmidt, 1986; Schmidt and Franks, 1986) decomposing the CSM matrix into signal and noise subspaces.

For an overview on the extraction process of dispersion curves from ambient vibration data, see Fig. 9.35. In order to extract the most coherent plane wave arrival, a grid search is performed in the wavenumber plane. For every frequency band of interest and every time window, a slowness map is computed. For each slowness map, the maximum is extracted resulting in the propagation parameters (azimuth and slowness as a function of frequency). These propagation parameters along with the beampower value are analyzed as histograms for their distributional characteristics, and the mean curve and variance are estimated. Such a dispersion curve is computed for each sub-array. Dispersion curves are subsequently combined within the resolution limits of the respective arrays (for a more detailed description, see, e.g., Bonnefoy-Claudet et al., 2005).

Aki (1957) developed a spatial autocorrelation method (SPAC). The spatial autocorrelation function describes the similarity between seismograms recorded by station pairs with similar interstation distance. By filtering with a narrow passband before correlation, the frequency-dependence of the spatial autocorrelation function can be investigated. Assuming a stochastic wavefield that is stationary in time and space, the spatial autocorrelation function depends directly on the phase velocity (Aki, 1957). By azimuthal averaging of the spatial autocorrelation function (averaging for station pairs having equal inter-station distances, but different azimuths), the directionally independent autocorrelation coefficients can be expressed by using Bessel functions. In the same manner as for the f-k method, the analysis is performed with sliding time windows. For each time window, the spatially averaged autocorrelation curve is computed and subsequently, the mean curve and variance are calculated over all time windows. The autocorrelation curve can either be inverted directly for the shear-wave velocity profile or transformed into a dispersion curve first. Since several station pairs with equal inter-station distance are required for azimuthal averaging, the array configuration has to be circular (Aki, 1957) or at least regular (extended spatial autocorrelation (ESPAC), see Ohori et al., 2002). Bettig et al. (2001) modified the SPAC method (MSPAC) for application to less ideal array configurations by grouping the seismic co-array (Haubrich, 1968) into rings of finite thickness introducing a radial integration corresponding to an average over all autocorrelation coefficients within a particular ring in addition to the azimuthal integration.



**Fig. 9.35** How to extract dispersion curves using f-k methods: a) compute slowness map for every frequency band of interest and every time window, pick maximum; b) insert slowness values into frequency-slowness plot, one slowness value is extracted for every frequency band and every time window, and a mean curve including variance is computed; c) a dispersion curve is extracted for every sub-array, these are combined within their respective resolution limits (given by colored lines) to give the final dispersion curve (figures partly produced using geopsy/dinver software; Wathelet et al., 2004; Wathelet, 2008).

### 9.10.5 Inversion of dispersion curves to obtain shear wave velocity profiles

A software package readily available to deal with the determination of dispersion curves from ambient vibration array recordings and their inversion is the *geopsy/dinver* software (Wathelet et al., 2004; Wathelet, 2008; see Fig. 9.36 for an example). The *geopsy* software package implements f-k methods (CVFK and a so-called high resolution f-k algorithm (HRFK)) as well as autocorrelation methods (MSPAC) following the implementation in the CAP software developed within the framework of the SESAME project. Within the *dinver* software, the inversion of dispersion curves is performed using a neighbourhood algorithm (Sambridge 1999a and b) linked to a revised version of the forward computation by Herrmann (1987).

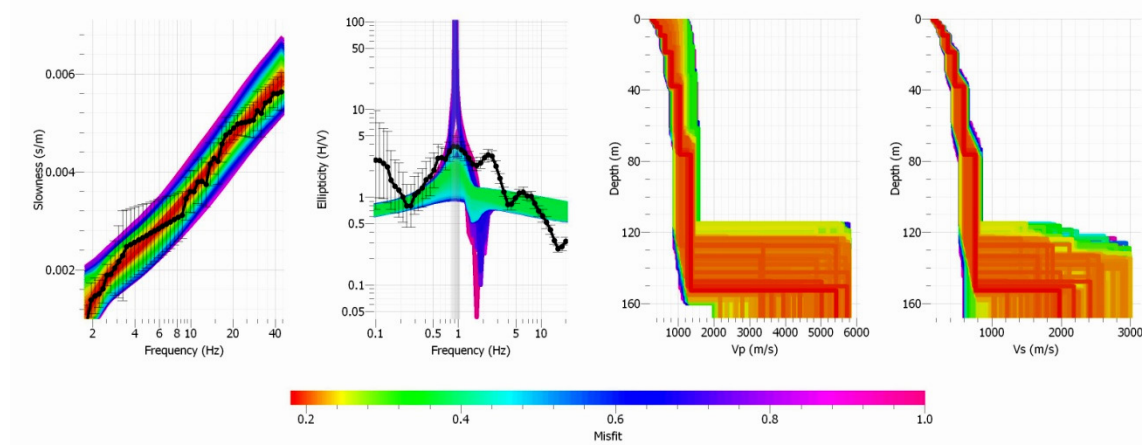
### 9.10.6 Difficulties

So far, only little work has been done on the reliability of the dispersion curve estimate and the dispersion curve inversion (for results of an international blind test see Cornou et al.,



2006). In general, dispersion curves derived from ambient vibration array analysis are reproducible and remain temporally stable although the sources exciting the wavefield may change. This stability is a consequence of the randomness of the source distribution and the attenuation properties of the medium (Bonnefoy-Claudet et al., 2005). However, the main assumptions on which the data inversion is based, namely that the subsurface is one-dimensional, i.e., horizontally stratified, and that ambient vibrations consist mainly of surface waves may not be a reasonable approximation under the given conditions (Jongmans et al., 2005).

One of the main difficulties in the analysis of ambient vibrations results from the lack of knowledge on the nature of the wavefield, e.g., spatio-temporal distribution of sources, source types, source depths, type and energy content of the radiated waves (Jongmans et al., 2005). The f-k techniques are best suited for the analysis of ambient vibrations resulting from a single source or an azimuthally constrained dominant source region, whereas for the MSPAC analysis, a random source distribution is most favorable. Thus, complementary information of both techniques can be used to enhance phase velocity estimates.



**Fig. 9.36** Inversion of dispersion curves; the four frames from left to right: 1) black line: mean dispersion curve with error bars, colored lines: forward computation of dispersion curves for ensemble of final models colored according to misfit; 2) black line: H/V ratio at central array station, colored lines: forward computation of H/V ratios for ensemble of final models; 3) P-wave velocity profile for ensemble of final models; 4) S-wave velocity profile for ensemble of final models, note high uncertainty of both P- and S-wave velocities below strong impedance contrast (figures produced using the dinver software; Wathelet et al., 2004; Wathelet, 2008).

Uncertainties in the phase velocity estimates may arise due to data recording, data processing, and the nature of the ambient vibration wavefield (Jongmans et al., 2005; Bonnefoy-Claudet et al., 2005). The main source of uncertainties is a relative arrival time error between sensor pairs due to instrumentation, phase determination, and violation of the plane wave signal. Time shifts may be either systematic or random (Jongmans et al., 2005). Systematic time shifts are caused by, e.g., unlocked GPS time synchronization, internal clock drifts of digitizers, station positioning errors, differences between phase delays of individual sensors caused by variations of seismometer constants and can be prevented by using only well-calibrated sensors in a sufficient number (at least 5, see Jongmans et al., 2005) having a corner frequency well below the frequency band of interest. For ambient vibration

measurements, the magnitude of random time shifts depends on the energy ratio between coherent versus non-coherent wavefield components (analogue to the SNR). The influence of random time shifts can be reduced by using long analysis windows (25-50 times the center period) in a large number. Additionally, in case of curved wavefronts, predictable time delays will distort the slowness estimate (Almendros et al., 1999) which means that nearby sources must be avoided ( $d/r < 2.5$ ; Jongmans et al., 2005).

Although wavefronts arriving at an array with aperture  $A$  from sources located at distances  $d \gg A$  are nearly plane as compared to the more curved wavefronts from closer sources (the presence of plane wavefronts is a basic assumption for the  $f$ - $k$  analysis), phase velocity curves derived from distant sources may show poor quality due to a lack of coherent energy attributed to the intrinsic attenuation properties of the medium (Jongmans et al., 2005). In addition, for strongly attenuating structures, higher mode energy contributions may dominate for certain frequency bands resulting in jumps of the estimated phase velocity between fundamental and higher mode branches, and intermediate phase velocities for comparable energy levels between different modes of propagation (for description and treatment of "apparent phase velocities", see Tokimatsu et al., 1992; Tokimatsu, 1997).

At sites having a sharp impedance contrast in the subsurface, spectral energy holes in the waveform spectra of the vertical component can be observed at least around the resonance frequency due to the filter characteristics of the subsurface hindering or impeding the determination of the dispersion curve in this frequency range (Scherbaum et al., 2003; Jongmans et al., 2005).

## 9.11 Array design for the purpose of maximizing the SNR gain

Signal detection at array stations is governed by the gain that can be achieved in the signal-to-noise ratio (SNR) through the process of beamforming. This subsection provides some guidance as to how an array can be designed to maximize this gain. Other aspects of array design (e.g., slowness resolution) have been dealt with elsewhere in this chapter.

### 9.11.1 The gain formula

The SNR gain  $G$  by beamforming achievable from seismic array data can be expressed by

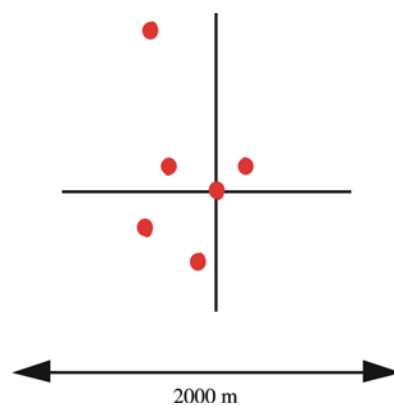
$$G^2 = \frac{\sum_{ij} C_{ij}}{\sum_{ij} \rho_{ij}} \quad (9.27)$$

where  $C_{ij}$  is the signal cross-correlation between sensors  $i$  and  $j$  of an array and  $\rho_{ij}$  is the noise cross-correlation between sensors  $i$  and  $j$  (see also Section 9.4.5). For an array with  $M$  sensors, this formula collapses to the well-known relation  $G^2 = M$  for perfectly correlating signals ( $C_{ij} = 1$  for all  $i$  and  $j$ ) and uncorrelated noise ( $\rho_{ij} = 0$  for  $i \neq j$  and  $\rho_{ij} = 1$  for  $i = j$ ).

For any array geometry it is thus possible to predict the array gain if the signal and noise cross-correlations are known for all pairs of sensors of the array layout. The remainder of this subsection describes how to design an array based on the availability of such correlation data.

### 9.11.2 Collection of correlation data during site surveys

Correlation data for use in the design phase should be collected in a carefully planned site survey. The sensor layout during the survey should be planned so as to represent as many intersensor distances as possible. The first layout for the experiments eventually leading to the deployment of the 25-element NORES array in Norway in 1984 utilized only 6 sensors, in a rather irregular geometry, as shown in Fig. 9.37.



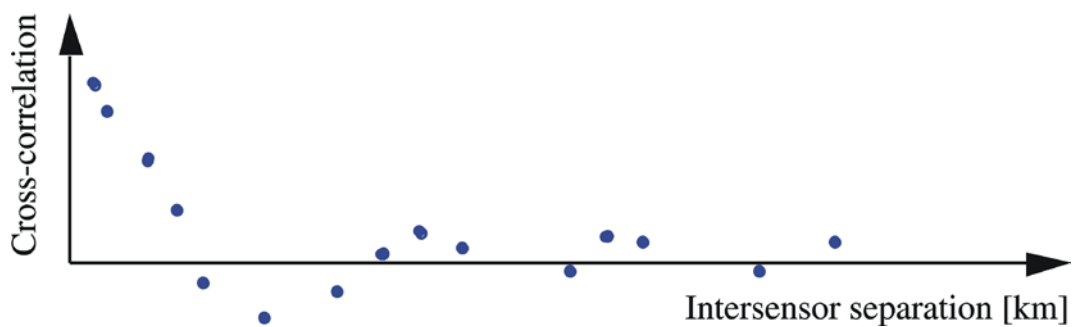
**Fig. 9.37** The first array layout for the experiments leading to the design of the 25-element NORES array in Norway in 1984.

The deployment for the collection of correlation data should be done in a way as simple as possible and should take advantage of outcropping bedrock where possible. The layout should preferably comprise ten sensors or more. If, however, for example only six sensors are available for the site survey, one could start out with a configuration something like that in Fig. 9.37 and record data continuously for about one week. At the end of this one-week period, one could redeploy four of the sensors and record data for another week. Two of the sensors would then occupy the same locations for the entire two-week recording period and would provide evidence (or lack thereof) of consistency in the results between the two one-week periods. The largest intersensor separation represented in these data should be, if possible, of the order of 3 km.

The experience from the design of the NORES array showed that the signal and noise correlation curves obtained from the early experiments (with six, and later twelve sensors) possessed most of the characteristic features (Mykkeltveit, 1985) and thus qualitatively resembled the curves derived later on from configurations comprising many more sensors (up to 25).

### 9.11.3 Correlation curves derived from experimental data

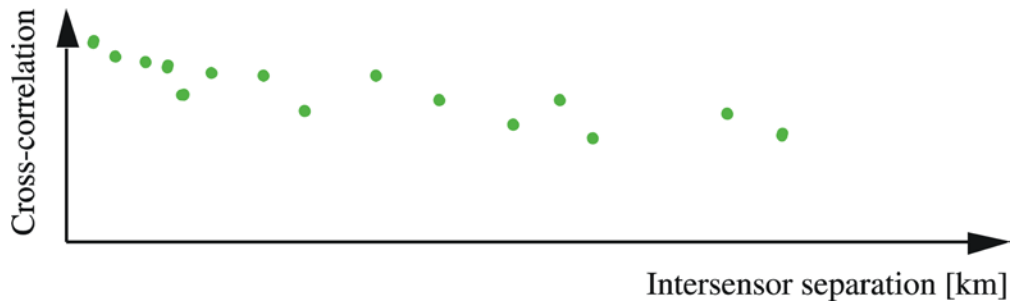
In the processing of the data from the site survey, cross-correlation values must be computed for all combinations of sensor pairs of the experimental layout. Consider, for example, a geometry of six sensors. This geometry comprises 15 unique pairs of sensors. Consider also a short interval of say 30 seconds of noise data (make sure no signal is contained in this time window) and compute the cross-correlation values for each of the 15 unique pairs of sensors (no time shifts are to be introduced for this computation). The time series are first bandpass filtered so as to derive the correlation values of one particular frequency (or frequency band). The 15 correlation values are then plotted in an x-y diagram, where the x-axis represents the intersensor separation and the y-axis the correlation value (between -1.0 and +1.0), resulting in a plot as shown in Fig. 9.38.



**Fig. 9.38** Noise cross-correlation values for a test layout of 6 sensors.

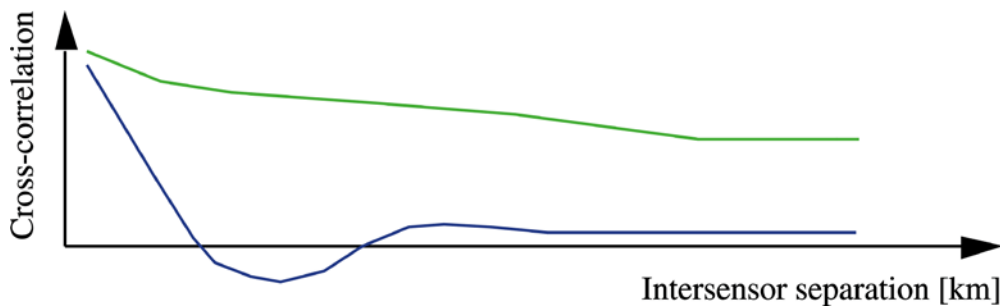
When plotting the cross-correlation values as a function of sensor separation only and thus disregarding possible directional dependencies, an implicit assumption is made of azimuthal symmetry in wavenumber space, over a longer time interval. This assumption is justified by the NORES experience, which shows that only a relatively small scatter is exhibited in the correlation data.

Computations of the kind described above should also be done for signals, which for the purpose of this section will be assumed to be P waves (although design strategies for the detection of S-type phases will be similar to those described here). A recording period of about 14 days during the site survey hopefully should be sufficient to record a reasonable number of representative P-wave arrivals. The time windows for these computations should be relatively short (approximately 5 seconds) to capture the coherent part of the signal arrival. Signal time series must be aligned in accordance with the signal slowness (phase velocity and direction of approach) before the cross-correlation is computed. Again, the time series must be filtered in a relatively narrow band around the peak frequency of the signal being considered. A plot like the one shown in Fig. 9.39 results from this, again assuming a six-sensor layout with 15 unique combinations of sensor pairs.



**Fig. 9.39** Schematic plot of P-wave cross-correlation values for a test layout of 6 sensors.

Computations as described above should be repeated for various time intervals for the noise, and for various P arrivals recorded during the site survey. Then, for each frequency interval of interest, all data (both noise and signal correlation data) should be combined in one diagram for the purpose of deriving curves (based on interpolation) that are representative of that frequency interval, and that provides correlation values for all intersensor separations. Such diagrams might then appear as shown in Fig. 9.40, in which the upper curve represents the signal correlation and the lower curve the noise correlation.



**Fig. 9.40** Schematic plot of signal (upper, green) and noise (lower, blue) correlation curves representing experimental data collected for a test array.

For the noise correlation curve in Fig. 9.40 to be representative for 2 Hz, e.g., the noise data should be filtered in a band where 2 Hz is close to the lower limit of the passband, due to the spectral fall-off of the noise. A passband of 1.8 – 2.8 Hz might be appropriate for the noise, but actual noise spectra for the site in question should be computed and studied before this passband is decided on. To generate a signal correlation curve representative for 2 Hz, signals should be used that have their spectral peaks close to this frequency, and some narrow filter passbands centered on 2 Hz should be applied to the data. These curves would then be used to predict gains for various array designs as detailed below.

It should be noted that the rather pronounced negative minima for the noise correlation curves (as schematically represented in Fig. 9.40) are consistently observed for the NORES array. It is the exploitation of this feature that provides for gains in excess of  $\sqrt{M}$  (see Section 9.4.5 and Eq. 9.7), commonly observed at the NORES array (or sub-geometries thereof). It should also be noted that this feature of negative noise correlation values is not a universal one; e.g., Harjes (1990) did not find consistently such pronounced negative minima for the GERES test array in Germany.

Braun and Schweitzer (2008) investigated the noise correlation behavior for a test-array installation of three-component sensors in Italy and similar intersensor-noise correlations were found for the rotated horizontal components as for the vertical sensors. The noise-correlation lengths (200 m for signals in the 2 – 4 Hz frequency passband) were about half as long as those observed for NORES. In addition, they also found specific BAZ directions with coherent noise with apparent velocities of Rg-type waves, which may be reducible by special array design.

## **9.12 Considerations when planning a new seismic array**

### **9.12.1 Basics**

Earlier in this chapter we described some older but still operational large seismic arrays. Today, with numerous digital seismic stations and the CTBTO monitoring arrays around the world, there is much less need for new arrays focusing on teleseismic monitoring. However, this is not true for small aperture arrays focusing on seismic sources at local and regional distances.

Based on very positive experiences during the last years with NORES-type, small aperture arrays, a recommended approach would be to use such an array geometry when planning a new array. However, it became also clear that an array does not necessarily need to consist of as many sites as NORES with its 25 sites. Quite good slowness (i.e., apparent velocity and backazimuth) estimates can already be achieved with SPITS-type arrays consisting of nine sites (corresponding to the A and B rings of NORES, see Fig. 9.41). The aperture of such a nine-element array should be between 750 and 1500 m depending on the local noise conditions (see Section 9.12.2). Then, even the identification and rough classification of teleseismic onsets is still possible.

With respect to instrumentation, it is no problem to keep the single sensors as broadband as possible. In particular when monitoring stronger events at local or regional distances the recording of higher frequencies in parallel with seismic low frequency energy can be of high importance, e.g., the new SPITS array is equipped with nine broadband, acceleration-proportional sensors, with a flat response between 100 s and 35 Hz.

A general problem of existing arrays is that they all record seismic waves predominantly with the use of vertical sensors only. Thereby, S-phase observation is much more difficult and limited to S phases with strong SV particle movements. This problem, discussed since some time, was partly solved for NORES-type arrays by installing three-component sensors at several sites (e.g., Schweitzer, 1994; Schweitzer and Kværna, 2002; Braun et al., 2004; Gibbons et al., 2011). Today, we can recommend equipping as many array sites as possible with three-component sensors as e.g., is the case for the refurbished Gräfenberg array, which has now three-component sensors at all 13 sites, and SPITS, which is equipped at six of nine sites with three-component sensors (Gibbons et al., 2011).

Finally, we would still like to recommend a central timing system for any array. During the last decades many cases were observed of malfunctioning single clocks at individual array sites, which produced unexpected (and sometimes undetected) wrong slowness results due to small time shifts. If the central clock of an array is showing a wrong time, the absolute onset times of seismic phases may be influenced, but the array itself is still functional.

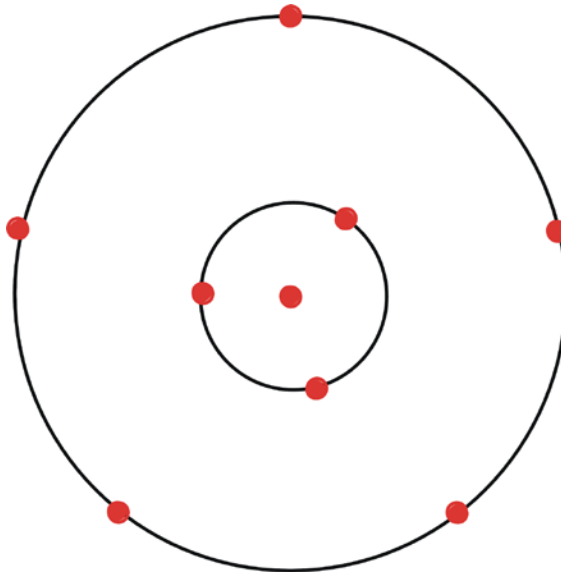
### 9.12.2 Example: A possible design strategy for a 9-element array

As an example of application of the design ideas outlined above, let us consider practical aspects of the design of a 9-element array. Several new arrays that have been built by the CTBTO for the International Monitoring System (IMS) for CTBT monitoring comprise 9 elements. A useful design for a new 9-element array would be one for which there are 3 and 5 elements equidistantly placed on each of two concentric rings, respectively, plus one element at the center of the geometry, as shown in Fig. 9.41.

The elements on the two rings should be placed so as to avoid radial alignment, which would produce stronger side lobes in the array transfer function as any regular pattern of the array geometry. If the five elements of the outer ring are placed at 0, 72, 144, 216 and 288 degrees from due North, the elements of the inner ring might be placed at 36, 156 and 276 degrees, as shown in Fig. 9.41. Within this class of design, the problem at hand is thus to find the radii of the two rings that for a given site would provide the best overall array gain. To constrain the design options even further, one might consider adopting the NORES design idea, limited to these two rings. The radii of the four NORES rings are given by the formula:

$$R = R_{\min} \cdot 2.15^n, (n = 0, 1, 2, 3) \quad (9.28)$$

For NORES,  $R_{\min} = 150$  m. For the design problem at hand, only  $R_{\min}$ , the radius of the inner ring, remains to be determined from the correlation data, whereas the radius of the outer ring would then be 2.15 times the radius of the inner ring.



**Fig. 9.41** The figure shows a possible design for a 9-element array.

The final step in the procedure outlined here is to compute expected gains for various array designs within this class of geometries. To this end, one must determine which signal frequencies are of the largest importance with regard to the detection capability of the array at the site under study. Assuming that three dominating P-wave signal frequencies,  $f_1$ ,  $f_2$  and  $f_3$ , (e.g., 1.8, 2.5 and 3.5 Hz) have been identified in the preparatory survey, these should be taken into account in the computations to derive the optimum array geometry. We would then

have available from the site survey empirically-based correlation curves in analytical or tabular form that would provide correlation values for **all** intersensor separations of interest. The gains as a function of frequency for various values of the parameter  $R_{min}$  could then be computed using the formula for the array gain and the correlation values derived from the correlation curves, for the relevant intersensor separations. The results of these computations could be tabulated as indicated in the (here artificial) Tab. 9.1. For more details and application of these analysis steps see Mykkeltveit (1985).

Note that for the lowest frequency considered ( $f_1$ ), it might pay in terms of array gain to exclude the elements of the inner ring from the gain computations, since noise correlation values for low frequencies may be high for many sensor pairs involving sensors of relative short intersensor distances (e.g., the innermost sensor rings).

**Tab. 9.1** Example for a table providing gains achievable by beamforming when applying different values of the parameter  $R_{min}$ .

| $R_{min}$ [m] | Gain ( $f_1$ ) [dB] | Gain ( $f_2$ ) [dB] | Gain ( $f_3$ ) [dB] |
|---------------|---------------------|---------------------|---------------------|
| 200           | 3                   | 6                   | 8                   |
| 300           | 4                   | 8                   | 9                   |
| 400           | 5                   | 9                   | 7                   |
| ...           | ...                 | ...                 | ...                 |
| 1000          | 9                   | 7                   | 4                   |
| ...           | ...                 | ...                 | ...                 |

The optimum geometry would correspond to the value of  $R_{min}$  that gives the best overall gain in Tab. 9.1. This judgment would be based on some appropriate weighting scheme for the frequencies considered.

The procedure outlined here could be generalized to a class of designs for which the radii of the two rings are varied independently. Gain values would then be tabulated as shown in Tab. 9.1, but there would now be a sequence of tables (each table would represent a fixed radius of one of the two rings). The search for the optimum geometry would then be performed across all these tables.

## 9.13 Routine processing of array data at NORSAR

### 9.13.1 Introduction

Now, we will explain the main features of the automatic routine processing of data recorded by regional arrays at NORSAR (Fyen 1989; 2001a and b).

As aforementioned in Section 9.2, the array data processing can be divided into three steps:

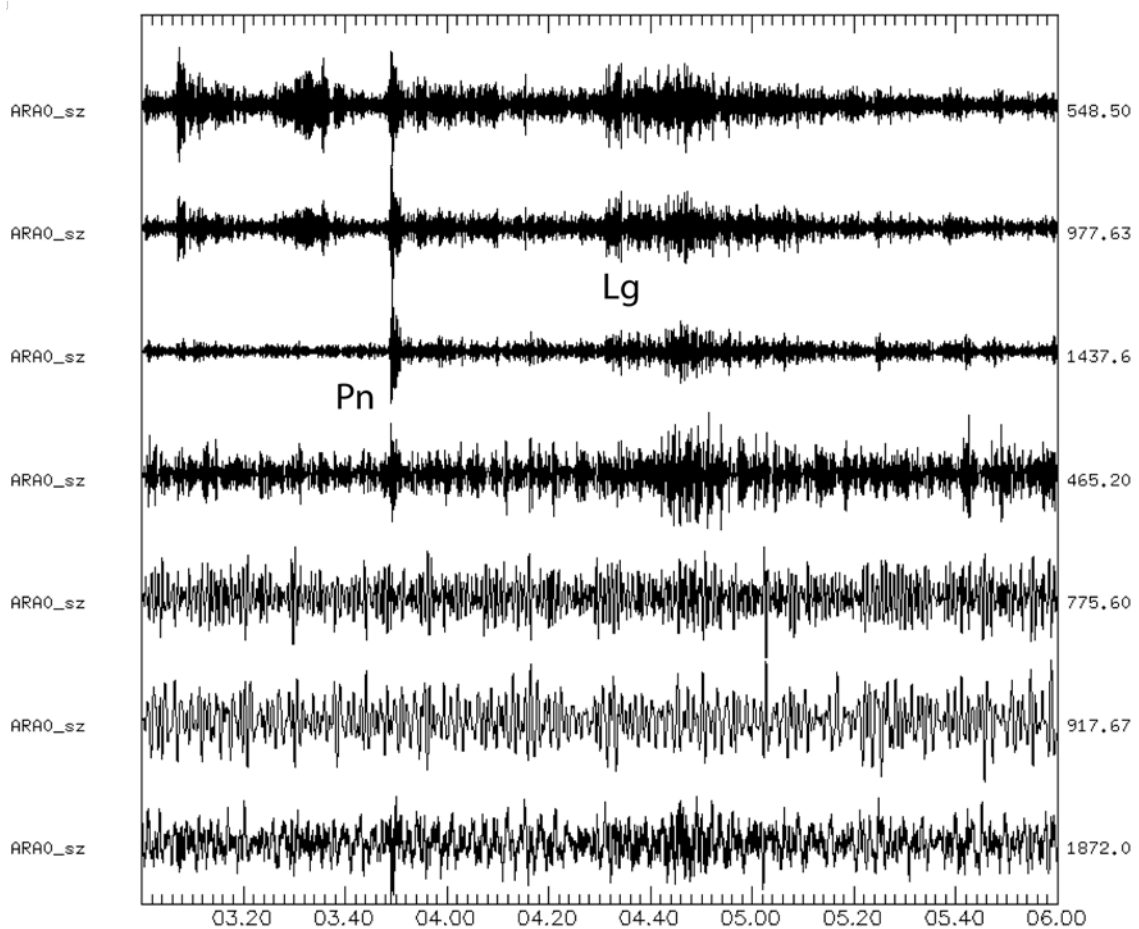
- Detection Processing (DP), i.e., performing STA/LTA triggering on a number of predefined beams;



- Signal Attribute Processing (SAP), i.e., performing signal feature extraction of detected signals, such as onset times, amplitudes, and frequencies/periods of identified phases; and
- Event Processing (EP), i.e., performing phase association, location processing and event plotting.

In Mykkeltveit et al. (1988) and in Kværna (1989), it is shown that different combinations of sensors, for example, within the NORES array, give different noise reduction for various frequency bands. The lesson is that it is not always optimal to use all seismometers of the array to form a beam; rather one should in general use different sub-configurations, tailored to the signal frequencies.

Fig. 9.16 showed different beams for the same P-wave signal. It is obvious from this figure that we need beams for various slowness vectors to detect the signal.



**Fig. 9.42** The bottom trace shows raw data from instrument A0 at the center of the ARCES array. The next traces from bottom to top are data from the same instrument filtered with 3rd order Butterworth bandpass filters using frequency bands 0.5 – 1.5, 1 – 3, 2 – 4, 4 – 8, 6 – 12, and 8 – 16 Hz, respectively.

We have earlier pointed out the importance of beamforming and filtering for signal enhancement. In the array detection process, several beams are formed, and several different filters are used (see Tab. 9.2). An STA/LTA detector is used on each beam, and as seen from

Fig. 9.17, we may get a trigger on several beams. The detector will compare the maximum STA/LTA (SNR) for every beam within a (narrow) time window. The influence of different filters on the detectability of seismic signals is demonstrated in Fig. 9.42. It shows ARCES data with one of the seismometer outputs filtered in different filter bands. An important point made in this figure is that the regional seismic phases Pn and Lg have their best SNR in different frequency bands. So to be sure to detect both phases, we should use several filter bands in the detector recipe.

**Tab. 9.2** An example for a detection beamset as it was in use at NORSAR for many years for the SPITS array. THR is the SNR threshold used to define a state of detection and “all” means that the whole SPITS array (SPA0, SPA1, SPA2, SPB1, SPB2, SPB3, SPB4, and SPB5) is used to form this beam (from Schweitzer, 1998).

| BEAM NAMES  | VELOCITY [km/s] | BACKAZIMUTH [deg]                          | FILTER        |       | THR | SITES (verticals only)        |
|-------------|-----------------|--------------------------------------------|---------------|-------|-----|-------------------------------|
|             |                 |                                            | bandpass [Hz] | order |     |                               |
| S001        | 99999.9         | 0.0                                        | 0.8 – 2.0     | 4     | 4.5 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| S002        | 99999.9         | 0.0                                        | 0.8 – 2.0     | 4     | 4.5 | all                           |
| S003        | 99999.9         | 0.0                                        | 1.0 – 3.0     | 3     | 4.5 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| S004        | 99999.9         | 0.0                                        | 1.0 – 3.0     | 3     | 4.5 | all                           |
| S005        | 99999.9         | 0.0                                        | 2.0 – 4.0     | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| S006        | 99999.9         | 0.0                                        | 2.0 – 4.0     | 3     | 4.0 | all                           |
| S007        | 99999.9         | 0.0                                        | 3.0 – 5.0     | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| S008        | 99999.9         | 0.0                                        | 3.0 – 5.0     | 3     | 4.0 | all                           |
| S009        | 99999.9         | 0.0                                        | 0.9 – 3.5     | 4     | 4.5 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| S010        | 99999.9         | 0.0                                        | 0.9 – 3.5     | 4     | 4.5 | all                           |
| S011        | 99999.9         | 0.0                                        | 1.0 – 4.0     | 3     | 4.5 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| S012        | 99999.9         | 0.0                                        | 1.0 – 4.0     | 3     | 4.5 | all                           |
| SA01 – SA04 | 10.0            | 0 90 180 270                               | 1.0 – 3.0     | 3     | 4.5 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SA05 – SA08 | 10.0            | 45 135 225 315                             | 1.0 – 3.0     | 3     | 4.5 | all                           |
| SA09 – SA12 | 10.0            | 0 90 180 270                               | 2.5 – 4.5     | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SA13 – SA16 | 10.0            | 45 135 225 315                             | 2.5 – 4.5     | 3     | 4.0 | all                           |
| SA17 – SA20 | 10.0            | 0 90 180 270                               | 4.0 – 8.0     | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SA21 – SA24 | 10.0            | 45 135 225 315                             | 4.0 – 8.0     | 3     | 4.0 | all                           |
| SA25 – SA28 | 10.0            | 0 90 180 270                               | 3.0 – 6.0     | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SA29 – SA32 | 10.0            | 45 135 225 315                             | 3.0 – 6.0     | 3     | 4.0 | all                           |
| SB01 – SB04 | 7.0             | 0 90 180 270                               | 1.0 – 4.0     | 3     | 4.5 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SB05 – SB08 | 7.0             | 45 135 225 315                             | 1.0 – 4.0     | 3     | 4.5 | all                           |
| SB09 – SB12 | 7.0             | 0 90 180 270                               | 3.0 – 6.0     | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SB13 – SB16 | 7.0             | 45 135 225 315                             | 3.0 – 6.0     | 3     | 4.0 | all                           |
| SB17 – SB20 | 7.0             | 0 90 180 270                               | 5.0 – 10.0    | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SB21 – SB24 | 7.0             | 45 135 225 315                             | 5.0 – 10.0    | 3     | 4.0 | all                           |
| SC01 – SC04 | 5.0             | 0 90 180 270                               | 1.0 – 4.0     | 3     | 4.5 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SC05 – SC08 | 5.0             | 45 135 225 315                             | 1.0 – 4.0     | 3     | 4.5 | all                           |
| SC09 – SC12 | 5.0             | 0 90 180 270                               | 3.5 – 5.5     | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SC13 – SC16 | 5.0             | 45 135 225 315                             | 3.5 – 5.5     | 3     | 4.0 | all                           |
| SC17 – SC20 | 5.0             | 0 90 180 270                               | 5.0 – 10.0    | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SC21 – SC24 | 5.0             | 45 135 225 315                             | 5.0 – 10.0    | 3     | 4.0 | all                           |
| SC25 – SC28 | 5.0             | 0 90 180 270                               | 8.0 – 16.0    | 3     | 4.0 | SPA0 SPB1 SPB2 SPB3 SPB4 SPB5 |
| SC29 – SC32 | 5.0             | 45 135 225 315                             | 8.0 – 16.0    | 3     | 4.0 | all                           |
| SD01 – SD08 | 4.0             | 0 45 90 135 180 225 270 315                | 0.9 – 3.5     | 4     | 4.5 | all                           |
| SD09 – SD16 | 4.0             | 0 45 90 135 180 225 270 315                | 3.0 – 6.0     | 3     | 4.0 | all                           |
| SD17 – SD24 | 4.0             | 0 45 90 135 180 225 270 315                | 4.0 – 8.0     | 3     | 4.0 | all                           |
| SE01 – SE08 | 3.3             | 0 45 90 135 180 225 270 315                | 1.5 – 3.5     | 3     | 4.5 | all                           |
| SE09 – SE16 | 3.3             | 0 45 90 135 180 225 270 315                | 3.0 – 6.0     | 3     | 4.0 | all                           |
| SE17 – SE24 | 3.3             | 0 45 90 135 180 225 270 315                | 5.0 – 10.0    | 3     | 4.0 | all                           |
| SF01 – SF08 | 2.5             | 0 45 90 135 180 225 270 315                | 1.0 – 4.0     | 3     | 4.5 | all                           |
| SF09 – SF16 | 2.5             | 0 45 90 135 180 225 270 315                | 2.0 – 4.0     | 3     | 4.0 | all                           |
| SF17 – SF24 | 2.5             | 0 45 90 135 180 225 270 315                | 3.0 – 5.0     | 3     | 4.0 | all                           |
| SN01        | 8.4             | 97.6                                       | 2.0 – 4.0     | 3     | 3.7 | all                           |
| SN02        | 8.4             | 97.6                                       | 3.0 – 5.0     | 3     | 3.7 | all                           |
| SN03        | 8.4             | 97.6                                       | 4.0 – 8.0     | 3     | 3.7 | all                           |
| SN04        | 8.4             | 97.6                                       | 6.0 – 12.0    | 3     | 3.7 | all                           |
| SN05        | 8.4             | 97.6                                       | 8.0 – 16.0    | 3     | 3.7 | all                           |
| SN06        | 4.7             | 97.6                                       | 2.0 – 4.0     | 3     | 3.7 | all                           |
| SN07        | 4.7             | 97.6                                       | 3.0 – 5.0     | 3     | 3.7 | all                           |
| SN08        | 4.7             | 97.6                                       | 4.0 – 8.0     | 3     | 3.7 | all                           |
| SN09        | 4.7             | 97.6                                       | 6.0 – 12.0    | 3     | 3.7 | all                           |
| SN10        | 4.7             | 97.6                                       | 8.0 – 16.0    | 3     | 3.7 | all                           |
| SG01 – SG12 | 2.0             | 0 30 60 90 120 150 180 210 240 270 300 330 | 1.5 – 3.5     | 3     | 4.5 | all                           |
| SG13 – SG24 | 2.0             | 0 30 60 90 120 150 180 210 240 270 300 330 | 2.5 – 4.5     | 3     | 4.0 | all                           |
| SG25 – SG36 | 2.0             | 0 30 60 90 120 150 180 210 240 270 300 330 | 3.5 – 5.5     | 3     | 4.0 | all                           |
| SM01 – SM12 | 1.7             | 0 30 60 90 120 150 180 210 240 270 300 330 | 1.0 – 3.0     | 3     | 4.5 | all                           |
| SM13 – SM24 | 1.7             | 0 30 60 90 120 150 180 210 240 270 300 330 | 2.0 – 4.0     | 3     | 4.0 | all                           |
| SM25 – SM36 | 1.7             | 0 30 60 90 120 150 180 210 240 270 300 330 | 3.0 – 6.0     | 3     | 4.0 | all                           |

Thus we have three parameter sets that make up the input for the STA/LTA detector: the specific array configuration, the slowness vector, and the filter band to use for this beam. Note that one could also use just a single seismometer instead of a beam. Based on experiments, a list of these parameters has been compiled at NORSAR that constitutes a “detector recipe” with, e.g., numerous beams using different slownesses, different configurations, and different filter bands. For a large signal, the detector program will trigger on many beams, and the program will use a detection reduction process to report only one detection for each signal.

As an example, a detector recipe listing the entire beam set composed of 254 beams for the online processing of data from the SPITS array, as it was in use at NORSAR for many years, is included in Tab. 9.2. The complete process is illustrated in the following Sections 9.13.2 – 9.13.4 by using a data example from the ARCES array.

### 9.13.2 Detection Processing – DP

The DP process continuously reads data off a disk loop or any other continuous database and uses beamforming, filtering, and the STA/LTA detector to obtain detections (triggers). Thus the DP program produces, e.g., a single file for each array and each day of the year (DOY).

The following list gives some example lines from such a file for DOY 199 in 1996 for the ARCES (ARC) array. The file contains the name of the detecting beam (e.g., F074), the time of detection (199:16.03.49.3), the end of the detection state (199:16.03.53.1), the maximum STA (242.4), the LTA at the time of detection (10.27), the SNR ( $\text{STA}/\text{LTA} = \text{SNR} = 23.601$ ), and the number of beams detecting (37). The detecting beam reported (here F074) is the one beam, normally out of many, that detected this signal with the highest SNR. The key parameters reported are the beam code, the trigger time, and the  $\text{SNR} = \text{STA}/\text{LTA}$ . The beam code (leftmost column in the list below) points to a file (see Fig. 9.43) containing information on beam configuration, slowness and filter used. The format of a detection output is not important. Important is to create a list of detections that can be used for further analysis.

| BEAM | DETECTION TIME           | STA    | LTA   | SNR    | # of BEAMS |
|------|--------------------------|--------|-------|--------|------------|
| FH04 | 199:16.02.18.6 - 02.19.9 | 106.52 | 32.15 | 3.313  | 1          |
| FH04 | 199:16.02.39.1 - 02.39.6 | 116.09 | 43.91 | 2.644  | 1          |
| FH04 | 199:16.02.44.4 - 02.48.6 | 296.52 | 56.63 | 5.236  | 5          |
| FI01 | 199:16.02.55.5 - 02.56.3 | 175.40 | 33.59 | 5.221  | 3          |
| F074 | 199:16.03.49.3 - 03.53.1 | 242.40 | 10.27 | 23.601 | 37         |
| FH03 | 199:16.04.32.9 - 04.50.9 | 291.46 | 73.01 | 3.992  | 5          |
| FH02 | 199:16.04.46.4 - 04.35.9 | 297.87 | 75.23 | 3.960  | 24         |
| FH04 | 199:16.09.57.6 - 09.58.9 | 80.86  | 24.07 | 3.359  | 2          |

```

THR = 4.000
BF1 = 3.500
BF2 = 5.500
BMVEL = 11.100
BMAZI = 150.000
REFLAT = 69.535
REFLON = 25.506
REFELE = 403.000
REFSIT = ARA0_sz
SELECTED CHANNELS : ARA0_sz ARB1_sz ARB2_sz ARB3_sz ARB4_sz ARB5_sz
ARC1_sz ARC2_sz ARC3_sz ARC4_sz ARC5_sz ARC6_sz ARC7_sz

```

**Fig. 9.43** Example of the contents of a file with the parameters that characterize beam F074. THR is the SNR detection threshold, BF1 and BF2 are the lower and the upper limits of the bandpass filter applied, BMVEL and BMAZI are the apparent velocity and the backazimuth for this beam, REFLAT, REFLON, REFELE, and REFSIT define the reference site of the beam, and SELECTED CHANNELS lists the site configuration.

### 9.13.3 Signal Attribute Processing – SAP

This process sequentially reads detections from the detection file and performs an f-k analysis for each detection to estimate apparent velocity and backazimuth. The estimated velocity and backazimuth is referred to as “observed slowness”. Waveform segments for the analysis are again read from a disk loop or any other database.

A special version of NORSAR’s EP program is used and produces a new file for the ARCES array on day-of-the-year (DOY) (here 199 in 1996). The key parameters reported in the EP-result files are the signal onset time, the beam code, the SNR, the estimated slowness, the signal amplitude and frequency, and the phase identification based on the slowness estimate.

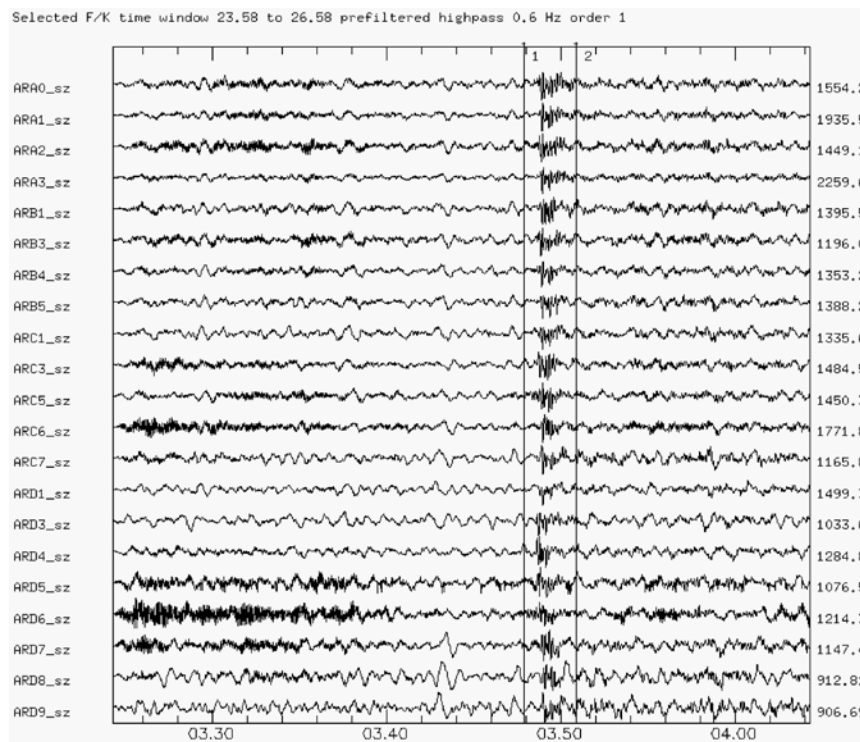
| #  | ONSET TIME       | DT   | BEAM | SNR  | VEL | PHASE | BAZ   | POWER | Q | FREQ | AMP    | STA   | IP    | IS | REC  | H/V  | INC 1 | INC 3 |
|----|------------------|------|------|------|-----|-------|-------|-------|---|------|--------|-------|-------|----|------|------|-------|-------|
| 5  | 199:16.02.18.314 | 0.29 | FH04 | 3.3  | 2.4 | nois  | 256.5 | 0.30  | 4 | 9.22 | 343.4  | 106.5 | 1     | -3 | 0.63 | 0.86 | 19.32 | 81.70 |
| 10 | 199:16.02.36.964 | 2.14 | FH04 | 2.6  | 2.4 | nois  | 246.6 | 0.37  | 3 | 9.79 | 466.6  | 116.1 | 1     | -3 | 0.75 | 0.29 | 6.02  | 88.69 |
| 15 | 199:16.02.43.014 | 1.39 | FH04 | 5.2  | 2.4 | nois  | 243.4 | 0.56  | 3 | 9.79 | 906.9  | 296.5 | 1     | -3 | 0.56 | 0.69 | 10.38 | 80.87 |
| 20 | 199:16.02.54.915 | 0.58 | F101 | 5.2  | 2.4 | nois  | 236.7 | 0.59  | 3 | 9.84 | 1021.5 | 175.4 | 1     | -3 | 0.63 | 0.42 | 10.56 | 86.32 |
| 25 | 199:16.03.48.409 | 0.89 | F074 | 23.6 | 7.5 | Pgn   | 122.7 | 0.71  | 2 | 4.85 | 518.5  | 242.4 | 0     | -3 | 0.72 | 0.61 | 48.76 | 68.81 |
| 30 | 199:16.04.32.160 | 0.74 | FH03 | 4.0  | 5.4 | SN    | 117.4 | 0.29  | 3 | 6.30 | 581.6  | 291.5 | D4_sz |    | 0.37 | 1.78 | 10.97 | 12.82 |
| 35 | 199:16.04.45.785 | 0.62 | FH02 | 4.0  | 4.1 | LG    | 127.2 | 0.28  | 3 | 4.88 | 552.1  | 297.9 | D4_sz |    | 0.44 | 1.10 | 78.22 | 72.30 |
| 40 | 199:16.09.55.770 | 1.83 | FH04 | 3.4  | 2.4 | nois  | 197.7 | 0.37  | 4 | 9.48 | 228.7  | 80.9  | -1    | -2 | 0.63 | 1.24 | 87.74 | 68.06 |
| 45 | 199:16.10.39.171 | 1.33 | F106 | 4.8  | 2.5 | nois  | 221.7 | 0.44  | 3 | 8.38 | 226.2  | 59.7  | 1     | -3 | 0.74 | 0.40 | 18.91 | 76.60 |

**Fig. 9.44** Example of the contents of a file with the Signal Attribute Processing results for the detections listed in Fig. 9.43.

A few lines from such a file are listed above (Fig. 9.44). The entries are the arrival id number (e.g., 25), the estimated onset time (199:16.03.48.409), the difference time (DT) between trigger and onset time (0.89 s), the beam name (F074), the SNR (23.6), the apparent velocity (VEL in km/s) from f-k analysis (7.4), the preliminary phase name by automatically considering apparent velocity and three-component polarization analysis (Pgn, which is used

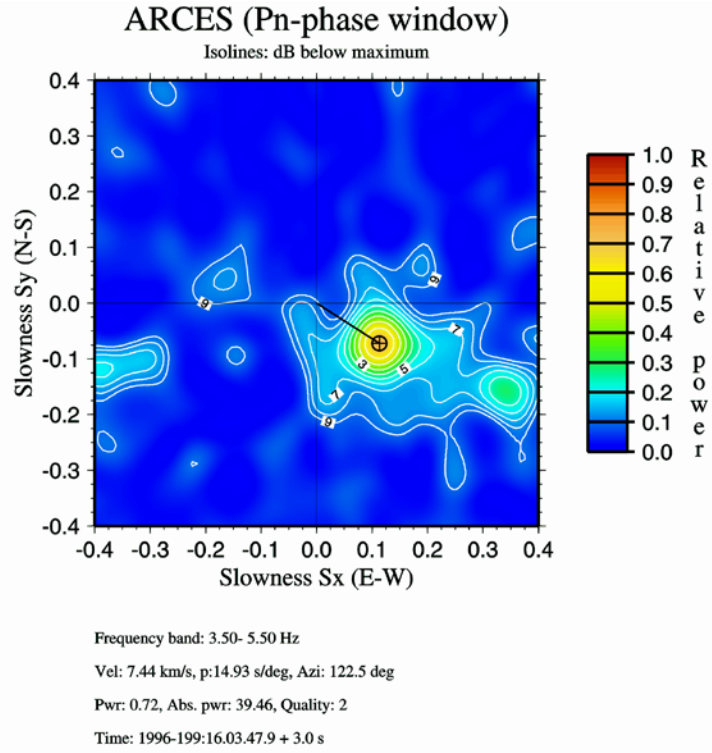
here as a code for ‘either Pg or Pn’), the estimated backazimuth (BAZ) from f-k analysis ( $122.5^\circ$ ), the relative power from f-k analysis (0.72, a number between 0.0 (no signal coherency) and 1.0 (perfect signal coherency)), the f-k analysis quality indicator (Q) (2, 1=best, 4=poor), the estimated dominant frequency (FREQ) in Hz (4.85), the maximum amplitude (AMP) in counts (476.9), the maximum STA of the detection (242.4), the polarization analysis classifications IP, IS (0 and  $-3$ , respectively), the polarization analysis rectilinearity (REC) (0.69), the horizontal/vertical ratio (H/V) (0.49), the inclination angle (INC) ( $41.26^\circ$ ), and the polarization angle (POL) ( $73.94^\circ$ ).

Fig. 9.45 shows raw data for the detection reported at time 199:16.03.49.3 (see the detection result listing). The signal attribute process will use this detection time to measure a first onset time of the signal and then to select a 3 second wide time window starting 0.5 second before the preliminary onset time (here 199:16.03.48.4, see also the time window indicated by vertical lines). The data from all vertical seismometers within this time window will then be used for f-k analysis to obtain the true apparent velocity of the signal.

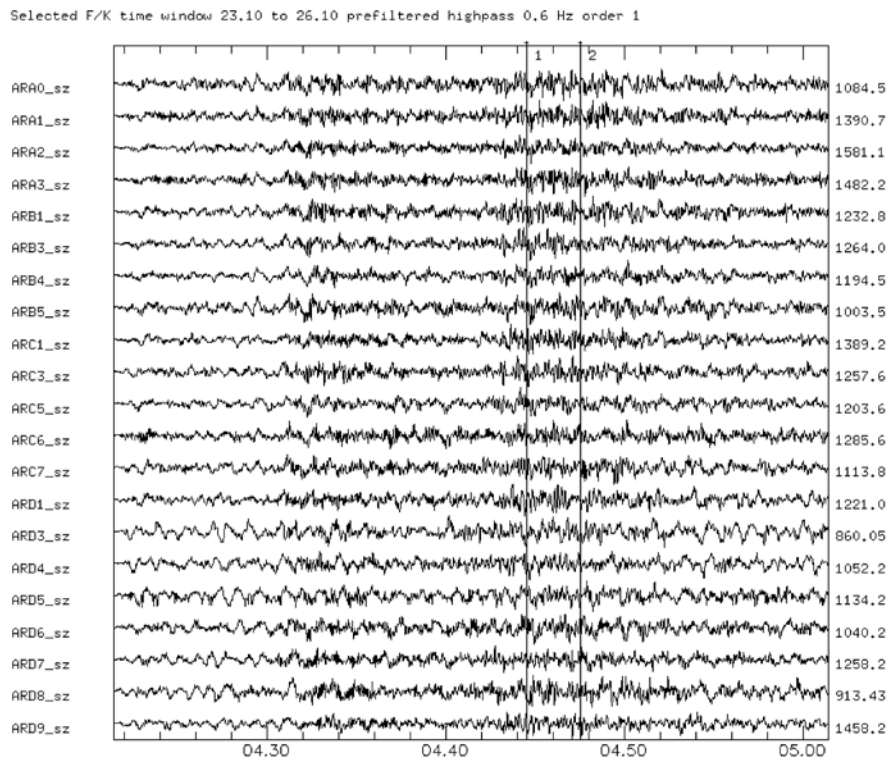


**Fig. 9.45** Raw data from all the vertical seismometers of the ARCES array. The time interval contains the Pn phase of a regional event. The vertical bars define a 3 second time window that is used for the f-k analysis.

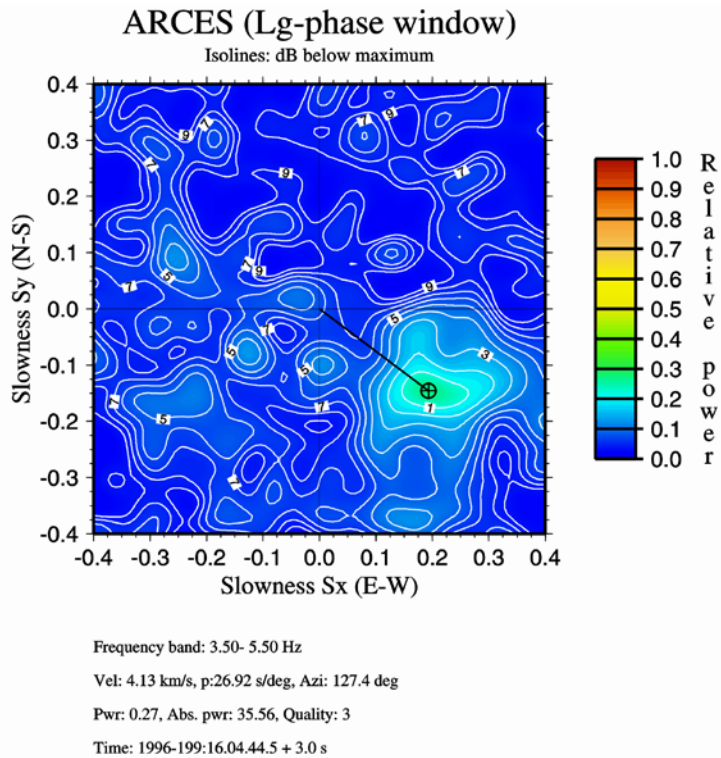
The result for the f-k analysis is shown in Fig. 9.46. This process is repeated for all detections and Fig. 9.47 shows the data interval selected for the Lg detection. The corresponding f-k analysis results are shown in Fig. 9.48. In automatic mode, of course, the EP program will not display any graphics. The figures are only produced for illustration purposes. However, the capability of displaying results graphically at any step of a process is essential to be able to develop optimum recipes and parameters. The EP program may output results into flat files or a database.



**Fig. 9.46** Result of the wide-band f-k analysis for the data in Fig. 9.45, pertaining to the Pn-phase interval.



**Fig. 9.47** Raw data from all the vertical seismometers of the ARCES array. The time interval contains the Lg phase of a regional event. The vertical bars define a 3 second time window that is used for the f-k analysis.



**Fig. 9.48** Result of the wide-band f-k analysis for the data in Fig. 9.47, pertaining to the Lg-phase interval.

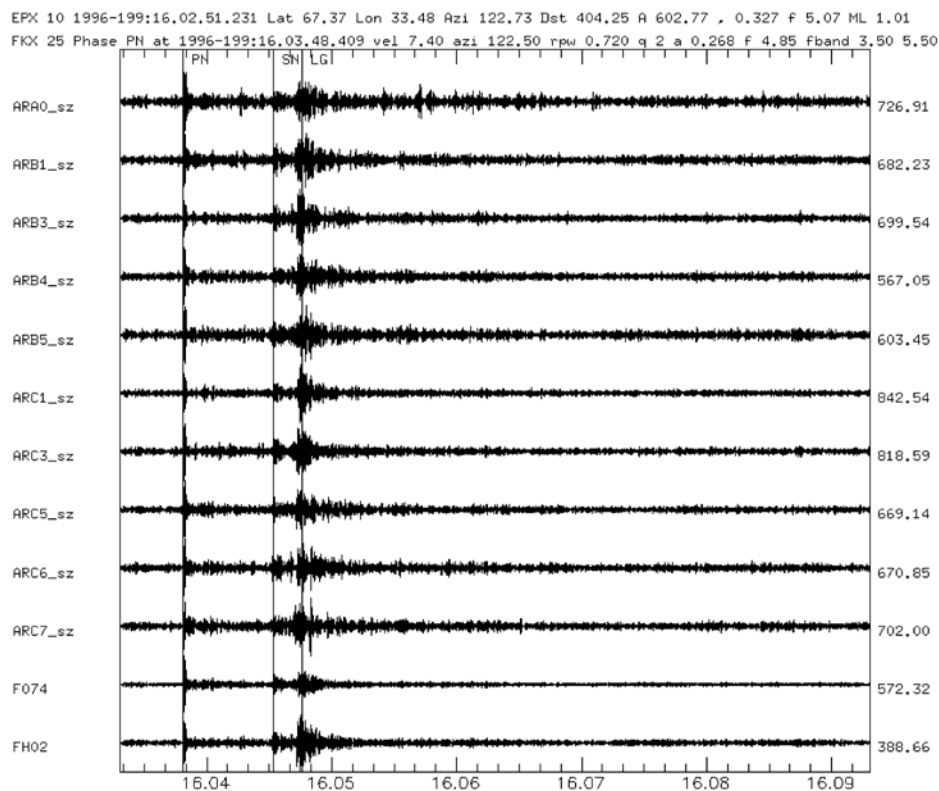
### 9.13.4 Event Processing – EP

This process sequentially reads all detections from the parameter extraction processing file. Whenever a detection with an apparent velocity greater than, e.g., 6.0 km/s is found, it is treated as P. Additional detections are searched for, and if additional detections are found with backazimuth estimates not more than, e.g., the predefined azimuth range of 30° from the first detection and a detection time not more than 4 minutes from the first detection, they are used as associated detections. If detections with an apparent velocity less than 6.0 km/s are found, then they are treated as S (Sn, Lg). If phases within 4 minutes and backazimuth deviation of less than 30° with a first P and an S later are found, then they are treated as observations from a regional event. A location routine, which uses the backazimuth information, is used to locate the regional event. More details on these topics are given in Mykkeltveit and Bungum (1984). The results are again written in a separate file for each DOY (here 199 in 1996). A few lines from such files are exemplarily listed below:

|    |                |     |      |                                 |        |       |      |       |       |           |
|----|----------------|-----|------|---------------------------------|--------|-------|------|-------|-------|-----------|
| 5  | 199:16.02.18.3 | FRS | nois | 0.148                           | 9.2    | 3.3   | FH04 | 2.4   | 256.5 | Noplot3ci |
| 10 | 199:16.02.37.0 | FRS | nois | 0.205                           | 9.8    | 2.6   | FH04 | 2.4   | 246.6 | Noplot3ci |
| 15 | 199:16.02.43.0 | FRS | nois | 0.398                           | 9.8    | 5.2   | FH04 | 2.4   | 243.4 | Noplot3ci |
| 20 | 199:16.02.54.9 | FRS | nois | 0.449                           | 9.8    | 5.2   | F101 | 2.4   | 236.7 | Noplot3ci |
| 10 |                | HYP | 724  | BALTIC STATES-BELARUS-NW RUSSIA |        |       |      |       |       |           |
| 10 | 199:16.02.51.2 | EPX |      | 67.382                          | 33.506 | 1.03  |      | 404.3 | 122.5 | 0F        |
| 25 | 199:16.03.48.4 | FRS | PN   | 0.291                           | 4.8    | 23.6  | F074 | 7.5   | 122.7 | LOCATE    |
| 30 | 199:16.04.32.2 | FRS | SN   | 0.273                           | 6.3    | 4.0   | FH03 | 5.4   | 117.4 | LOCATE    |
| 10 | 199:16.04.44.6 | MAG | ML   | 0.337                           | 4.9    | 602.8 | aVG  | 3.6   | 122.5 | LOCATE    |
| 35 | 199:16.04.45.8 | FRS | LG   | 0.309                           | 4.9    | 4.0   | FH02 | 4.1   | 127.2 | LOCATE    |
| 40 | 199:16.09.55.8 | FRS | nois | 0.099                           | 9.5    | 3.4   | FH04 | 2.4   | 197.7 | Noplot3ci |
| 45 | 199:16.10.39.2 | FRS | nois | 0.097                           | 8.4    | 4.8   | F106 | 2.5   | 221.7 | Noplot3ci |

Whenever an event is declared, a location is performed and reported with two lines (**HYP** and **EPX**) that contain event number (10), origin time (199:16.02.51.2), latitude (67.369°), longitude (33.479°), ML (1.01), distance in [km] (404.3), backazimuth (122.7°), fixed depth at 0 km (0F). The associated phases are listed thereafter, and for the Pn phase we have the id number (25), the arrival time (199:16.03.48.4), the station name (FRS, the old, NORSAR-internal code for ARC5), the phase name (PN), the maximum amplitude in [nm] (0.268), the corresponding dominant frequency in [Hz] (4.8), the SNR (23.6), the beam name (F074), the apparent velocity (7.4), the backazimuth (122.5°), and an explanatory code from the location process. **LOCATE** means that this phase was used for locating the event and **ASSOC** means that this onset was associated but not used for the location processing. **Tele** means that this phase is interpreted as a teleseismic onset and **Noplots3ci** means that this phase was not used for any event definition. Such onsets are then also often declared as **nois** (i.e., noise) observations. The “beam-name” **aVG** in a **MAG** line means that the corresponding arrival time is used for measuring the amplitude for **ML** as average amplitude of all traces, and the apparent velocity is the group velocity in that case. The key parameters reported are the origin time, the hypocenter, and the magnitude for each located event, and onset time, amplitude and frequency, SNR, beam code, and apparent velocity for all associated detections.

The example above identifies one group of phases with backazimuth around 123° that are arriving within 4 minutes. The first phase within the group has a regional P-wave apparent velocity, and it is followed by a phase with a regional S apparent velocity. Those are the predefined criteria for defining an event. For each declared event, an event plot may be created (see Fig. 9.49). Results of the automated regional array processing at NORSAR can be found at <http://www.norsardata.no/NDC/bulletins/dpep>.



**Fig. 9.49** Summarizing plot for the above discussed regional event.



### 9.13.5 Event location with a network of arrays

It became very soon obvious that the location precision of single small aperture arrays is quite limited mostly due to scatter in the backazimuth estimates and uncertainties in automatically measuring the travel-time differences between P- and S-type onsets. After building up a network of small aperture arrays in Northern and Central Europe during the late 1980s and early 1990s, a joint interpretation of observations from several small aperture arrays could be tested. One successful approach became the Generalized Beam Forming (GBF) location algorithm. This algorithm, developed at NORSAR, can automatically utilize the results of several seismic arrays in a common bulletin (Ringdal and Kværna, 1989; Kværna et al., 1999). Today, data from the highly sensitive regional arrays ARCES, FINES, HFS, SPITS, and NORES, and the teleseismic NORSAR array (NOA) are automatically processed in on-line mode applying this regional and local event-location process (<http://www.norsardata.no/NDC/bulletins/gbf>). For events observed at several arrays (i.e., events with local magnitudes above 2) these fully automatically achieved GBF locations are usually within a distance of 10 to 50 km from the analyst reviewed event locations (<http://www.norsardata.no/NDC/bulletin/regional/>). However, the location capabilities are very much depending on the relative location of the events to the geometry of the array network and the number of observed phases.

### 9.13.6 Teleseismic event location

Benndorf (1906; 1907) published that in the case of a spherically symmetric Earth model apparent velocities (or the seismic ray parameters) are constant along their whole ray path through the Earth (Benndorf's Law). If the velocities inside the Earth are known, it can easily be shown that the ray parameter of seismic onsets changes with the epicentral distance and that an observed ray parameter (or apparent velocity) can directly be inverted for the epicentral distance.

For modern spherically symmetric Earth models and seismic arrays of at least 10 km aperture, this principle works fine for first arriving P-type onsets from seismic events at teleseismic distances (i.e., from about 25° to about 100° epicentral distance). Events at shorter distances are hard to locate because the derivative of the apparent velocity with respect to distance is very small and triplications of the travel-time curve do not allow for a unique correspondence between apparent velocity and distance. At distances beyond ~90° these derivatives become for P waves again very small and in the Earth's shadow zone for P ( $D > 100^\circ$ ) the interpretation of the different core-phase onsets is also quite difficult and limits the location capabilities of a seismic array. Knowing the epicentral distance, the observed BAZ can then be used to define the epicentral coordinates.

The described event location technique has been in use at least since the 1960s and a quick look in the bulletins of the International Seismological Centre shows the huge amount of reported teleseismic event locations made with, e.g., the Large Aperture Seismic Array (LASA) in Montana, USA, the Yellowknife Array (YKA) in Northern Canada, the Gräfenberg Array (GRF) in Bavaria, Germany, or the large Norwegian Seismic Array (NOA) in southern Norway.

For the large NORSAR array (NOA), we may choose between frequency-domain f-k analysis or the beampacking process. The benefit of using beampacking rather than frequency domain f-k analysis is that for every point in slowness space, we can use time delay corrections (see Section 9.5.4) and obtain a calibrated slowness. On the basis of phase identification and measured slowness, we then get an epicentral distance by screening a slowness table and thereby a location using distance and backazimuth. So, for each detection a corresponding location is provided by NORSAR's near real-time array data processing. All results of the automatic array processing of the NOA array can be found at <http://www.norsardata.no/NDC/bulletins/dpep>.

## **9.14 Seismic arrays in Earthquake Early Warning Systems (EEWS)**

### **9.14.1 Introduction**

The advantage of applying seismic array techniques in EEWS is connected with the capability of an array not only to observe a seismic signal but also to measure its propagation direction and apparent velocity. However, the capability of an array to measure the backazimuth (BAZ) and apparent velocity with sufficient accuracy and to suppress other than the target signals is very much depending on the array geometry and the number of its sensors. Therefore, not each array is equally suitable for an Earthquake Early Warning System (EEWS). Here, we will focus on the usage of seismic arrays as EEWSs in general and in particular on real-time algorithms and discuss the advantages and disadvantages of applying array-analysis techniques as input for any EEWS.

### **9.14.2 Examples for array contributions to fast event locations**

#### **9.14.2.1 Single array results**

As mentioned earlier, it is possible to locate seismic events with data observed by one or more seismic arrays. However, in the case of any fast event location algorithm array analysis can only contribute if the whole data processing is automated. On the other hand, the recorded data volume of seismic arrays is very large, thus also requiring automated data processing techniques. Therefore, array data processing algorithms were as much as possible automated since the 1960s. For example, NORSAR's program package DP/EP as described in Section 9.13 was mostly developed in the 1970s – 1990s and later adapted to many other array installations (Fyen, 1989; 2001a and b).

After international exchange of emails was becoming more reliable and common in the early 1990s, it became possible to report event locations or strong P-phase observations based on fully automatic data processing algorithms. Thereby, results from the NORSAR, YKA, or the GERES array were automatically sent to, e.g., the USGS for its Quick Epicenter Determinations (QEDs), the European-Mediterranean Seismological Center (EMSC), the Swiss Seismological Service (SED), and the wider interested seismological community.

#### **9.14.2.2 The Fast Earthquake Information Service (FEIS) algorithm**

At the University of Bochum a special alert system was developed in the mid 1990s, which combined the mentioned single array observations with recordings of the newly at that time installed German Regional Seismological Network (GRSN). The so-called Fast Earthquake Information Service (FEIS) algorithm (Schulte-Theis et al., 1995; Harjes et al., 1996) was triggered by strong local or regional events observed by the GERES array. For this, the data of the regional GERES array were automatically analyzed in real time by applying the DP/EP array software developed at NORSAR (Fyen, 1989; 2001a and b, see also Section 9.13). After each automatic GERES location with a local magnitude above 3, the FEIS algorithm was triggered, consisting of the following steps:

- recalculation of an initial location from the GERES data alone;
- calculation of theoretical onset times for regional phases (Pn, Pg, Sn, Sg) at all GRSN stations;
- polling of all automatic GRSN-detection lists via telecommunication lines for a larger time interval around the assumed arrival times;
- searching a small time interval around the theoretical onset times for Pg or Pn detections, depending on the epicentral distance;
- in the case that a P-type phase could be associated, the detection lists were searched for possible S-type detections in a distance-depending predefined time window;
- relocation of the event with the GERES-observation parameters (phase names, onset times, BAZs) and the applicable GRSN detections;
- in the case of a stable location result, the determined location was distributed automatically to interested addresses in Europe.

A comparison with PDE (USGS) locations indicated that the automatic FEIS-relocation procedure significantly improved the original automatically achieved GERESS location accuracy, in particular for events, which occurred within the GRSN.

#### **9.14.2.3 NORSAR's Event Warning System (NEWS)**

Since 2000, a new quick event-location system was developed at NORSAR to provide fast and reliable solutions in the case of strong events: NORSAR's Event Warning System (NEWS) (Schweitzer, 2003). The whole NEWS algorithm is based on high Signal-to-Noise Ratio (SNR) detections; whenever one of the contributing arrays observes a P-type onset with an SNR larger than a predefined threshold, the NEWS process is initialized.

Once triggered, the NEWS process searches the automatic result lists of all other available arrays for corresponding onsets. Corresponding in this context means that the other onsets have to come from a backazimuth, and with an apparent velocity, which is consistent with the triggering onset. Formulating robust rules for which onsets can eventually be associated with the same event, was a quite cumbersome procedure. However, as implemented today, these rules are built on travel-time differences between the onset times at the different stations, measured backazimuth and apparent velocity of the signals, and SNR of the onsets. In the case of a presumably local or regional event, NEWS also searches for S-type onsets in the onset lists of the arrays.

After all available lists are searched the NEWS process locates the seismic event. To make this automatic event location as robust as possible, onset times and apparent velocity values are only used from first P and S arrivals. However, to use as much as possible information from the seismic arrays, all onsets in compliance with the selection rules and the measured backazimuth values are used to locate the event. Depending on the mean apparent velocity of all detected P onsets, the program defines the event as probably regional, or as near, far or very far teleseismic. Then, together with the mean backazimuth estimation, an initial source region is chosen. Depending on this initial solution, either a regional or a global velocity model is used to locate the event. The observed P amplitudes can be used to calculate a preliminary event magnitude.

For the determination of the source parameters NORSAR's location program HYPOSAT (see PD.11.1) is used. With the limited amount of data available for locating the event, the event's depth cannot be resolved and has therefore to be fixed to a predefined value. However, until now, such preliminary locations have been sufficient for preliminary information to the public in the case of local or regional events. Although the used network of seismic arrays has an aperture of about 18 degrees in the north-south direction, teleseismic events are usually observed over only a very small azimuth range. Therefore, the small number of available observations produces solutions with limited accuracy and large error bars, and some events are even not locatable. This is in particular true for events in the South Pacific, for which only PKP-type onsets can be observed.

On average, NEWS solutions are available between a few and up to about 10 minutes after the first P onsets have been recorded at one of the seismic arrays. A listing of the most recent NEWS solutions is available on the web (<http://www.norsardata.no/NDC/ael/eventmap.html>) and NEWS solutions are disseminated to international data centers, which also work on quick epicenter determinations. The delay of several minutes between the source time and the dissemination of source parameters of regional events by today's NEWS implementation is due to several factors:

- usually, the distance between a seismic event and the closest array recording it is in the order of several hundreds of kilometers;
- it takes several additional minutes until other arrays of the sparse network of arrays in Northern Europe can record the event;
- to achieve a more stable solution for the event location the NEWS algorithm is implemented in such a way that it also waits for possible S-type onsets;
- the location algorithm HYPOSAT (see PD 11.1) used for locating the event may be better optimized for shorter computation time.

### **9.14.3 Usage of seismic arrays to monitor an EEWS relevant site**

With its unique capability to measure not only onset times and amplitudes, but also BAZs and apparent velocities of seismic onsets, an array gives us several possibilities to locate an event. The only question is, which algorithm and data processing scheme should be used to provide quick locations for an EEWS. Working with the above mentioned methods and software packages, one can conclude that with today's computer capacities the most critical parameter for using seismic arrays in an EEWS is the epicentral distance to the array installation(s). All discussed algorithms are on today's computers so fast that the actual calculation times for the different algorithms do not really contribute to EEWS delays. More important are the actual

transmission times of seismic signals since all data connection algorithms work with data frames containing a specific amount of data. The delay time between the actual recording of a signal and its arrival at a data center can vary between seconds and minutes and has to be added to the EEWS times achievable by the discussed location algorithms.

#### **9.14.3.1 The single array case**

In the case of single array locations at local or near regional distances, the travel-time difference between source and arrival time of the first P phase is in the order of tens of seconds for local or near regional events. Additional tens of seconds will be needed to record the first S onset, necessary for calculating the epicentral distance. Therefore, such an array used as an EEWS tool will most likely need more time to locate the event than a traditional seismic network installed in the area of interest. The situation changes in many cases where several seismic active areas or a longer tectonic fault contribute to seismic hazard. Dense, local networks cannot be installed at all places and in particular if more remote or off-shore located zones contribute to a hazard scenario, single array installations can contribute, within a few minutes, with quite reliable event locations for all events within some hundred kilometers epicentral distance. Moreover, as shown by Gibbons et al. (2005), a single array can be tuned for a specific target area and the resulting location precision can become as high as that of a local network, assuming that sufficient calibration information is available. This is of particular interest in case of monitoring the aftershock sequence of a very large earthquake.

In the case of source regions with a higher risk to generate tsunamis, it is of large interest to know quickly not only the hypocenter of the earthquake but also at least a rough estimate of the magnitude and details of the rupture lengths. In such cases, algorithms as proposed and tested by Roessler et al. (2010) could be implemented (see also Section 9.8.6) in a tsunami warning system.

#### **9.14.3.2 A single array and a sparse regional network**

In the case that data from an array and additionally a regional or local network are available, a FEIS-type algorithm can be used. Knowing the BAZ and apparent velocity of the first P-type onset directly gives information about the direction in which the event occurred and if it was at a local or a regional distance. For regional events, the first P onset should have an apparent velocity typical for Pn phases and for local events typical for Pg onsets, respectively. With this information, the array result for the first P onset directly indicates, which single station records should be added to achieve a fast and more reliable event location.

An EEWS based on a single array and a sparse network can provide a first, quick and reasonably good event location at local or near regional distance ( $D < 400$  km) within the first minutes after the event origin time (OT) as long as one of the network stations is located as close as the array or closer to the event.

#### **9.14.3.3 Multiple array configuration**

In the case of observations from two or more arrays, a GBF- (see Section 9.13.5) or NEWS-type algorithm can be implemented. Recording one onset from each array with a BAZ

estimate is already sufficient to locate the source area. If the target fault zone is located between two arrays, which have a distance of about 200 km from each other, such an installation is sufficient to locate the main shock and the whole aftershock sequence on the fault zone within about 30 seconds after OT. Events, which are not located between the two arrays, will be located within 20 s plus the absolute travel time of the first P onset to one of the arrays.

This scenario of course assumes that the data of the two arrays are available in real time for fully automated array processing software. The location capabilities and precision will in general increase with a larger number of arrays. In such cases, different arrays may be also combined to monitor simultaneously different target areas.

## Acknowledgments

The authors thank John B. Young (Blacknest) for reprints of publications from the old days of array seismology in Great Britain, and Peter Bormann, Brian Kennett, Frank Scherbaum, and Lyla Taylor for critical reviews of the manuscript of this chapter in the first NMSOP edition of 2002. We thank Hanneke Paulssen, John Coyne and Peter Bormann for critically reviewing the new 2011 version of this chapter and for many helpful comments on how to improve it. Main parts of Section 9.14 were compiled during NORSAR's participation in the SAFER project, funded under the Sixth Framework Programme of the European Commission (Project Number 036935).

## References

- Abrahamson, N. A., Bolt, B. A., Darragh, R. B., Penzien J., and Tsai Y. B. (1987). The SMART 1 accelerograph array (1980-1987): a review. *Earthquake Spectra*, **3**, 263-287.
- Abt, A. (1907). *Vergleichung seismischer Registrierungen von Göttingen und Essen (Ruhr)*. Inaugural-Dissertation (Ph.D. thesis), Philosophische Fakultät, Georg-August-Universität zu Göttingen, Göttingen 1907, 26 pp. + curriculum vitae.
- Aki, K. (1957). Space and time spectra of stationary stochastic waves with special reference to microtremors. *Bull. Earthqu. Res. Inst.*, **35**, 415-456.
- Aki, K., and Richards, P. G. (1980). *Quantitative seismology – theory and methods*, Volume II, Freeman and Company, ISBN 0-7167-1059-5(v. 2), 609-625.
- Aki, K., Christoffersson, A., and Husebye, E. S. (1977). Determination of the three dimensional seismic structure of the lithosphere. *J. Geophys. Res.*, **82**, 277-296.
- Almendros, J., Ibáñez, J. M., Alguacil, G., and Del Pezzo, E. (1999). Array analysis using circular-wave-front geometry: an application to locate the nearby seismo-volcanic source. *Geophys. J. Int.*, **136**, 159-170.
- Almendros, J., Ibáñez, J. M., Alguacil, G., Del Pezzo, E., and Ortiz, R. (1997). Array tracking of the volcanic tremor source at Deception Island, Antarctica. *Geophys. Res. Lett.*, **24**, 3069-3072.
- Arai, H., and K. Tokimatsu (2004). S-wave velocity profiling by inversion of microtremor H/V spectrum. *Bull. Seism. Soc. Am.*, **94**, 53-63.

- Asten, M. W., and Henstridge, J. D. (1984). Array estimators and the use of microseisms for reconnaissance of sedimentary basins. *Geophys.*, **49**, (11), 1828-1837.
- Barber, N. S. (1958). Optimum arrays for direction finding. *New Zealand. J. of Sci.*, **1**, 35-51.
- Barber, N. S. (1959). Design of 'optimum' arrays for direction finding. *Electronic and Radio Engineer*, **36**, 333-232.
- Bard, P.-Y. (1999). Microtremor measurements: a tool for site effect estimation? In: Kudo, K., Okada H., and Sasatani T. (eds.). Proc. of the 2<sup>nd</sup> Int. Symp. on Effects of Surface Geology on Seismic Motion, Yokohama, Japan, *Balkema*, **3**, 1251-1279.
- Beauchamp, K. G. (ed.) (1975). *Exploitation of seismograph networks*, NATO Advanced Study Institutes Series E 111, Nordhoff Leiden, XII + 647 pp.
- Benndorf, H. (1905). Über die Art der Fortpflanzung der Erdbebenwellen im Erdinneren. 1. Mitteilung. *Sitzungsberichte der Kaiserlichen Akademie in Wien. Mathematisch-Naturwissenschaftliche Klasse*, **114**, *Mitteilungen der Erdbebenkommission*, **Neue Folge** **29**, 1-42.
- Benndorf, H. (1906). Über die Art der Fortpflanzung der Erdbebenwellen im Erdinneren. 2. Mitteilung. *Sitzungsberichte der Kaiserlichen Akademie in Wien. Mathematisch-Naturwissenschaftliche Klasse*, **115**, *Mitteilungen der Erdbebenkommission*, **Neue Folge** **31**, 1-24.
- Benz, H. M., Vidale, J. E., and Mori, J. (1994). Using regional seismic networks to study the Earth's deep interior. *EOS Trans. Am. Geophys. Soc.*, **75**, 225, 229.
- Berteussen, K. A. (1974). NORSAR location calibrations and time delay corrections. *NORSAR Sci. Rep.*, **2-73/74**, 41 pp.
- Berteussen, K. A. (1976). The origin of slowness and azimuth anomalies at large arrays. *Bull. Seism. Soc., Am.*, **66**, 719-741.
- Bettig, B., Bard, P.-Y., Scherbaum, F., Riepl, J., and Cotton, F. (2001). Analysis of dense array noise measurements using the modified spatial autocorrelation method (SPAC). Application to the Grenoble area. *Boll. Geof. Teor. Appl.*, **42**, (3/4), 281-304.
- Birtill, J. W., and Whiteway, F. E. (1965). The application of phased arrays to the analysis of seismic body waves. *Phil. Trans. Royal Soc. of London, Math. and Phys. Sciences*, **A-258**, (1091), 421-493.
- Bondár, I., North R. G., and Beall, G. (1999). Teleseismic slowness-azimuth station corrections for the International Monitoring System seismic network. *Bull. Seism. Soc. Am.*, **89**, 989-1003.
- Bonnefoy-Claudet, S., Cornou, C., Bard, P.-Y., and Cotton, F. (2004). Nature of noise wavefield. *SESAME Deliverable D13.08*, available at [http://sesame-fp5.obs.ujf-grenoble.fr/Delivrables/D13.08\\_finalreport\\_new.pdf](http://sesame-fp5.obs.ujf-grenoble.fr/Delivrables/D13.08_finalreport_new.pdf).
- Bonnefoy-Claudet, S., Cornou, C., Di Giuglio, G., Guillier, B., Jongmans, D., Köhler, A., Ohrnberger, M., Savvaidis, A., Roten, D., Scherbaum, F., Schissele, E., Vollmer D., and Wathelet, M. (2005). Report on the FK/SPAC capabilities and limitations. *SESAME Deliverable D19.06*, available at <http://sesame-fp5.obs.ujf-grenoble.fr/Delivrables/Del-D19-Wp06.pdf>.
- Braun, T., and Schweitzer, J. (2008). Spatial noise-field characteristics of a three-component small aperture test array in Central Italy. *Bull. Seism. Soc. Amer.*, **98**, 1876-1886, doi: 10.1785/0120070077.

- Braun, T., Schweitzer, J., Azzara, R. M., Piccinini, D., Cocco, M., and Boschi E. (2004). Results from the temporary installation of a small aperture seismic array in the Central Apennines and its merits for the local event detection and location capabilities. *Annals of Geophysics (Annali di Geofisica)*, **47**, 1557-1568.
- Bullen, K. E., and Bolt, B. A. (1985). *An introduction to the theory of seismology*. Cambridge University Press, 4<sup>th</sup> revised ed., xvii+499 pp., ISBN: 0-521-28389-2, 451-454.
- Bulletin of the Seismological Society of America (1990). Special Issue: "Regional seismic arrays and nuclear test ban verification". *Bull. Seism. Soc. Am.*, **80**, (6B), 1775-2281.
- Bungum, H., Husebye, E. S., and Ringdal F. (1971). The NORSAR array and preliminary results of data analysis. *Geophys. J. R. astr. Soc.*, **25**, 115-126.
- Bungum, H., Olesen, O., Pascal, C., Gibbons, S., Lindholm, C., and Vestøl, O. (2010). To what extent is the present seismicity of Norway driven by postglacial rebound? *J. Geolog. Soc. London*, **167**, 373-384, doi: 10.1144/0016-76492009-009.
- Burg, J. P. (1964). Three-dimensional filtering with an array of seismometers. *Geophysics*, **29**, 693-713.
- Burg, J. P. (1967). Maximum entropy spectral analysis. 37<sup>th</sup> Ann. Int. SEG Meeting, Soc. Explor. Geophys., Oklahoma City, Oklahoma, Oct. 31, 1967.
- Buttkus, B. (2000). *Spectral analysis and filter theory in applied geophysics*. Springer Verlag, xv+667 pp., ISBN 3-540-62674-3.
- Buttkus, B. (ed.) (1986). Ten years of the Gräfenberg array: defining the frontiers of broadband seismology. *Geologisches Jahrbuch E-35*, Hannover, 135 pp.
- Cansi, Y. (1995). An automatic seismic event processing for detection and location: P.M.C.C. method. *Geophys. Res. Lett.*, **22**, 1021-1024.
- Cansi, Y., Plantet, J. L., and Massion, B. (1993). Earthquake location applied to a min-array: k-spectrum versus correlation method. *Geophys. Res. Lett.*, **20**, 1819-1822.
- Capon, J. (1969a). High-resolution frequency-wavenumber spectrum analysis. *Proc. IEEE*, **57**, 1408-1418.
- Capon, J. (1969b). Investigation of long-period noise at the large aperture seismic array. *J. Geophys. Res.*, **12**, 3182-3194.
- Capon, J. (1973). Signal Processing and frequency-wavenumber spectrum analysis for a large aperture seismic array. In: Bolt, B. A. (ed.) *Methods in computational physics*, **13 Geophysics**. Academic Press, xiii+473 pp., IBN 0-12-460813-2, 1-59.
- Capon, J., Greenfield, R. J., and Kolker, K. J. (1967). Multidimensional maximum-likelihood processing of a Large Aperture Seismic Array. *Proc. IEEE*, **55**, 2, 192-211.
- Cessaro, R. K (1994). Sources of primary and secondary microseisms. *Bull. Seism. Soc. Am.*, **84**, 142-148.
- Chiu, H. C., Yeh, Y. T., Ni, S. D., Lee, L., Liu, W. S., Wen, C. F., and Liu, C. C. (1994). A new strong-motion array in Taiwan: Smart-2. *TAO*, **5**, 463-475.
- Chouet, B. (1996). New methods and future trends in seismological volcano monitoring. In: Scarpa, R., and Tilling, R. (eds.). *Monitoring and mitigation of volcano hazards*, Berlin, Springer-Verlag, 1996, xx + 841 pp., ISBN 3-540-60713-7, 23-97.



- Christoffersson, A., and Husebye, E. S. (2011). Seismic tomographic mapping of the Earth's interior – back to basics revisiting the ACH inversion. *Earth-Science Rev.*, **106**, 293-306, doi: 10.1016/j.earscirev.2011.02.007.
- Cornou, C., Ohrnberger, M., Boore, D. M., Kudo, K., and Bard, P.-Y. (2006). Derivation of structural models from ambient vibration array recordings: results from an international blind test. *3<sup>rd</sup> Int. Symp. on the Effects of Surf. Geol. on Seismic Motion*, Grenoble, France, 30 August – 1 September 2006.
- Dahlman, O., Mykkeltveit, S., and Haak, H. (2009). *Nuclear test ban – converting political visions to reality*. Springer Verlag, xviii+277 pp., ISBN 978-1-4020-6883-6, doi: 10.1007/978-1-4020-6885-0.
- Davies, D., Kelly, E. J., and Filson, J. R. (1971). The VESPA process for the analysis of seismic signals. *Nature*, **232**, 8-13.
- Del Pezzo, E., and Giudicepietro, F. (2002). Plane wave fitting method for a plane, small aperture, short period seismic array: a MATHCAD program. *Computers and Geosciences*, **28**, 59-64, doi: 10.1016/S0098-3004(01)00076-0.
- Delgado, L., López, C. C., Giner, J., Estévez, A., Cuenca, A., and Mouna, S. (2000). Microtremors as geophysical exploration tool: applications and limitations. *Pure Appl. Geophys.*, **157**, 1445-1462.
- Dietrich, K., Schweitzer, J., and Meier, T. (2001). S-velocities in the crust and uppermost mantle of northern Fennoscandia deduced from dispersion analysis of Rayleigh waves. *NORSAR Sci. Rep.*, **1-2001**, 71-76.
- Doornbos, D. J., and Husebye, E. S. (1972). Array analysis of PKP phases and their precursors. *Phys. Earth Planet. Inter.*, **6**, 387-399.
- Douglas, A. (2002). Seismometer arrays – their use in earthquake and test ban seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (eds.) (2002): *Handbook of Earthquake and Engineering Seismology*. Academic Press.
- Essen, H. H., Krüger, F., Dahm, T., and Grevemeyer, I. (2003). On the generation of secondary microseism observed in north and central Europe. *J. Geophys. Res.*, **108**, B10, ESE 15, 15 pp., doi: 10.1029/2002JB002318.
- Fäh, D., Kind, F., and Giardini, D. (2003). Inversion of local S-wave velocity structures from average H/V ratios, and their use for the estimation of site effects. *J. Seism.*, **7**, 449-467.
- Foti, S., Parolai, S., Albarello, D., and Picozzi, M. (2011). Application of surface-wave methods for seismic site characterization. *Surv. Geophys.*, doi: 10.1007/s10712-011-9134-2.
- Frankel, A., Hough, S., Friberg, P., and Rusby, R. (1991). Observations of Loma Prieta aftershocks from a dense array in Sunnyvale, California. *Bull. Seism. Soc. Amer.*, **81**, 1900-1922.
- Freiberger, W. F. (1963). An approximation method in signal detection. *Quart. J. App. Math.*, **20**, 373-378.
- Friedrich, A., Krüger, F., and Klinge, K., (1998). Ocean generated microseismic noise located with the Gräfenberg array. *J. Seism.*, **2**, 47-64.
- Frosch, R. A., and Green, P. E., Jr. (1966). The concept of a large aperture seismic array. *Proc. Royal Soc.*, **A-290**, 368-384.

- Fujiwara, H. (1997). Windowed f-k spectra of a three-dimensional wavefield excited by a point source in a two-dimensional multilayered elastic medium. *Geophys. J. Int.*, **128**, 571-584.
- Furumoto, M., Kunitomo, T., Inoue, H., Yamada, I., Yamaoka, K., Ikami, A., and Fukao, Y. (1990). Twin Sources of High-Frequency Volcanic Tremor of Izu-Oshima Volcano. Japan. *Geophys. Res. Lett.*, **17**, 25-27.
- Fyen, J. (1989). Event processor program package. *NORSAR Sci. Rep.*, **2-88/89**, 117-123.
- Fyen, J. (2001a). *NORSAR seismic data processing – user guide and command reference*. NORSAR (contribution 748), Kjeller, Norway.
- Fyen, J. (2001b). *NORSAR seismic detection processing – user guide and command reference*. NORSAR (contribution 731), Kjeller, Norway.
- Geller, R. J., and Mueller, C. S. (1980). Four similar earthquakes in central California. *Geophys. Res. Lett.*, **7**, 821-824.
- Gibbons, S. J., and Ringdal, F. (2006). The detection of low magnitude seismic events using array-based waveform correlation. *Geophys. J. Int.*, **165**, 149-166.
- Gibbons, S. J., Kväerna, T., and Ringdal, F. (2005). Monitoring of seismic events from a specific source region using a single regional array: A case study. *J. Seism.* **9**, 277-294.
- Gibbons, S. J., Kväerna, T., and Ringdal, F. (2009). Considerations in phase estimation and event location using small-aperture regional seismic arrays. *Pure Appl. Geophys.*, doi: 10.1007/s00024-009-0024-1.
- Gibbons, S. J., Schweitzer, J., Ringdal, F., Kväerna, T., Mykkeltveit, S., and Paulsen, B. (2011). Improvemnets to seismic monitoring of the European Arctic using three-component array processing at SPITS. *Bull. Seism. Soc. Am.*, **101**, 2737-2754, doi: 10.1785/0120110109.
- Goldstein, P., and Archuleta, R. J. (1991a). Deterministic frequency-wavenumber methods and direct measurements of rupture propagation during earthquakes using a dense array: theory and methods. *J. Geophys. Res.*, **96**, 6173-6185.
- Goldstein, P., and Archuleta, R. J. (1991b). Deterministic frequency-wavenumber methods and direct measurements of rupture propagation during earthquakes using a dense array: data analysis. *J. Geophys. Res.*, **96**, 6187-6198.
- Goldstein, P., and Chouet, B. (1994). Array measurements and modeling of sources of shallow volcanic tremor at Kilauea Volcano, Hawaii. *J. Geophys. Res.*, **99**, 2637-2652.
- Halldorsson, B., Sigbjornsson, S., and Schweitzer, J. (2009). ICEARRAY: the first small-aperture, strong-motion array on Iceland. *J. Seism.*, **13**, 173-178, doi: 10.1007/s10950-008-9133-z.
- Hao, C., and Zheng, Z. (2009). Slowness-azimuth corrections of teleseismic events for IMS primary arrays in China. *J. Seism.*, **13**, 437-448, doi: 10.1007/s10950-008-9137-8.
- Harjes, H.-P. (1990). Design and siting of a new regional seismic array in Central Europe. *Bull. Seism. Soc. Am.*, **80**, 1801-1817.
- Harjes, H.-P., and Henger, M. (1973). Array-Seismologie, *Zeitschrift für Geophysik*, **39**, 865-905.
- Harjes, H.-P., and Seidl, D. (1978). Digital recording and analysis of broad-band seismic data at the Gräfenberg (GRF-) array. *J. Geophys.*, **44**, 511-523.

- Harjes, H.-P., Schulte-Theis, H., Jost, M. L., and Schweitzer, J. (1996). Fast Earthquake Information Service (FEIS). *CSEM / EMSC, Newsletter*, **9**, 2-4, 1996.
- Haubrich, R. A. (1968). Array design. *Bull. Seism. Soc. Am.*, **58**, 977-991.
- Herrmann, R.-B. (1987). *Computer programs in seismology, Volume 4*, St Louis University.
- Horike, M. (1985). Inversion of phase velocity of long-period microtremors to the S-wave velocity structure down to the basement in urbanized areas. *J. Phys. Earth*, **33**, 59-96.
- Huang, B.-S. (2001). Evidence for azimuthal and temporal variation of the rupture propagation of the 1999 CH-Chi, Taiwan earthquake from dense seismic array observations. *Geophys. Res. Lett.*, **28**, 3377-3380.
- IBM (1972). *Seismic array design handbook*. ARPA Order No. 800, USAF Electronic System Division Contract No. F19628-68-C-0400, Gaithersburg, Maryland, August 1972, xv + 483 pp.
- Ibs von Seht, M., and Wohlenberger, R. (1999). Microtremor measurements used to map thickness of soft soil sediments. *Bull. Seism. Soc. Am.*, **89**, 250-259.
- Ishii, M., Shearer, P., and Vidale, J. (2005a). Extent, duration and speed of the 2004 Sumatra-Andaman earthquake imaged by the Hi-Net array. *Nature*, **435**, 933-936, doi: 10.1038/nature03675.
- Ishii, M., Shearer, P. M., Houston, H., and Vidale, J. E. (2005b). Teleseismic P wave imaging of the 26 December 2004 Sumatra-Andaman and 28 March 2005 Sumatra earthquake ruptures using the Hi-Net array. *J. Geophys. Res.*, **112**, B11307, doi: 10.1029/2006JB004700.
- Iyer, H. M. (1968). Determination of frequency-wave-number spectra using seismic arrays. *Geophys. J. R. astr. Soc.*, **16**, 97-117.
- J-Array Group (1993). The J-array program: system and present status. *J. Geomag. Geoelectr.*, **45**, 1265-1274.
- Johnson, D. H., and Dudgeon, D. E. (1993, 2002). *Array signal processing: concepts and techniques*. Prentice Hall signal processing series, PTR Prentice Hall, XIII+533 pp., ISBN 0-13-048513-6.
- Jongmans, D., Ohrnberger, M., and Wathelet, M. (2005). Recommendations for array measurements and processing. *Deliverable D24.13*, available at <http://sesame-fp5.obs.ujf-grenoble.fr/Delivrables/Del-D24-Wp13.pdf>.
- Journal of Seismology (2011). Special Issue: "Array seismology in Europe: recent developments and applications", *J. Seism.*, **15**, (3), 429-556.
- Kanasewich, E. R., Hemmings, C. D., and Alpaslan, T. (1973). N-th root stack nonlinear multichannel filter. *Geophysics*, **38**, 327-338.
- Keen, C. G., Montgomery, J., Mowat, W. M. H., and Platt, D. C. (1965). British seismometer array recording systems. *The Radio and Electronic Engineer*, **30**, 297-306.
- Kelly, E. J. Jr., and Levin, M. J. (1964). Signal parameter estimation for seismometer arrays. *M.I.T., Lincoln Laboratory Techn. Rep.*, **339**, Lexington, Massachusetts.
- Kennett, B. L. N. (2000). Stacking three-component seismograms. *Geophys. J. Int.*, **141**, 263-269.

- Kennett, B. L. N. (2002). *The seismic wavefield, Volume II: Interpretation of seismograms on regional and global scale*. Cambridge University Press, xii+534 pp., ISBN 0-521-00665-1, 163-170.
- Kind, F. (2002). *Development of microzonation methods: application to Basle, Switzerland*. Dissertation ETH No. 14548, Swiss Federal Institute of Technology, Zürich, 110 p.
- Krüger, F., and Ohrnberger, M. (2005a). Tracking the rupture of the  $M_w = 9.3$  Sumatra earthquake over 1150 km at teleseismic distance. *Nature*, **435**, 937-939, doi: 10.1038/nature0369kao.
- Krüger, F., and Ohrnberger, M. (2005b). Spatiuo-temporal source characteristics of the 26 December 2004 Sumatra earthquake as imaged by teleseismic broadband arrays. *Geophys. Res. Lett.*, **32**, L24312, doi: 10.1029/2005GL023939.
- Krüger, F., and Weber, M. (1992). The effect of low velocity sediments on the mislocation vectors of the GRF array. *Geophys. J. Int.*, **105**, 387-393.
- Krüger, F., Weber, M., Scherbaum, F., and Schlittenhardt, J. (1993). Double beam analysis of anomalies in the core-mantle boundary region. *Geophys. Res. Lett.*, **20**, 1475-1478.
- Kuleli, S., Zor, E., Türkelli, N., Sandvol, E., Seber, D., and Barazangi, M. (2001). The IMS Belbaşı seismic array (BRAR) in Central Turkey. *Seism. Res. Lett.*, **72**, 60-69.
- Kværna, T. (1989). On exploitation of small-aperture NORESS type arrays for enhanced P-wave detectability. *Bull. Seism. Soc. Am.*, **79**, 888-900.
- Kværna, T., and Doornbos, D. J. (1986). An integrated approach to slowness analysis with arrays and three-component stations. *NORSAR Sci. Rep.*, **2-85/86**, 60-69.
- Kværna, T., and Ringdal, F. (1986). Stability of various f-k estimation techniques. *NORSAR Sci. Rep.*, **1-86/87**, 29-40.
- Kværna, T., Schweitzer, J., Taylor, L., and Ringdal, F. (1999). Monitoring of the European Arctic using regional generalized beamforming. *NORSAR Sci. Rep.*, **2-98/99**, 78-94.
- Lacoss, R. T. (1965). Geometry and patterns of Large Aperture Seismic Arrays. *M.I.T., Lincoln Laboratory, Techn. Note*, **1965-64**, Lexington, Massachusetts.
- Lacoss, R. T., Kelly, E. J., Toksöz, and M. N. (1969). Estimation of seismic noise structure using arrays. *Geophysics*, **34**, 21-38.
- Levander, A., Humphreys, E. D., Ekstrom, G., Meltzer, A. S., and Shearer, P. M. (1999). Continental Assembly, Stability, and Instability: USArray: An Earth sciences tool for investigating North America. *EOS Trans. Am. Geophys. Un.*, **80**, 245-249.
- Levin, M. J. (1964). Maximum-likelihood array processing. *Lincoln Laboratory Report*, **31**, December 1964.
- Liaw, A. L., and McEvilly, T. V. (1979). Microseisms in geothermal exploration – studies in Grass Valley, Nevada. *Geophysics*, **44**, 1097-1115.
- Linville, F. A., and Laster, S. J. (1966). Numerical experiments in the estimation of frequency-wavenumber spectra of seismic events using linear arrays. *Bull. Seism. Soc. Am.*, **56**, 1337-1355.
- Malischewsky, P., and Scherbaum, F. (2004) Love's formula and H/V ratio (ellipticity) of Rayleigh waves. *Wave Motion*, **40**, (1), 57-67.

- Manchee, E. B. and Weichert, D. H. (1968). Epicentral uncertainties and detection probabilities from the Yellowknife seismic array data. *Bull. Seism. Soc. Am.*, **58**, 1359-1377.
- Muirhead, K. J. (1968). Eliminating false alarms when detecting seismic events automatically. *Nature*, **217**, 553-554.
- Muirhead, K. J., and Ram Datt (1976). The N-th root process applied to seismic array data. *Geophys. J. R. Astr. Soc.*, **47**, 197-210.
- Mykkeltveit, S. (1985). A new regional array in Norway: design work and results from analysis of data from a provisionally installation. In: A.U. Kerr (editor). *The VELA Program – A twenty-five year review of basic research*. Defence Advanced Research Project Agency, xviii + 964 pp., 1985, 546-553.
- Mykkeltveit, S., and Bungum, H. (1984). Processing of regional seismic events using data from small-aperture arrays. *Bull. Seism. Soc. Am.*, **74**, 2313-2333.
- Mykkeltveit, S., Åstebøl, K., Doornbos, D. J., and Husebye, E. S. (1983). Seismic array configuration optimization. *Bull. Seism. Soc. Am.*, **73**, 173-186.
- Mykkeltveit, S., Fyen, J., Ringdal, F., and Kværna, T. (1988). Spatial characteristics of the NORESS noise field and implications for array detection processing. *Phys. Earth Planet. Inter.*, **63**, 277-283.
- Nakamura, Y. (1989). A method for dynamic characteristics estimation of subsurface using microtremor on the ground surface. *Quat. Rep. RTRI*, **30**, 25-33.
- Nogoshi, M., and Igarashi, T. (1971). On the amplitude characteristics of microtremor (part 2). *J. Seism. Soc. Japan*, **24**, 26-40 (Japanese with English abstract).
- Ødegaard, E., Doornbos, D. J., and Kværna, T. (1990). Surface topographic effects at arrays and three-component stations. *Bull. Seism. Soc. Am.*, **80**, 2214-2226.
- Ohori, M., Nobata, A., and Wakamatsu, K. (2002). A comparison of ESAC and FK methods of estimating phase velocity using arbitrarily shaped microtremor arrays. *Bull. Seism. Soc. Am.*, **92**, (6), 2323-2332.
- Okada, H. (2003). The microseismic survey method. *Soc. Expl. Geophys. Japan, Geophys. Mon. Ser.*, **12** (translated by Koya Suto).
- Parolai, S., Bormann, P., and Milkereit, C. (2001). Assessment of the natural frequency of the sedimentary cover in the Cologne area (Germany) using noise measurements. *J. Earthq. Eng.*, **5**, 541-564.
- Parolai, S., Bormann, P., and Milkereit, C. (2002). New relationship between Vs, thickness of the sediments, and resonance frequency calculated by the H/V ratio of seismic noise for the Cologne area (Germany). *Bull. Seism. Soc. Amer.*, **92**, 2521-2527.
- Parolai, S., Picozzi, M., Richwalski, S. M., and Milkereit, C. (2005). Joint inversion of phase velocity dispersion and H/V ratio curves from seismic noise recordings using a genetic algorithm, considering higher modes. *Geophys. Res. Lett.*, **32**, doi: 10.1029/2004GL021115.
- Picozzi, M., Parolai, S., and Richwalski, S. M. (2005). Joint inversion of H/V ratios and dispersion curves from seismic noise: Estimating the S-wave velocity of bedrock. *Geophys. Res. Lett.*, **32**, doi: 10.1029/2005GL022878.

- Ringdal, F., and Kværna, T. (1989). A multi-channel processing approach to real time network detection, phase association, and threshold monitoring. *Bull. Seism. Soc. Am.*, **79**, 1927-1940.
- Roessler, D., Krueger, F., Ohrnberger, M., and Ehlert, L. (2010). Rapid characteristics of large earthquakes by multiple seismic broadband arrays. *Nat. Hazards Earth. Syst. Sci.*, **10**, 923-932, [www.nat-hazards-earth-syst-sci.net/10/923/2010/](http://www.nat-hazards-earth-syst-sci.net/10/923/2010/).
- Rost, S., and Thomas, C. (2002). Array Seismology: Methods and applications. *Rev. Geophys.*, **40**, (3), 1008, doi: 10.1029/2000RG000100.
- Rost, S., and Thomas, C. (2009). Improving seismic resolution through seismic arrays. *Surv. Geophys.*, **30**, 271-299, doi: 10.1007/s10712-009-9070-6.
- Sambridge, M. (1999a). Geophysical inversion with a neighbourhood algorithm – I. Searching a parameter space. *Geophys. J. Int.*, **138**, 479-494.
- Sambridge, M. (1999b). Geophysical inversion with a neighbourhood algorithm – II. Appraising the ensemble. *Geophys. J. Int.*, **138**, 727-746.
- Scherbaum, F. (2001). *Of Poles and Zeros: Fundamentals of Digital Seismology*. Kluwer Academic Publishers, 2nd edition, 265 pp. (including a CD-ROM by E. Schmidtke and F. Scherbaum, with examples written in Java).
- Scherbaum, F., Hinzen, K. G., and Ohrnberger, M. (2003). Determination of shallow shear wave velocity profiles in the Cologne, Germany area using ambient vibrations. *Geophys. J. Int.*, **152**, 597-612.
- Schimmel, M., and Paulssen, H. (1997). Noise reduction and detection of weak, coherent signals through phase-weighted stacks. *Geophys. J. Int.*, **130**, 497-505.
- Schmidt, R. O. (1986). Multiple emitter location and signal parameter estimation. *IEEE Trans. Ant. Prop.*, **AP-34**, 276-280.
- Schmidt, R. O., and Franks, E. R. (1986). Multiple source df signal processing: an experimental system. *IEEE Trans. Ant. Prop.*, **AP-34**, 281-290.
- Schulte-Pelkum, V., Earle, P. S., and Vernon, F. L. (2003). Strong directivity of ocean-generated seismic noise. *Geochem. Geophys. Geosyst.*, **5**, Q03004, doi: 10.1029/2003GC000520.
- Schulte-Theis, H., Jost, M. L., and Schweitzer, J. (1995). Fast earthquake information service (FEIS): optimized location of seismic events in Europe. In: *Advanced waveform research methods for GERESS recordings*. Scientific Report 4, 1 December 1994 – 30 June 1995, DARPA Grant MDA 972-93-1-0022, 17-28, 1995
- Schweitzer, J. (1994). Some improvements of the detector / SigPro-system at NORSAR. *NORSAR Sci. Rep.*, **2-93/94**, 128-139.
- Schweitzer, J. (1998). Tuning the automatic data processing for the Spitsbergen array (SPITS). *NORSAR Sci. Rep.*, **1-98/99**, 110-125.
- Schweitzer, J. (2001). Slowness corrections – one way to improve IDC products. *Pure Appl. Geophys.*, **158**, 375-396.
- Schweitzer, J. (2003). NORSAR's event warning system (NEWS). *NORSAR Sci. Rep.*, **1-2003**, 27-31.

- Schweitzer, J., and Krüger, F. (2011). Foreword (special issue “Array seismology in Europe: recent developments and applications”). *J. Seismology*, **15**, (3), 429-430, doi: 10.1007/s10950-011-9241-z.
- Schweitzer, J., and Kværna, T. (2002). Design study for the refurbishment of the SPITS array (AS72). *NORSAR Sci. Rep.*, **2–2002**, 65–77.
- Semin, K. U., Ozel, N. M., and Necmioglu, O. (2011). Detection and identification of low-magnitude seismic events near Bala, Central Turkey, using array-based waveform correlation. *Seism. Res. Lett.*, **82**, (1), 97-103.
- SESAME group (2004). Guidelines for the implementation of the H/V spectral ratio technique on ambient vibrations measurements, processing and interpretation. *SESAME Deliverable D23.12*, available at <http://sesame-fp5.obs.ujf-grenoble.fr/Delivrables/Del-D23-HV-User-Guidelines.pdf>.
- Shen, X., and Ritter, J. R. R. (2010). Small-scale heterogeneities below the Lanzhou CTBTO seismic array, from seismic wave field fluctuations. *J. Seism.*, **14**, 481-493.
- Somers, H., and Manchee, E. B. (1966). Selectivity of the Yellowknife seismic array. *Geophys. J. R. astr. Soc.*, **10**, 401-412.
- Spudich, P., and Bostwick, T. (1987). Studies of the seismic coda using an earthquake cluster as deeply buried seismograph array. *J. Geophys. Res.*, **92**, 10526-10546.
- Spudich, P., and Cranswick, E. (1984). Direct observation of rupture propagation during the 1979 Imperial Valley earthquake using a short baseline accelerometer array. *Bull. Seism. Soc. Am.*, **74**, 2083-2114.
- Tibuleac, I. M., and Stroujkova, A. (2009). Calibrating the Chiang Mai seismic array (CMAR) for improved event location. *Seism. Res. Lett.*, **80**, 579-590, doi: 10.1785/gssrl.80.4.579.
- Tokimatsu, K. (1997). Geotechnical site characterization using surface waves. In: Ishihara, K. (ed.): Proc. IS-Tokyo '95, 1<sup>st</sup> Intl. conf. Earthquake Geotech. Eng., Tokyo, Japan, 14-16 November 1995, *Balkema*, **3**, 1333-1368.
- Tokimatsu, K., Tamura, S., and Kojima, H. (1992). Effects of multiple modes on Rayleigh wave dispersion. *J. Geotech. Eng. ASCE*, **118**, (10), 1529-1543.
- van der Kulk, W., Rosen, F., and Lorenz, S. (1965). Large aperture array signal processing study. *IBM Final Report*, ARPA Contract SD-296, Rockville.
- Wang, J. (2002). A scheme for initial beam deployment for the International Monitoring System arrays. *Pure Appl. Geophys.*, **159**, 1005-1020.
- Wathelet, M. (2008). An improved neighbourhood algorithm: parameter conditions and dynamic scaling. *Geophys. Res. Lett.*, **35**, L09301, doi: 10.1029/2008GL033256.
- Wathelet, M., Jongmans, D., and Ohrnberger, M. (2004). Surface wave inversion using a direct search algorithm and its application to ambient vibration measurements. *Near Surf. Geophys.*, **2**, 211-221.
- Weichert, D. H. (1975). The role of medium aperture arrays: the Yellowknife system. In: Beauchamp, K. G., (1975), 167-195.
- Weichert, D. H., Manchee, E. B., and Whitham, K. (1967). Digital experiments at twice real-time speed on the capabilities of the Yellowknife seismic array. *Geophys. J. R. astr. Soc.*, **13**, 277-295.

- Whiteway, F. E. (1965). The recording and analysis of seismic body waves using linear cross arrays. *The Radio and Electronic Engineer*, **29**, 33-46.
- Whiteway, F. E. (1966). The use of arrays for earthquake seismology. *Proc. Royal Soc.*, **A-290**, 328-342.
- Willmore, P. L. (1979). *Manual of seismological observatory practice*. World Data Center A, Report **SE-20**, 165 pp., reprinted 1982.



# Chapter 10

## Seismic Data Formats, Archival and Exchange

(Version August 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch10)

Bernard Dost<sup>1)</sup>, Jan Zednik<sup>2)</sup>, Jens Havskov<sup>3)</sup>,  
Raymond J. Willemann<sup>4)</sup>, and Peter Bormann<sup>5)</sup>

<sup>1)</sup> Seismology Division KNMI, P.O. Box 201, 3730 AE De Bilt, The Netherlands; E-mail: Bernard.Dost@knmi.nl

<sup>2)</sup> Geophysical Institute AS CR, Bocni II/1401, 141 31 Prague 4, Czech Republic; E-mail: jzd@ig.cas.cz

<sup>3)</sup> University of Bergen, Department of Earth Science, Allégaten 41, N-5007 Bergen, Norway; E-mail: Jens.Havskov@geo.uib.no; Lars.Ottmoller@geo.uib.no;

<sup>4)</sup> IRIS Headquarters, 1200 New York Ave., Suite 800, Washington, DC 2005; E-mail: ray@iris.edu

<sup>5)</sup> Formerly GFZ German Research Center for Geosciences, Telegrafenberg, 14473 Potsdam, Germany; E-mail: pb65@gmx.net.

|                                                            | <b>page</b> |
|------------------------------------------------------------|-------------|
| <b>10.1 Introduction</b>                                   | 1           |
| <b>10.2 Parameter formats</b>                              | 3           |
| 10.2.1 HYPO71                                              | 4           |
| 10.2.2 HPOINVERSE                                          | 4           |
| 10.2.3 Nordic format                                       | 5           |
| 10.2.4 The GMS/IMS formats                                 | 6           |
| 10.2.5 The IASPEI Seismic Format (ISF)                     | 7           |
| <b>10.3 Digital waveform data</b>                          | 8           |
| 10.3.1 Data archival                                       | 10          |
| 10.3.2 Data exchange formats                               | 11          |
| 10.3.3 Formats for data base systems                       | 12          |
| 10.3.4 Continuous data protocols and formats               | 12          |
| <b>10.4 Some commonly encountered digital data formats</b> | 13          |
| <b>10.5 Format conversions</b>                             | 15          |
| 10.5.1 Why convert?                                        | 15          |
| 10.5.2 Ways to convert                                     | 16          |
| 10.5.3 Conversion programs                                 | 16          |
| <b>Acknowledgments</b>                                     | 18          |
| <b>Special references</b>                                  | 18          |
| <b>References</b>                                          | 19          |

### 10.1 Introduction

Seismology entirely depends on co-operation, both national and international cooperation. Only the accumulation of large sets of compatible high quality data in standardized formats from many stations and networks around the globe and over long periods of time will yield sufficiently reliable long-term results in event localization, seismicity rate and hazard

assessment, investigations into the structure and rheology of the Earth interior and other priority tasks in seismological research and applications.

For almost a century, only parameter readings taken from seismograms were exchanged with other stations and regularly transferred to national or international data centers for further processing. Because of the uniqueness of traditional paper seismograms and lacking opportunities for producing high-quality copies at low cost, original analog waveform data, cumbersome to handle and prone to damage or even loss, were rarely exchanged. The procedures for carefully processing, handling, annotating and storing such records have been extensively described in the 1979 edition of the *Manual of Seismological Observatory Practice* (Willmore, 1979) in the chapter *Station operation*. They are not repeated here. Also the traditional way of reporting parameter readings from seismograms to international data centers such as the U.S. Geological Survey National Earthquake Information Center (NEIC), the International Seismological Centre (ISC) or the European Mediterranean Seismological Centre (EMSC) are outlined in the old manual in detail in the section *Reporting output*. They have not changed essentially since then, although methods to exchange parameter data did change dramatically. On the other hand, respective working groups on parameter formats of the IASPEI and of its regional European Seismological Commission (ESC) have meanwhile debated for many years, how to make these formats more homogeneous, consistent and flexible so as to better accommodate also other seismologically relevant parameter information.

Any data report must, of course, follow a format known to the recipient in order to be successfully parsed. Some of the goals for any format are:

- *concise*                avoiding unnecessary expense in transmission and storage
- *complete*            providing all of the information required to use the data
- *transparent*        easily read by a person, perhaps without documentation
- *simple*                straightforward to write and parse with computer programs

Traditional formats for reporting parameter data sacrificed simplicity, transparency and even sometimes completeness in favor of the other goals. With the falling cost of data storage and exchange, modern formats more often sacrifice conciseness in favor of transparency and simplicity.

Modern parameter formats, in addition, are usually extensible and include “metadata”. An extensible format includes some way for new types of data to be introduced without either collecting all of the new information into unformatted comment strings or making messages with the new data types unreadable by old parsers. “Metadata” are information about the data, such as how and by whom the data were prepared. For waveform data the term metadata includes all information on the station and sensor/data acquisition system that produces the data.

The Telegraphic Format (TF), as documented in the *Manual of Seismic Observatory Practice* (Willmore, 1979) is an extreme example of a traditional format for reporting and exchanging parameter data. Since telex was very expensive compared with the modern communication costs, conciseness was the paramount goal even to the point of occasional ambiguity. The year of the data, for example, might be excluded if the recipient could probably infer it. The format was intended for use in an era when many stations were isolated and could report little more than their own phase readings, so event parameters such as hypocenter and magnitude are relegated to a secondary role. The TF incorporated further restrictions due to the special

limitations of telex messages, such as no lower-case letters and sometimes no control over line breaks.

A seismic network with modern, calibrated instruments can provide far more information than telegraphic format allows, while low-cost e-mail has eliminated the restrictions and high costs of telex messages. Consequently, since at least 1990 most seismic parameter data have been stored and exchanged in modern formats that are more complete, simpler and usually more transparent than the Telegraphic Format. But until recently there was no generally accepted standard modern format. A major step forward in this direction was made by the Group of Scientific Experts (GSE) organized by the United Nations Conference on Disarmament. It developed GSE/IMS formats (see 10.2.4) for exchanging parametric seismological data in tests of monitoring the Comprehensive Test Ban Treaty (CTBT) (see 10.2.4) which became popular also with other user groups. Seismological research, however, has a broader scope than the International Monitoring System (IMS) for the CTBT. Therefore, a new *IASPEI Seismic Format* (ISF), compatible with the IMS format but with essential extensions, has been developed and adopted by the Commission on Seismological Observation and Interpretation of the International Association of Seismology and Physics of the Earth's Interior at its meeting in Hanoi, August 2001. It is the conclusion of a 16-year process seeking consensus on a new format and fully exploits the much greater flexibility and potential of E-mail and Internet information exchange as compared to the older telegraphic reports (see 10.2.5). Since 2001 new developments have been introduced due to technology changes. The introduction of XML resulted in the requirement of a standard schema for the representation of parameter data: QuakeML (<https://quake.ethz.ch/quakeml/QuakeML>). QuakeML is a collaborative effort, which started within the NERIES project, and presents an open standard.

Digital waveform data, however, are nowadays by far the largest volume of seismic data stored and exchanged world-wide. The number of formats in existence and their complexity far exceeds the variability for parameter data. With the wide availability of continuous digital waveform data and unique communication technologies for world-wide transfer of such complete original data, their reliable exchange and archival has gained tremendous importance. Several standards for exchange and archival have been proposed, yet a much larger number of formats are in daily use. The purpose of the section on digital waveform data is to describe the international standards and to summarize the most often used formats. In addition, there will be a description of some of the more common conversion programs.

Due to the rapid increase of seismological stations, the practice of an international registry of station names (ISC, NEIC) became difficult to maintain. New definitions of station codes have been discussed within IASPEI and FDSN since 2005 and approved by IASPEI and its Commission on Seismic Observation and Interpretation (CoSOI) at the 2009 Cape Town General Assembly (see IS 10.3).

However, beforehand, a short description of the most common parameter formats will be given in the following section.

## 10.2 Parameter formats

Parameter formats deal with all earthquake parameters like origin times, hypocenter coordinates (longitude, latitude and depth), phase names and phase arrival times, maximum amplitudes and periods of different seismic phases or wave groups as recorded by different

types of seismographs, magnitudes calculated from such amplitude and period readings, etc. Until recently, there were no real standards, except the Telegraphic Format (TF) used for many years to report phase arrival, amplitude and period data to international agencies (Willmore, 1979; chapter “Reporting output”). The format is not used for processing. There have been attempts to modernize TF for many years through the IASPEI Commission of Practice (now the Commission on Seismological Observation and Interpretation) and as mentioned in the introduction, the IASPEI Seismic Format (ISF) was approved as a standard in 2001 (see section 10.2.5). Recently IASPEI (2005 and 2011) has adopted measurement standards for several widely used magnitudes. The standard nomenclature for reporting such magnitudes and related amplitude and period data in accordance with the ISF format have been outlined in IS 3.3 as well as in Chapter 3, sections 3.2.7.3 and 3.2.7.5. An example is given in Tab. 10.5 below. In practice, however, still many different formats are used and the most dominant ones have come from popular processing systems. In the following, some of the most well known formats will be briefly described. For complete description of the formats, the reader is referred to original manuals or publications.

### 10.2.1 HYPO71

The very popular location program HYPO71 (Lee and Lahr, 1975) has been around for many years and has been the most used program for local earthquakes. The format was therefore limited to work with only a few of the important parameters. Tab. 10.1 gives an example.

**Tab. 10.1** Example of an input file in HYPO71 format. Each line contains, from left to right: Station code (max 4 characters), E (emergent) or I (impulsive) for onset clarity, polarity (C – compression; D – dilatation), year, month, day, and time (hours, minutes, seconds, hundredth of seconds) for P-Phase, second for S-phase (seconds and hundredth of seconds only), S-phase onset and, in the last column, duration. The blank space between ES and duration has been used for different purposes like amplitude. The last line is a separator line between events and contains control information.

|         |                 |         |     |
|---------|-----------------|---------|-----|
| FOO EPC | 96 6 6 64848.47 | 62.67ES | 136 |
| MOL EPC | 96 6 6 64849.97 | 65.87ES | 144 |
| HYA EP  | 96 6 6 64856.78 | 78.07ES | 135 |
| ASK EP  | 96 6 6 649 2.94 | 34.72ES | 183 |
| BER EPC | 96 6 6 649 7.56 | 36.61ES |     |
| EGD EPD | 96 6 6 649 5.76 | 40.53ES |     |
|         | 10 5.0          |         |     |

The format is rather limited since only P or S phase names can be used and the S-phase is referenced to the same hour-minute as the P-phase and the format cannot be used with teleseismic data. However, the format is probably one of the most popular formats ever for local earthquakes. The HYPO71 program has seen many modifications and the format exists in many forms with small changes.

### 10.2.2 HYPOINVERSE

Following the popularity of HYPO71, several other popular location programs followed like Hypoinverse (Klein, 1978) and Hypoellipse (Lahr, 1989). Tab. 10.2 gives an example of the input format for Hypoinverse.

**Tab. 10. 2** Example of the Hypoinverse input format. Note that year, month, day, hour, min is only given in the header and only one phase is given per line.

```

96 6 60648
FOO EPC 48.5 136
FOO ES 62.7
MOL EPC 50.0 144
MOL EPC 50.9
MOL ES 65.9

```

### 10.2.3 Nordic format

In the 1980's, this was one of the first attempts to create a more complete format for data exchange and processing. The initiative came from the need to exchange and store data in Nordic countries and the so-called Nordic format was agreed upon among the 5 Nordic countries. The format later became the standard format used in the SEISAN data base and processing system and is now widely used. The format tried to address some of the shortcomings in HYPO71 format by being able to store nearly all parameters used, having space for extensions and useful for both input and output. An example is given in Tab. 10.3.

**Tab. 10. 3** Example of Nordic format. The data is the same as seen in Tabs. 10.1 and 10.2. The format starts with a series of header lines with type of line indicated in the last column (80) and the phase lines are following the header lines with no line type indicator. There can be any number of header lines including comment lines. The first line gives among other things, origin time, location and magnitudes, the second line is the error estimate, the third line is the name of the corresponding waveform file and the fourth line is the explanation line for the phases (type 7). The abbreviations are: STAT: Station code, SP: component, I: I or E, PHAS. Phase, W: Weight, D: polarity, HRMM SECON: time, CODA: Duration, AMPLIT: Amplitude, PERI: Period, AZIMU: Azimuth at station, VELO: Apparent velocity, SNR: Signal-to-noise ratio, AR: Azimuth residual of location, TRES: Travel time residual, W: Weight in location, DIS: Epicentral distance in km and CAZ: Azimuth from event to station.

```

1996 6 6 0648 30.4 L 62.635 5.047 15.0 TES 13 1.4 3.0CTES 2.9LTES 3.0LNA01
GAP=267 5.92 18.8 43.0 31.8 -0.5630E+03 0.8720E+03 -0.3916E+03E
1996-06-06-0647-46S.TEST__011 6
STAT SP IPHASW D HRMM SECON CODA AMPLIT PERI AZIMU VELO SNR AR TRES W DIS CAZ7
FOO SZ EP C 648 48.47 136 -0.110 116 180
FOO SZ ESG 649 2.67 0.710 116 180
FOO SZ E 649 2.89 426.4 0.3 116 180
MOL SZ EP C 648 49.97 144 -0.310 129 92
MOL SZ EPG C 648 50.90 0.410 129 92
MOL AZ E 649 5.86 129 92
MOL SZ ESG 649 5.87 0.410 129 92
MOL SZ E 649 6.98 328.6 0.6 129 92
HYA SZ EP 648 56.78 135 0.810 174 159
HYA SZ IP D 648 56.78 0.810 174 159
HYA SZ EPG D 648 57.56 0.110 174 159
HYA SZ ESG 649 18.07 0.610 174 159
NRA0 SZ Pn 0649 24.03 309.6 8.5 139 5 -0.410 403 119
NRA0 SZ Pg 0649 32.60 305.6 7.285.2 1 0.410 403 119
NRA0 SZ Lg 0650 22.05 302.0 4.016.0 -1 -0.410 403 119

```

## 10.2.4 The GSE/IMS formats

Also in the late eighties, a new format was created for exchange of data within the International Monitoring System (IMS) of the Comprehensive Test Ban Treaty Organization (CTBTO), formally called the GSE (Group of Scientific Experts) parameter format. The format used in the GSE's Technical Test 3 (GSETT-3) was designated GSE2.0 and came to be used even by seismologists uninvolved in CTBT monitoring.

Following GSETT-3, a significantly revised format was originally designated GSE2.1, but renamed IMS1.0 when it was adopted for use by the International Data Center planned to monitor the CTBT when it enters into force. The format IMS1.0 is similar in structure to the Nordic format, however more complete in some respects and lacking features in other respects. A major difference is that the line length can be more than 80 characters long, which is not the case for any of the previously described formats. After SEISAN, it is the first major format for which completeness or readability was recognized as a more important design goal than conciseness. The IMS1.0 format is available from the ISC web site via <http://www.isc.ac.uk/standards/isf> and [ftp://ftp.isc.ac.uk/pub/isf/isf\\_ext.pdf](ftp://ftp.isc.ac.uk/pub/isf/isf_ext.pdf). IMS1.0 has been used extensively for data exchange within the institutions participating in the IMS and also been used for data exchange outside IMS like in the popular AutoDRM system ([http://www.seismo.ethz.ch/prod/autodrm/index\\_EN](http://www.seismo.ethz.ch/prod/autodrm/index_EN)) and for transmission to international centers. However, IMS has been used less as a processing format than HYPO71 and Nordic formats.

**Tab. 10.4** An example of the IMS1.0 parameter format. The example contains the same data as given in Tabs. 10.1 to 10.3. The first lines are message information etc. The remaining lines are more or less self-explanatory. Note that more information, with a higher accuracy, can be given for each phase (like magnitude) than in the Nordic format. On the other hand, information like component and event duration is missing. These are added in the new ISF format.

```
BEGIN GSE2.0
MSG_TYPE DATA
MSG_ID 1900/10/19_1711 ISR_NDC
DATA_TYPE ORIGIN GSE2.0
EVENT 00000001
  Date      Time      Latitude Longitude      Depth      Ndef Nsta Gap      Mag1  N      Mag2  N
rms  OT_Error      Smajor Sminor Az      Err      mdist  Mdist  13 267      ML 2.9 8
1.40 +- 5.92      0.0 0.0 0 +- 31.8 1.04 4.84 +-0.3
Sta Dist EvAz      Phase Date      Time      TRes Azim AzRes Slow SRes Def SNR
Amp Per Mag1 Mag2 Arr ID
F00 1.04 180.0 mc P 1996/06/06 06:48:48.5 -0.1 T
F00 1.04 180.0 m SG 1996/06/06 06:49:02.7 0.7 T
F00 1.04 180.0 m 1996/06/06 06:49:02.9
426.4 0.30 ML 3.2 00000003 (from previous line)
MOL 1.16 92.0 mc P 1996/06/06 06:48:50.0 -0.3 T
MOL 1.16 92.0 mc PG 1996/06/06 06:48:50.9 0.4 T
MOL 1.16 92.0 m 1996/06/06 06:49:05.9
MOL 1.16 92.0 m SG 1996/06/06 06:49:05.9 0.4 T
MOL 1.16 92.0 m 1996/06/06 06:49:07.0
NRA0 3.62 119.0 m Pn 1996/06/06 06:49:24.0 -0.4 309.6 5.0 8.5 TAS 13.9
(from previous line)
NRA0 3.62 119.0 m Pg 1996/06/06 06:49:32.6 0.4 305.6 1.0 7.2 TAS 85.2
NRA0 3.62 119.0 m Lg 1996/06/06 06:50:22.0 -0.4 302.0 -1.0 4.0 TAS 16.0
(from previous line)
STOP
```

## 10.2.5 The IASPEI Seismic Format (ISF)

The need for a common and widely accepted parameter format for comprehensive seismological data exchange has led to the IASPEI Seismic Format (ISF), adopted as standard in August 2001. ISF conforms to the IMS1.0 standard but has essential extensions for reporting additional types of data. This allows the contributor to include complementary data considered to be important for seismological research and applications by the IASPEI Commission on Seismological Observation and Interpretation. The format looks almost like the IMS1.0 example in Tab. 10.4 above, except for the extensions. The ISF has been comprehensively tested at the ISC and NEIC and incompatibilities have been eliminated. The detailed description of the ISF is available from the ISC home page and kept up-to-date there (see [ftp://ftp.isc.ac.uk/pub/isf/isf\\_ext.pdf](ftp://ftp.isc.ac.uk/pub/isf/isf_ext.pdf)). It is not reproduced in this manual.

Consensus on the ISF was reached partly by including many optional items, so the format is not as simple as some alternatives. Despite this, the completeness, transparency, extensibility and metadata of ISF are expected to make it very widely used. Wide use of ISF will bring back the advantages of a generally accepted standard so that it becomes easier to exchange data, re-use data collected for past projects, and employ programs developed elsewhere.

In the Information Sheets IS 10.1 and IS 10.2, examples are given how event parameter data and unassociated parameter readings by seismic stations are reported according to the IMS format with ISF extensions.

Tab. 10.5 below gives an abridged example of an USGS/NEIC Hydra System data printout in the IMS1.0 format of calculated event parameters and station parameter readings. Yet, for brevity, several columns in the original listing have been left out in order to highlight the new, more differentiated way of magnitude, amplitude and period reporting. It helps to recognize easily which amplitudes and magnitudes have been measured according to the IASPEI (2005) recommended standards and which ones not, whether the amplitudes are displacement or velocity amplitudes and at what “amplitude phase” time these parameters have been “picked”. This assures later on an unambiguous reproduction or check of these parameter determinations.

**Tab. 10.5** Cut-out of an event and station parameter plot in IMS1.0/ISF parameter format produced by the USGS/NEIC HYDRA automatic location and analysis system. Note the IASPEI (2005) standard magnitudes mb, mb\_Lg, mB\_BB, Ms\_BB and Ms\_20. BB stands for unfiltered velocity broadband records, 20 for surface waves with periods of 20±2 s measured in WWSSN-LP filtered records. In contrast, mb and mb\_Lg are measured on WWSS-SP filtered records. Letters preceding the magnitude symbol in the “amplitude-phase name” column stand for: I - IASPEI standard, A - displacement amplitude in nm and V - velocity amplitude in nm/s. The next column gives the time at which the respective amplitudes and periods have been measured. Additionally, NEIC also determines several non-standard magnitudes.

```
BEGIN IMS1.0
MSG_TYPE DATA
MSG_ID 30000Z9N_43 HYDRA_ORANGE
DATA_TYPE BULLETIN IMS1.0:SHORT
The following is an UNCHECKED, FULLY AUTOMATIC LOCATION from the USGS/NEIC Hydra
System
Event      15694
```

| Date       | Time        | Err  | RMS  | Latitude | Longitude | Smaj | Smin | Az  | Depth |
|------------|-------------|------|------|----------|-----------|------|------|-----|-------|
| 2009/09/29 | 17:48:13.26 | 2.94 | 1.45 | -15.5267 | -172.0703 | 6.5  | 5.8  | 133 | 28.4  |

| Magnitude | Err | Nsta | Author    | OrigID |
|-----------|-----|------|-----------|--------|
| Mb_Lg     | 6.0 | 0.5  | 1 NEIC    | 0      |
| Ms_VX     | 8.2 | 0.1  | 23 NEIC   | 0      |
| mb        | 7.2 | 0.0  | 243 NEIC  | 0      |
| Mwp       | 7.8 | 0.0  | 179 NEIC  | 0      |
| mB_BB     | 7.7 | 0.0  | 246 NEIC  | 0      |
| Ms_BB     | 8.3 | 0.1  | 134 NEIC  | 0      |
| Mwb       | 7.7 | 0.0  | 97 NEIC   | 0      |
| Ms_20     | 8.0 | 0.0  | 161 NEIC  | 0      |
| Mwc       | 8.1 | 0.0  | 71 NEIC   | 0      |
| M         | 7.9 | 0.1  | 1064 NEIC | 0      |
| Mwp       | 7.9 | 0.0  | 4 PMR     | 0      |

| Sta  | Dist  | EvAz  | Phase   | Time         | Amp       | Per   | Magnitude | ArrID   |
|------|-------|-------|---------|--------------|-----------|-------|-----------|---------|
| KNTN | 12.68 | 1.6   | IAmb_Lg | 17:55:00.090 | 7785.9    | 0.98  | Mb_Lg 6.0 | BHZIU00 |
| KNTN | 12.68 | 1.6   | IVMs_BB | 17:56:03.398 | 2169276.1 | 10.00 | Ms_BB 7.7 | LHZIU00 |
| TARA | 22.39 | 317.2 | P       | 17:53:11.829 |           |       |           | BHZIU00 |
| TARA | 22.39 | 317.2 | IAmb    | 17:53:45.055 | 12080.1   | 1.25  | mb 7.2    | BHZIU00 |
| TARA | 22.39 | 317.2 | MMwp    | 17:53:53.279 | 1728974.9 | 41.45 | Mwp 7.6   | BHZIU00 |
| TARA | 22.39 | 317.2 | IVmB_BB | 17:54:07.329 | 206688.6  | 5.40  | mB_BB 7.8 | BHZIU00 |
| OUZ  | 23.45 | 210.6 | P       | 17:53:22.710 |           |       |           | HHZNZ10 |
| OUZ  | 23.45 | 210.6 | IAmb    | 17:53:47.450 | 11261.4   | 1.22  | mb 7.3    | HHZNZ10 |
| OUZ  | 23.45 | 210.6 | IVmB_BB | 17:53:49.880 | 314772.2  | 9.74  | mB_BB 8.0 | HHZNZ10 |
| OUZ  | 23.45 | 210.6 | IAMS_20 | 18:01:02.679 | 6314.2    | 20.00 | Ms_20 8.1 | LHZNZ10 |
| OUZ  | 23.45 | 210.6 | IVMs_BB | 18:02:47.679 | 3821858.4 | 16.00 | Ms_BB 8.4 | LHZNZ10 |
| OUZ  | 23.45 | 210.6 | AMs_VX  | 18:02:54.449 | 6326077.5 | 15.00 | Ms_VX 8.3 | LHZNZ10 |

## 10.3 Digital waveform data

Many different formats for digital data are used today in seismology. For a summary and the abbreviations used, see the following sections. Most formats can be grouped into one of the following five classes:

1. Local formats in use at individual stations, networks or used by a particular seismic recorder (e.g., ESSTF, PDR-2, BDSN, GDSN; mini-SEED).
2. Formats used in standard analysis software (e.g., SEISAN, SAC, AH, BDSN; mini-SEED).
3. Formats designed for data exchange and archiving (SEED, GSE).



4. Formats designed for database systems (CSS, SUDS).
5. Formats for real time data transmission (IDC/IMS, Earthworm; mini-SEED).

Use of the term "designed" in describing Class 3 and 4 formats is intentional. It is usually only at this level that very much thought has been given to the subtleties of format structure which result in efficiency, flexibility, and extensibility.

The four classes (1-4) show a hierarchical structure. Class 4 forms a superset of the others, meaning that classes 1-3 can be deduced from it. The same argument applies to class 3 with respect to classes 1 and 2. Nearly all format conversions performed at seismological data centers are done to move upwards in the hierarchy for the purpose of data archiving and exchange with other data centers. Software tools are widely available to convert from one format to another and particularly upwards in the hierarchy.

This hierarchy also explains why there are so many formats. The design of class 1 formats depends on the manufacturer of the data acquisition system. In the early days of digital seismometry, display and analysis software was often proprietary and marketed specifically for a certain manufacturer's equipment and data format. There was no real need for manufacturers to adhere to a standard recording format, until users began to realize the advantages of exchanging data with other seismologists and discovered that this was quite difficult unless the other party was using the same hardware and/or software.

Station operators who were not satisfied with the proprietary analysis software supplied with the procured data acquisition systems started to convert data from Class 1 formats into the Class 2 formats which were used by more powerful and widely available analysis packages such as SAC. These programs usually provide subroutines that make conversion from local formats fairly easy. Analysis packages (e.g., SeisGram) which are developed around a Class 1 format (BDSN in this case) implicitly offer their format preference as a candidate for a new standard in Class 2, but it hardly matters as long as the necessary software tools are available to convert to and from the data exchange formats.

The GDSN (Global Digital Seismic Network) format began as a Class 1 format, but because it was used by an important global seismograph network (DWWSSN, SRO) it became accepted as a de facto standard for data exchange (Class 3). The beginning of widespread international data exchange within the FDSN (Federation of Digital Seismic networks) and GSE (Group of Scientific Experts) groups in the late 1980's revealed the GDSN format's weaknesses in this role and put in motion the process of defining more capable exchange formats.

The volume of commonly available digital seismic data continues to increase dramatically. It increased from 600 MB annually in 1980 to 300 GB in 1992 and today we are talking about many terabytes. Database systems, which are specially designed to handle these large datasets, have therefore begun to appear as a superset of the standard data exchange formats. The SUDS system was an example of this type of format, however today it is little used.

In the 1990's, several activities (e.g., the GSETT-3 experiment and the U.S. National Seismograph Network (USNSN)) emerged featuring real-time exchange of seismological data, and interest focused on formats which are suitable for such applications. In the late 1990's, this idea was carried farther by systems such as Earthworm, which implemented format-independent protocols. Earthworm also is designed to exchange data across a peer

network of multiple, independent nodes, as well as in a traditional network of dependent nodes with a centralized collection and distribution center.

Following is a brief description of some of the classes of formats as defined above.

### 10.3.1 Data archival

Data archival requires the storage of complete information on station, channel(s) and the structure of the data. Most existing formats are designed to provide part of the information. Most archival formats presently in use do include info on station and channel, but are not always complete in the description of the data. What we envisage is demonstrated through several features in the Standard for the Exchange of Earthquake Data (SEED) format:

- Data Description Language (DDL)
- Reference to byte order
- Response information

The DDL is defined to enable the data itself to be stored in any data format (integer, binary, compressed). The language consists of a number of keys defining, for example, the applied compression scheme, number of bytes per sample, mantissa and gain length in bits and the use of the sign convention. The reader interprets the DDL and knows exactly how to deal with the data. The advantage of the DDL is that the original data structure can be maintained and is known. A disadvantage is that readers will have to interpret the DDL and have less performance in reading. However, the decoding information is available directly with the data and this is extremely important, since data are collected on platforms having different byte orders. In SEED the byte order of the original data is defined in the header, so the reader will be able to decide whether bytes should be swapped.

In most archival formats, response information can be supplied in terms of poles and zeroes. Fewer efforts are undertaken to give the FIR filter coefficients in the header, although they are accounted for in the definition of SEED and GSE2.X. A problem occurs when a description of the instrument response is given only in measured amplitude and phase data as a function of frequency, as is the case in the GSE1.0 format. Also, the GSE2.X does not specify the minimum requirement. The main purpose of the response information is to correct for instrument response and thus the user will have to find the best fitting poles and zeroes to the given response. Although tools are available to calculate poles and zeroes from frequency, amplitude and phase data (e.g., in Preproc), results from the multiple inversion of the discrete frequency, amplitude and phase data will be different from the original data.

The deployment of large mobile arrays consisting of heterogeneous instrumentation is an important research tool. Data archival of these data is important. Although there is a tendency to store the data in a common format, the responses of sensors and data acquisition systems are often poorly known. It is recommended to pay attention to this issue before the experiment starts!

Finally, an issue in data archival is the responsibility of the data quality and the mechanism of reporting data errors. The network/station operator is responsible for the quality of the original data. However, the data may be subjected to format conversion at a remote data center. This last stage could introduce errors and it is the originator of the data, which must be

responsible for data quality and should agree on the final conversion, if such a conversion is done externally.

### 10.3.2 Data exchange formats

The data exchange formats are closely related to the way data is exchanged. Therefore, these formats are described separately in this section. Essentially, any format can be used for exchange, however the idea of an exchange format is to make it easy to send it electronically, to have a minimum standard of content and be readable on all computer platforms. Today, the de-facto standard within the FDSN is the mini-SEED or SEED format ([http://www.iris.edu/manuals/SEED\\_chpt1.htm](http://www.iris.edu/manuals/SEED_chpt1.htm)).

At present, there are many different techniques in use to exchange data, either between data users and data centers or between data centers. Since the beginning of the 21<sup>st</sup> century techniques to transmit and exchange data in real-time became very popular and efficient, and the volume of continuous waveform data has grown very fast. With this rapid increase of continuous data, maintenance and adaptation of existing techniques and development of new techniques for requesting selected datasets only. Nowadays, the most common format which is used in real-time and request mechanisms for data exchange is (mini-)SEED. An overview of existing data request techniques is given below.

|                         | Technique                                                                      | Advantage                                     | Disadvantage                            |
|-------------------------|--------------------------------------------------------------------------------|-----------------------------------------------|-----------------------------------------|
| <b>Indirect on-line</b> | autoDRM, NetDC, breq-fast                                                      | email based<br>(no connection time)           | small volume or<br>download through ftp |
| <b>Direct on-line</b>   | ftp, WWW, DRM<br>(Wilber/FARM); DHI,<br>ArcLink, webservices,<br>QuakeExplorer | direct access, enables<br>easy data selection | slow for large data<br>volumes          |
| <b>Off-line</b>         | CD-ROM (DVD)                                                                   | direct access                                 | no real-time data                       |

Indirect on-line data exchange is arranged through (automated) *Data Request Managers* (DRMs) where the request mechanism is based on email traffic. One of the earliest systems was AutoDRM (<http://seismo.ethz.ch/autodrm>), a standardized protocol using the GSE defined syntax. One step further is the implementation of a communication protocol for exchange between data centers in such a way that a user only has to send one request to a nearby data center node. His/her request is then automatically routed through the data centers that may contribute to the requested data set. Such a protocol is the NetDC system (Casey and Ahern, 1996). A similar system, called ArcLink, was developed by GFZ (GEOFON) using a more simple communication protocol over a TCP/IP socket, making it a direct on-line request mechanism

One basic problem in using email as the transport mechanism is the restricted data volume that can be exchanged. Also, the format sometimes will have to be ASCII. The format issue is taken care of in the GSE format, although in the description of the AutoDRM protocol it is mentioned that also a format like SEED can be used. The only difference is that the user is requested to get the data through anonymous ftp (pull) or the data is pushed into an anonymous ftp area defined by the user. The AutoDRM system at the Orfeus Data Centre

(ODC) supports both the SEED and GSE format in data exchange. Both NetDC and ArcLink provide mini-SEED waveform data.

*Direct on-line access to data* is arranged at the ODC (<http://www.orfeus-eu.org/>) for example through ftp, a web-interface and web services. Web services provide machine-accessible data services typically accessed through the standard HTTP protocol. In many cases these services are publicly exposed so that users may access them directly through their own applications. Other data centers, like IRIS-DMC and GEOFON, have similar web-based techniques. An example of an application using web services is the NERIES portal (<http://www.neries-eu.org>) through which a user can search for earthquake data in the EMSC database and download waveform data through the ODC.

Internet speed presently may still be limiting the usefulness of this direct on-line data exchange, especially since the volumes that are to be transferred may be huge. One major advantage of direct on-line availability of the data is the capability to make a selection out of the vast amount of digital data. Procedures are presently under development to increase the power of these selection tools, for example through the above mentioned portal and webservices developments.

*Off-line data access* provides complete, quality controlled data that are locally available at each institute in the form of CD-ROMs or DVD's. The completeness and quality control takes time and CD-ROMs and DVD's have a limited storage capacity.

### **10.3.3 Formats for data base systems**

Formats for data base systems are specially designed and no details will be given here. Examples of such formats are CSS and the derived "IDC Database Schema" (see Information Sheet IS 10.3) and SUDS.

### **10.3.4 Continuous data protocols and formats**

With better communication systems, real time transmission of digital data has become common. There is no internationally agreed upon protocol for this and equipment manufacturers often use their own formats. However, the number of protocols is limited, and the most common systems are SeedLink, SCREAM, NAQS, CD1.x, Antelope, EarthWorm, NRTS and LISS. A number of open-source and commercial data exchange systems exist between the different systems, and may be included in the protocol software or may be found on the appropriate web sites. Complete documentation for the CD-1.0 protocol used for the transmission of IMS data as described under 10.2.4. is now being replaced by the CD-1.1 protocol. Descriptions for both can be found on the secure web site <https://www2.ctbto.org> (authorized users only) and openly on <http://www.geoinstr.com/pub/manuals/343r02p.pdf>.

At present, a large number of channels from a station or array of stations can be transmitted in near-real time using a single connection. Digital data are provided in compressed or uncompressed format and with or without authentication signatures. The protocol uses units of information called frames to establish or alter a connection and to exchange data between

the sender and the receiver. Only one frame is being transmitted or received at any instance. A time-out is used in case of lost connection.

*Establishing connections.* The sender initiates the connection with the receiver to a pre designated IP address and port by sending a Connection Request Frame. The receiver validates the authenticity of the sender and provides a new port and Internet Protocol (IP) address in a Port Assignment Frame. The sender drops the original connection and connects to the assigned IP address and port that is subsequently used for all data transfer.

*Transmitting data.* After the connection is established, the sender sends a Data Format Frame, which describes the format of the subsequent Data Frames. The sender can then send Data Frames data. The Data Format Frame provides information about itself and about Data Frames that will follow. The Data Frame contains the raw time series data. Each Data Frame has a single Data Frame Header and multiple channel sub-frames.

*Altering connections.* Either sender or the receiver can alter the connection through the exchange of Alert Frames. The receiver sends the Alert Frame to notify the sender to use a different port. The sender uses Alert Frames to notify the receiver that the communication will cease or that a new data format is about to be used.

*Terminating connections.* Typically, an established connection remains active and in use until the sender or receiver terminates it for maintenance or reconfiguration. The connection can be intentionally terminated by sending an Alert Frame. Unintentional termination due to a slow or failed communications system is detected after the time-out period.

Documentation for the Earthworm protocol can be found on the USGS website <http://folkworm.ceri.memphis.edu/ew-doc/> .

## 10.4 Some commonly encountered digital data formats

This section gives an alphabetical list of common formats in use by several recording or analysis programs. For each format some description is given. For several more specific, proprietary, now outdated or rarely used and platform formats we refer to Chapter 10, pages 12-16, in the first edition of NMSOP (DOI: 10.2312/GFZ.NMSOP\_r1\_ch10; <http://nmsop.gfz-potsdam.de> and <http://www.isc.ac.uk/standards> ).

### AH

Class: 2

The Ad Hoc (AH) format is used in the AH waveform analysis software package developed at Lamont Doherty Geological Observatory, New York, USA. This package also supports a number of conversion tools.

### CSS

Class: 2,4

The Center for Seismic Studies (CSS) Database Management System (DBMS) was designed to facilitate storage and retrieval of seismic data for seismic monitoring of test ban treaties [CSS]. The seismic data separate into two categories: Waveform data and parametric data.

For the parameter data, the design utilizes a commercial relational database management system. Information is stored in relations that resemble flat, two-dimensional tables as in the ISF format (see Information Sheet IS 10.1). The description of waveform data is physically separated from the waveform data itself. The index to the waveform archive is maintained within the relational database. Data are stored in plain files, called non-DBMS files. Each non-DBMS file is indexed by a relation that contains information describing the data and the physical location of the data in the file system. Each waveform segment contains digital samples from only one station and one channel. The time of the first sample, the number of samples and the sample rate of the segment are noted in an index record. The index also defines in which file and where in the file the segment begins, and it identifies the station and channel names. A calibration value at a specified frequency is noted. The index records are maintained in the **wfdisc** relation. Each **wfdisc** record describes a specific waveform segment and contains an id number to designate detailed information on the station and instrumentation of the trace.

## **GSE**

### Class 3

The format proposed by the Group of Scientific Experts (GSE format) has been extensively used with the GSETT projects on disarmament. The GSE2.1, now renamed IMS1.0, is the most recent version. The manual can be downloaded from [http://www.seismo.ethz.ch/prod/autodrm/manual/provisional\\_GSE2.1.pdf](http://www.seismo.ethz.ch/prod/autodrm/manual/provisional_GSE2.1.pdf).

A GSE2.1 waveform data file consists of a waveform identification line (WID2) followed by the station line (STA2), the waveform information itself (DAT2), and a checksum of the data (CHK2) for each DAT2 section (Provisional GSE 2.1 Message Formats & Protocols, 1997). The default line length is 132 bytes. No line may be longer than 1024 bytes. The response data type allows the complete response to be given as a series of response groups than can be cascaded. Response description is made up of the CAL2 identification line plus one or more of the PAZ2, FAP2, GEN2, DIG2 and FIR2 response sections in any order.

Waveform identification line WID2 gives the date and time of the first data sample; the station, channel and auxiliary codes; the sub-format of the data, the number of samples and sample rate; the calibration of the instrument represented as the number of nanometers per digital count at the calibration period; the type of the instrument, and the horizontal and vertical orientation.

Line STA2 contains the network identifier, latitude and longitude of the station, reference coordinate system, elevation and emplacement depth.

Data section after DAT2 may be in any of six different sub-formats recognized in the GSE2.1 waveform format: INT, CM6, CM8, AUT, AU6, and AU8. INT is a simple ASCII sub-format, "CM" sub-formats are for compressed data and "AU" sub-formats are for authentication data. All represent the numbers as integers and therefore can be sent by email.

A checksum CHK2 must be provided in the GSE2.1 format. The checksum is computed from integer data values prior to converting them to any of the sub-formats.

## **SAC**

Class 2

Seismic Analysis Code (SAC) is a general-purpose interactive program designed for the study of time sequential signals [SAC]. Emphasis has been placed on analysis tools used by research seismologists. A SAC data file contains a single data component recorded at a single seismic station. Each data file also contains a header record that describes the contents of that file. Certain header entries must be present (e.g., the number of data points, the file type, etc.). Others are always present for certain file types (e.g., sampling interval, start time, etc. for evenly spaced time series). Other header variables are simply informational and are not used directly by the program. Although the SAC analysis software only runs on Unix platforms and the general format is binary, there is also an ASCII version that can be used on any platform.

## **SUDS**

Class: 1,2,4 Platform: PC

SUDS stands for “The Seismic Unified Data System”. The SUDS format was launched to be a more well thought out format useful for both recording and analysis and independent of any particular equipment manufacturer. The format has seen widespread use, but has lost some momentum, partly because it is not made platform independent.

## **10.5 Format conversions**

### **10.5.1 Why convert?**

Ideally, we should all use the same format. Unfortunately, as the previous descriptions have shown, there are a large number of formats in use. With respect to parameter formats, one can get a long way with HYPO71, Nordic and GSE/ISF formats for which converters are available, such as in the SEISAN system. For waveform formats, the situation is more complicated. First of all, there are many different formats, and, since most are binary, there is the added complication that some will work on some computer platforms and not on others. This is particularly a problem with binary files containing real numbers as for example, the SeisGram format. Additional problems are: Some formats have seen slight changes and exist in different versions, different formats have different contents so not all parameters can be transferred from one format to another and conversion programs might not be fully tested for different combinations of data.

Many processing systems require a higher level format than the often primitive recording formats so that is probably the most common reason for conversion, and a similar reason is to move from one processing system to another.

The SEED format has become a success for archival and data exchange. Initially, it was not very useful for processing purposes, however now it is more widely used also for processing. Sometimes it is important to be able to move down in the hierarchy to be able to use a particular processing system. Thus, the main reasons for format conversion can be summarized as:

- Moving upwards in the hierarchy of formats for the purpose of data archiving and exchange

- Moving downward from the archive and exchange formats for analysis purposes
- Moving across the hierarchy for analysis purposes
- Moving from one computer platform to another

### 10.5.2 Ways to convert

There are essentially two ways of converting. The first is to request data from a data center in a particular format or to log into a data center and use one of their conversion programs. The other more common way is to use a conversion program on the local computer. Such conversion programs are available both as free standing software and as part of processing systems. Equipment manufactures will often supply at least a program to convert recorder data to some ASCII format and very often also to the more standard format MiniSeed.

### 10.5.3 Conversion programs

Since conversion programs are often related to analysis programs, we list in Tab. 10.6 some of the better-known analysis systems and the format they use directly.

**Tab.10.6** Examples of popular analysis programs.

| Program        | Author(s)                           | Input format(s)                  | Output format(s) |
|----------------|-------------------------------------|----------------------------------|------------------|
| Geotool        | J.Coyne                             | CSS, SAC, GSE                    | CSS, SAC, GSE    |
| SAC            | IRIS                                | SAC                              | SAC              |
| SEISAN         | J.Havskov, L. Ottemöller,<br>P.Voss | SEISAN, GSE                      | SEISAN, GSE, SAC |
| SeismicHandler | K.Stammler                          | miniSEED, GSE, AH, ESSTF,<br>GCF | GSE, miniSEED    |
| SNAP           | M.Baer                              | SED, GSE                         | SED, GSE         |

An overview of available format conversion programs can be found on the ORFEUS Web pages under ORFEUS Seismological Software Library (<http://orfeus.knmi.nl/wirjung.groups/wg4/index.html>). Here we present just a few packages in alphabetical order. Only those programs are mentioned which are able to read at least one of the formats mentioned in sub-chapter 10.4.

#### Codeco

Program **codeco** was written by U. Kradolfer and modified by K. Stammler and K. Koch. Input files can be in SAC binary or ASCII, or GSE formats. Output formats are: integer or compressed GSE1.0 or GSE2.0, SAC binary or ASCII, and miniSEED. **Codeco** is available through the SZGRF software library (<ftp://ftp.szgrf.bgr.de/pub/software> ).



## GSE to SEED

Program **gse2seed**, developed by R. Sleeman (Orfeus Data Centre, de Bilt), converts a GSE2.X file to the SEED2.3 format. Multiple traces are handled. For each WID2 section, the GSE file must contain corresponding data types STATION, CHANNEL and RESPONSE. This program was originally developed to convert both metadata and waveform data, but is maintained nowadays only to convert GSE metadata into dataless SEED. Conversion of GSE waveform data into mini-SEED can be done by **gse2mseed**, developed by Chad Trabant (IRIS DMC).

## PASSCAL package

The PASSCAL package was written by P. Friberg, S. Hellman, and J. Webber, developed on SUN under SunOs4.1.4, compiled under Solaris 2.4 and higher and also under LINUX. It converts RefTek to SEG Y and miniSEED. Program **pql** provides a quick and easy way to view SEG Y, SAC, miniSEED or AH seismic data. **pql** operates in the X11 window environment. The package is available from the PASSCAL instrument center (<http://www.passcal.nmt.edu>) at New Mexico Tech., Socorro.

## Preproc

**Preproc** has been designed to assist the seismologist who wishes to analyze large sets of raw digital data that need to be preprocessed in some standard way prior to the analysis. Preproc was written by Miroslav Zmeskal for the ISOP project in the period 1991-1993. It was rewritten recently in the object-oriented form. As a by-product, **preproc** can perform data conversion from GSE / PITSA ISAM to GSE / PITSA ISAM. In the near future new input/output formats will be implemented (ESSTF, miniSEED). **preproc** was successfully compiled on HP, SUN, Linux and DOS. Program package **preproc** and a detailed manual are available through the ORFEUS Seismological Software Library

## Rdseed

**Rdseed** reads from the input tape or file in the SEED format. According to the command line function option specified by the user, **rdseed** will read the volume and recover the volume table of contents ( -c), the set of abbreviation dictionaries ( -a), or station and channel information and instrument response table ( -s). In order to extract data from the SEED volume for analysis by other packages, the user must run **rdseed** in user prompt mode (without any command line options). As data is extracted from the SEED volume, **rdseed** looks at the orientation and sensitivity of each channel and corrects the header information on request. Implemented output formats are (option d): SAC, AH, CSS 3.0, miniSEED and SEED. A Java version of rdseed is to be released in 2001. **Rdseed** was developed by Dennis O'Neill and Allen Nance, IRIS DMC.

**SeedStuff** is a set of basic programs provided by the GEOFON DMS software library in Potsdam (<ftp://ftp.gfz-potsdam.de/pub/home/st/GEOFON/software>) to process and compile raw data from Quanterra, Comserv and RefTek data loggers. The goal is to check and extract

data from station files/tapes to miniSEED files and to assemble miniSEED files to full SEED volumes. The SeedStuff package was written by Winfried Hanka and compiled on the SUN, HP and Linux. The following tools are available:

extr\_qic: extracts multiplexed raw Quanterra station tapes to demultiplexed miniSEED files containing only one station / stream / component  
extr\_file: like extr\_qic for multiplexed miniSEED, RefTec files  
extr\_fseed: disassemble full SEED tapes. SEED headers are skipped, data are stored into station / stream / component files  
check\_seed: checks the contents of miniSEED data files or tapes  
check\_qic: analysis the contents of a Quanterra data tape  
copy\_seed: assembles a full SEED volumes from miniSEED files for a given set of station / stream / component defined in the copy\_seed.cfg configuration file  
make\_dlsv: generates a dataless (header only) SEED volume for a set of station/stream/component defined in copy\_seed.cfg

## **SEED to GSE**

Recently (2009), a SEED to GSE converter was developed at ORFEUS Data Center (ODC), mainly to support the GSE format within AutoDRM at ODC. This tool will become available at the ORFEUS software library.

## **SEISAN**

The SEISAN analysis system has about 40 conversion programs, mostly from some binary format to SEISAN. The SEISAN format can then be converted to any standard format like SEED, SAC or GSE. SEISAN has format converters for most recorders on the market including Kinemetrics, Nanometrics, Teledyne, GeoSig, Reftek, Lennartz, Güralp and Sprengnether.

## **Acknowledgement**

The authors acknowledge with thanks the careful review by Bruce Presgrave of the US Geological Survey. It has improved both the language of the original draft and provided useful references to the Earthworm system. In addition we acknowledge discussion and comments from Lars Ottemöller and Reinoud Sleeman.

## **Special references**

- [CSS] Anderson, J., W. Farrell, K. Garcia, J. Given, and H. Swanger, Center for Seismic Studies Version 3 Database: Schema Reference Manual, SAIC Technical Report C90-01, 1990.
- [IDC3.4.1] Formats and Protocols for Messages, Rev. 3, 2001.
- [GSE] Provisional GSE2.1 Message Formats & Protocols, 1997. Operations Annex 3, GSETT-3.

- [LEN] SAS-58000 User's Guide and Reference Manual, 1986. Lennartz electronic GmbH
- [SAC] W.C. Tapley & J.E. Tull, 1992. SAC - Seismic Analysis Code. LLNL, Regents of the University of California
- [SEED] Standard for the Exchange of Earthquake Data, 1992. Reference Manual, SEEDFormat v2.3, FDSN, IRIS, USGS

## References

- Casey, R., and Ahern, T. (1996). Technical manual for Networked Data Centers (NETDC) protocol (*IRIS, internal report*) or <http://www.iris.washington.edu/manuals/netdc/>
- IASPEI (2005 and 2011). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. Preliminary version October 2005, updated version September 2011; <http://www.iaspei.org/commissions/CSOI.html>.
- Klein, F. W. (1978). Hypocenter location program HYPOINVERSE. *U.S. Geol. Surv. Open-File Report*. **78-694**.
- Lahr, J. C. (1981) REFERENCE IS MISSING!
- Lee, W. H. K., and Lahr, J. C. (1975). HYPO71 (revised): A computer program for determining hypocenter, magnitude and first motion pattern of local earthquakes. U.S. Geological Survey Open-File Report 75-311, 116 pp.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.

|

| Title   | Data-Type Bulletin IMS1.0: Short                                                                                                                  |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | Raymond J. Willemann, IRIS, 1200 New York Ave, NW, Suite 800, Washington DC 20005, U.S.A.; E-mail: <a href="mailto:ray@iris.edu">ray@iris.edu</a> |
| Version | September 2001; DOI: 10.2312/GFZ.NMSOP-2_IS_10.1                                                                                                  |

Below an example is given of an ISF Bulletin which is in accordance with the IMS1.0 format of the International Monitoring System and the current version of the document defining ISF extensions of IMS1.0 (see ISC home page <http://www.isc.ac.uk/standards/isf>).

The example includes two events because an example of only one event would fail to show how consecutive events are to follow each other. The first event is small; with only a few data. This makes it possible to realize the typical way of data presentation at a glance. The second event is larger and this examples shows better how multiple magnitudes and event parameters are to be included.

An example can not, of course, explain which elements are required and which are optional, nor give the units in which each parameter is required to be given. Thus, it is essential for an agency intending to write ISF bulletins to read the format description as well as look at an example. The format description in this case includes the IMS1.0 (a.k.a. GSE2.1) documentation, as well as the extensions of IMS1.0 that constitute the ISF (see IS 10.2). The final ISF description of the extensions document has been posted on the ISC web site (<http://www.isc.ac.uk/standards/isf>) as a PDF document.

In the ISC Bulletin for the time period 01-09-1999 06:00:00 to 01-09-1999 06:45:00 the following 2 events were found:

**Event 1847567 Turkey**

| Date       | Time        | Err  | RMS  | Latitude | Longitude | Smaj | Smin | Az | Depth | Err  | Ndef | Nsta | Gap | mdist | Mdist | Qual | Author | OrigID  |
|------------|-------------|------|------|----------|-----------|------|------|----|-------|------|------|------|-----|-------|-------|------|--------|---------|
| 1999/09/01 | 06:03:51.10 |      |      | 40.7170  | 30.7580   |      |      |    | 14.0  |      |      | 5    |     |       |       |      | ISK    | 2576986 |
| 1999/09/01 | 06:03:50.70 | 3.88 | 0.58 | 40.7830  | 30.7590   | 41.2 | 13.6 | 0  | 5.0   | 23.4 | 5    | 5    |     | 0.47  | 1.80  | m i  | ISC    | 3325719 |

| Magnitude | Err | Nsta | Author | OrigID  |
|-----------|-----|------|--------|---------|
| MD 2.6    |     |      | ISK    | 2576986 |

| Sta | Dist | EVaz  | Phase | Time         | TRES | Azim | AzRes | Slow | SRes | Def | SNR | Amp | Per | Qual | Magnitude | ArrID    |
|-----|------|-------|-------|--------------|------|------|-------|------|------|-----|-----|-----|-----|------|-----------|----------|
| MDU | 0.47 | 132.3 | Pg    | 06:04:00.100 | -0.0 |      |       |      |      | T__ |     |     |     | _i   |           | 40036747 |
| EYL | 0.51 | 244.9 | Pg    | 06:04:01.200 | 0.4  |      |       |      |      | T__ |     |     |     | _i   |           | 40036748 |
| EYL | 0.51 | 244.9 | SG    | 06:04:07.700 |      |      |       |      |      | T__ |     |     |     | _e   |           | 40036749 |
| YLV | 1.08 | 258.9 | Pg    | 06:04:11.800 | -0.4 |      |       |      |      | T__ |     |     |     | _e   |           | 40036750 |
| IZI | 1.08 | 246.0 | PN    | 06:04:12.100 | -0.2 |      |       |      |      | T__ |     |     |     | _e   |           | 40036751 |
| CTT | 1.80 | 282.4 | PN    | 06:04:22.800 | 0.2  |      |       |      |      | T__ |     |     |     | _i   |           | 40036752 |

**Event 1717835 Central Mid-Atlantic Ridge**

| Date                          | Time        | Err  | RMS  | Latitude | Longitude | Smaj | Smin | Az  | Depth | Err | Ndef | Nsta | Gap | mdist | Mdist  | Qual | Author | OrigID  |  |
|-------------------------------|-------------|------|------|----------|-----------|------|------|-----|-------|-----|------|------|-----|-------|--------|------|--------|---------|--|
| 1999/09/01                    | 06:42:34    |      |      | 3.0000   | -34.0000  |      |      |     |       |     |      |      |     |       |        |      | NAO    | 3125978 |  |
| 1999/09/01                    | 06:42:41.63 | 0.21 | 0.85 | 4.6760   | -32.6130  | 6.7  | 4.3  | 152 | 10.0F |     | 125  | 125  | 64  | 25.28 | 157.07 | se   | NEIC   | 2932984 |  |
| (#PARAM SCALAR_MOMENT=2.1E16) |             |      |      |          |           |      |      |     |       |     |      |      |     |       |        |      |        |         |  |
| 1999/09/01                    | 06:42:44.00 | 0.25 |      | 4.3726   | -32.2802  | 18.7 | 9.3  | 133 | 33.0  |     | 31   |      |     | 27.20 | 82.59  | ke   | LDG    | 3015245 |  |
| 1999/09/01                    | 06:42:45.23 | 1.14 | 0.80 | 4.7536   | -32.7216  | 28.4 | 23.0 | 50  | 18.8  | 3.5 | 19   | 11   | 171 | 43.40 | 157.20 | uk   | EIDC   | 3004603 |  |
| 1999/09/01                    | 06:42:49.00 | 0.80 |      | 5.1800   | -32.7000  | 11.1 | 11.1 | -1  | 15.0F |     |      |      |     |       |        | se   | HRVD   | 2932985 |  |

| (#CENTROID) | (#MONTENS | sc    | MO     | fCLVD | MRR   | MTT   | MPP   | MRT   | MTP | NST      | Author |
|-------------|-----------|-------|--------|-------|-------|-------|-------|-------|-----|----------|--------|
| (#          | eMO       | eCLVD | eRR    | eTT   | ePP   | eRT   | eTP   | ePR   | NCO | duration |        |
| (#          | 16        | 4.40  | -5.010 | 2.560 | 2.450 | 0.000 | 1.220 | 0.000 | 12  | HRVD     |        |
| (#          |           |       | 0.650  | 1.330 | 0.840 |       | 0.430 |       | 17  |          |        |

| (+ Data Used: GSN.) | (#FAULT_PLANE | Typ    | Strike | Dip    | Rake | NP | NS | Plane | Author |
|---------------------|---------------|--------|--------|--------|------|----|----|-------|--------|
| (#                  | BDC           | 226.00 | 45.00  | -90.00 |      |    |    | HRVD  |        |
| (+                  | BDC           | 46.00  | 45.00  | -90.00 |      |    |    | HRVD  |        |

| (#PRINAX | sc | T_val | T_azim | T_pl | B_val | B_azim | B_pl | P_val | P_azim | P_pl  | Author |
|----------|----|-------|--------|------|-------|--------|------|-------|--------|-------|--------|
| (#       | 16 | 3.72  | 136.00 | 0.00 | 1.29  | 46.00  | 0.00 | -5.01 | 180.00 | 90.00 | HRVD   |

| 1999/09/01                   | 06:42:41.81 | 0.28 | 0.89 | 4.6780 | -32.5870 | 6.1 | 4.9 | 0 | 10.0F |  | 137 | 177 | 25.30 | 157.06 | m i | ISC | 3325720 |
|------------------------------|-------------|------|------|--------|----------|-----|-----|---|-------|--|-----|-----|-------|--------|-----|-----|---------|
| (#PARAM pP_DEPTH=13.30+0.80) |             |      |      |        |          |     |     |   |       |  |     |     |       |        |     |     |         |

| Magnitude | Err | Nsta | Author | OrigID  |
|-----------|-----|------|--------|---------|
| mb 4.9    |     | 48   | NEIC   | 2932984 |
| MSZ 4.6   |     | 59   | NEIC   | 2932984 |
| Mw 5.1    |     |      | HRVD   | 2932985 |
| Mb 4.8    | 0.2 | 25   | LDG    | 3015245 |
| Ms 4.1    | 0.2 | 7    | LDG    | 3015245 |
| Mb 5.0    |     |      | NAO    | 3125978 |
| mb 4.4    | 0.1 | 10   | EIDC   | 3004603 |
| mmsle 4.1 | 0.1 | 6    | EIDC   | 3004603 |

3

| Sta   | Dist  | EVaz  | Phase | Time         | TRes | Azim  | AzRes | Slow | SRes | Def | SNR  | Amp   | Per   | Qual | Magnitude | ArrID        |
|-------|-------|-------|-------|--------------|------|-------|-------|------|------|-----|------|-------|-------|------|-----------|--------------|
| SCHQ  | 57.13 | 337.2 | P     | 06:52:29.600 | -1.2 | 123.7 |       | 8.3  | T    |     | 13.7 | 15.4  | 0.79  | —    | mb        | 5.1 42115750 |
| SCHQ  | 57.13 | 337.2 | SP    | 06:52:35.675 | 0.5  | 139.5 |       | 8.8  | T    |     | 7.1  | 15.0  | 0.92  | —    |           | 42115751     |
| PTCC  | 57.45 | 36.2  | P     | 06:52:33.100 | -0.1 |       |       |      | T    |     |      |       |       | —    |           | 42839620     |
| VOY   | 57.57 | 36.8  | P     | 06:52:34.200 | 0.1  |       |       |      | T    |     |      |       |       | e    |           | 42808147     |
| GRF   | 58.08 | 32.0  | P     | 06:52:37.600 | 0.0  |       |       |      | T    |     |      | 38.7  | 1.30  | e    | mb        | 5.3 40722849 |
| GRF   | 58.08 | 32.0  | SP    | 06:52:43.200 | 1.1  |       |       |      | —    |     |      |       |       | e    |           | 40722850     |
| GRF   | 58.08 | 32.0  | LR    | 06:52:41.190 | -1.1 |       |       |      | T    |     |      | 300.0 | 18.50 | —    | MS        | 4.4 40722851 |
| AAM   | 58.74 | 317.5 | P     | 06:52:41.190 | -1.1 |       |       |      | T    |     |      | 19.7  | 0.80  | e    | mb        | 5.2 40722765 |
| AAM   | 58.74 | 317.5 | LR    | 06:52:41.190 | -1.1 |       |       |      | T    |     |      | 380.0 | 19.00 | —    | MS        | 4.5 40722766 |
| PTJ   | 58.76 | 37.7  | P     | 06:52:39     | -3.4 |       |       |      | T    |     |      |       |       | e    |           | 36882044     |
| MOX   | 58.85 | 31.2  | P     | 06:52:42     | -1.0 |       |       |      | T    |     |      |       |       | e    |           | 45438984     |
| MOX   | 58.85 | 31.2  | L     | 07:15:53     |      |       |       |      | T    |     |      |       |       |      |           | 45438985     |
| GEC2  | 58.95 | 33.9  | P     | 06:52:41.800 | -1.8 |       |       |      | T    |     |      | 2.4   | 1.00  | e    | mb        | 4.2 40722841 |
| GERES | 58.95 | 33.9  | P     | 06:52:42.125 | -1.5 | 246.4 |       | 4.2  | T    |     | 25.5 | 4.5   | 0.95  | —    | mb        | 4.5 42115739 |
| GERES | 58.95 | 33.9  | SP    | 06:52:48.825 | 0.7  | 228.7 |       | 5.1  | —    |     | 13.4 | 9.4   | 1.10  | —    |           | 42115740     |
| WCI   | 59.03 | 312.2 | PFAKE | 06:52:50.000 | 5.6  |       |       |      | —    |     |      |       |       | —    |           | 40722992     |
| WCI   | 59.03 | 312.2 | LR    | 06:52:50.000 | 5.6  |       |       |      | —    |     |      | 480.0 | 22.00 | —    | MS        | 4.6 40722993 |
| KHC   | 59.03 | 33.5  | P     | 06:52:44.000 | -0.2 |       |       |      | T    |     |      |       |       | e    |           | 44689671     |
| WVT   | 59.50 | 309.5 | P     | 06:52:45.930 | -1.7 |       |       |      | T    |     |      | 10.9  | 1.00  | e    | mb        | 4.8 40722999 |
| WVT   | 59.50 | 309.5 | LR    | 06:52:45.930 | -1.7 |       |       |      | —    |     |      | 530.0 | 19.00 | —    | MS        | 4.7 40723000 |
| CLL   | 59.94 | 31.2  | P     | 06:52:50     | -0.5 |       |       |      | T    |     |      |       |       | —    |           | 45569564     |
| PRU   | 60.03 | 33.1  | P     | 06:52:50.600 | -0.5 |       |       |      | T    |     |      | 13.0  | 1.20  | i    | mb        | 4.8 40722943 |
| PRU   | 60.03 | 33.1  | SP    | 06:52:56.200 | 0.6  |       |       |      | —    |     |      |       |       | i    |           | 40722944     |
| BRG   | 60.19 | 32.0  | P     | 06:52:52.700 | 0.5  |       |       |      | T    |     |      | 18.0  | 1.40  | i    | mb        | 4.9 40722793 |
| BRG   | 60.19 | 32.0  | SP    | 06:52:57.800 | 1.2  |       |       |      | —    |     |      |       |       | i    |           | 40722794     |
| BRG   | 60.19 | 32.0  | LR    | 06:52:57.800 | 1.2  |       |       |      | —    |     |      |       |       | —    |           | 40722795     |
| BRG   | 60.19 | 32.0  | LR    | 06:52:57.800 | 1.2  |       |       |      | —    |     |      | 160.0 | 24.00 | —    | MS        | 4.1 40722795 |
| BRG   | 60.19 | 32.0  | LR    | 06:52:57.800 | 1.2  |       |       |      | —    |     |      | 170.0 | 24.00 | —    | MS        | 4.0722796    |
| BRG   | 60.19 | 32.0  | LR    | 06:52:57.800 | 1.2  |       |       |      | —    |     |      | 130.0 | 24.00 | —    | MS        | 4.2 40722797 |
| OXF   | 60.26 | 307.2 | PFAKE | 06:53:00.000 | 7.1  |       |       |      | —    |     |      |       |       | —    |           | 40722933     |
| OXF   | 60.26 | 307.2 | LR    | 06:53:00.000 | 7.1  |       |       |      | —    |     |      | 970.0 | 21.00 | —    | MS        | 4.9 40722934 |
| VRAC  | 60.81 | 34.6  | P     | 06:52:55.800 | -0.6 |       |       |      | T    |     |      |       |       | —    |           | 41194520     |
| VAY   | 61.17 | 44.9  | P     | 06:53:00.500 | 1.5  |       |       |      | T    |     |      |       |       | e    |           | 40722986     |
| MORC  | 61.58 | 34.5  | P     | 06:53:01.500 | -0.1 |       |       |      | T    |     |      |       |       | —    |           | 41194519     |
| OKC   | 61.95 | 34.6  | P     | 06:53:03.800 | -0.4 |       |       |      | T    |     |      |       |       | e    |           | 44689672     |
| SUR   | 62.63 | 130.1 | P     | 06:53:09.940 | 1.0  | 7.5   |       | 8.8  | T    |     | 3.7  | 9.9   | 0.97  | —    | mb        | 4.9 42115752 |
| UALR  | 62.66 | 306.8 | P     | 06:53:07.230 | -1.9 |       |       |      | T    |     |      |       |       | e    |           | 40722983     |
| CCM   | 62.68 | 310.6 | P     | 06:53:06.830 | -2.3 |       |       |      | T    |     |      | 10.4  | 0.70  | e    | mb        | 5.1 40722802 |
| CCM   | 62.68 | 310.6 | LR    | 06:53:06.830 | -2.3 |       |       |      | —    |     |      | 380.0 | 21.00 | —    | MS        | 4.5 40722803 |
| OJC   | 63.07 | 34.8  | P     | 06:53:12.300 | 0.8  |       |       |      | T    |     |      |       |       | e    |           | 36778884     |
| JFWS  | 63.47 | 316.1 | PFAKE | 06:53:20.000 | 5.7  |       |       |      | —    |     |      |       |       | —    |           | 40722872     |
| JFWS  | 63.47 | 316.1 | LR    | 06:53:20.000 | 5.7  |       |       |      | —    |     |      | 440.0 | 19.00 | —    | MS        | 4.7 40722873 |
| KONO  | 63.59 | 22.4  | PFAKE | 06:53:30.000 | 15.2 |       |       |      | —    |     |      |       |       | —    |           | 40722880     |
| KONO  | 63.59 | 22.4  | LR    | 06:53:30.000 | 15.2 |       |       |      | —    |     |      | 200.0 | 19.00 | —    | MS        | 4.3 40722881 |
| LBTB  | 63.73 | 120.7 | P     | 06:53:16.570 | 0.2  |       |       |      | T    |     |      |       |       | e    |           | 40722889     |
| HKT   | 64.55 | 300.9 | PFAKE | 06:53:30.000 | 8.3  |       |       |      | —    |     |      |       |       | —    |           | 40722861     |
| HKT   | 64.55 | 300.9 | LR    | 06:53:30.000 | 8.3  |       |       |      | —    |     |      | 390.0 | 22.00 | —    | MS        | 4.5 40722862 |
| BOSA  | 64.67 | 124.6 | P     | 06:53:21.600 | -0.8 |       |       |      | T    |     |      | 12.2  | 0.90  | e    | mb        | 5.1 40722791 |
| BOSA  | 64.67 | 124.6 | LR    | 06:53:21.600 | -0.8 |       |       |      | —    |     |      | 540.0 | 22.00 | —    | MS        | 4.7 40722792 |



| Sta   | Dist  | EVaz  | Phase | Time         | TRes | Azim  | AzRes | Slow | SRes | Def | SNR  | Amp   | Per   | Qual | Magnitude | ArrID        |
|-------|-------|-------|-------|--------------|------|-------|-------|------|------|-----|------|-------|-------|------|-----------|--------------|
| NOA   | 65.11 | 21.8  | P     | 06:53:24.150 | -0.6 | 230.9 |       | 6.3  | T    |     | 6.4  | 5.4   | 1.01  | —    | mb        | 4.7 42115745 |
| NOA   | 65.11 | 21.8  | SP    | 06:53:30.243 | 1.2  | 224.9 |       | 7.1  |      |     | 6.3  | 6.0   | 1.16  | —    |           | 42115746     |
| NOA   | 65.11 | 21.8  | LR    | 07:20:09.114 |      | 235.0 |       | 34.5 |      |     |      | 164.1 | 18.14 | —    | MS        | 4.3 42115747 |
| CBKS  | 69.33 | 309.7 | PFAKE | 06:54:00.000 | 8.3  |       |       |      |      |     |      |       |       | —    |           | 40722800     |
| CBKS  | 69.33 | 309.7 | LR    |              |      |       |       |      |      |     |      | 540.0 | 22.00 | —    | MS        | 4.8 40722801 |
| ULM   | 69.53 | 322.3 | LR    | 07:20:46.063 |      | 152.0 |       | 32.9 |      |     |      | 170.6 | 18.34 | —    |           | 42115753     |
| LTX   | 71.23 | 299.3 | PFAKE | 06:54:10.000 | 6.5  |       |       |      |      |     |      |       |       | —    |           | 40722904     |
| LTX   | 71.23 | 299.3 | LR    |              |      |       |       |      |      |     |      | 290.0 | 21.00 | —    | MS        | 4.5 40722905 |
| FINES | 71.33 | 25.7  | P     | 06:54:03.775 | 0.3  | 284.5 |       | 7.4  | T    |     | 7.5  | 3.1   | 0.81  | —    | mb        | 4.5 42115737 |
| FINES | 71.33 | 25.7  | SP    | 06:54:10.200 | 2.5  | 260.8 |       | 6.8  |      |     | 7.4  | 15.2  | 1.27  | —    |           | 42115738     |
| GDL2  | 72.11 | 302.2 | P     | 06:54:08.200 | -0.4 |       |       |      | T    |     |      |       |       | e    |           | 40722840     |
| GDL   | 73.68 | 309.9 | PFAKE | 06:54:30.000 | 12.2 |       |       |      |      |     |      |       |       | —    |           | 40722845     |
| GDL   | 73.68 | 309.9 | LR    |              |      |       |       |      |      |     |      | 420.0 | 20.00 | —    | MS        | 4.7 40722846 |
| ANMO  | 74.12 | 304.9 | P     | 06:54:20.520 | 0.1  |       |       |      | T    |     |      | 24.8  | 1.60  | e    | mb        | 5.0 40722768 |
| ANMO  | 74.12 | 304.9 | LR    |              |      |       |       |      |      |     |      | 760.0 | 19.00 | —    | MS        | 5.0 40722769 |
| LPM   | 74.12 | 304.2 | P     | 06:54:21.110 | 0.2  |       |       |      | T    |     |      |       |       | e    |           | 40722900     |
| OBN   | 74.36 | 33.8  | P     | 06:54:21.500 | 0.1  |       |       |      | T    |     |      |       |       | —    |           | 40722928     |
| LAZ   | 74.63 | 304.3 | P     | 06:54:23.740 | 0.4  |       |       |      | T    |     |      |       |       | e    |           | 40722887     |
| RW3   | 75.42 | 308.2 | P     | 06:54:28.440 | 0.6  |       |       |      | T    |     |      |       |       | —    |           | 40722951     |
| KEV   | 75.43 | 18.1  | P     | 06:54:26.200 | -1.1 |       |       |      | T    |     |      |       |       | e    |           | 45302922     |
| KIV   | 76.17 | 46.0  | P     | 06:54:33.060 | 1.1  |       |       |      | T    |     |      | 192.0 | 1.50  | e    | mb        | 6.0 40722877 |
| KVAR  | 76.18 | 46.0  | P     | 06:54:33.975 | 2.0  | 135.0 |       | 2.4  | T    |     | 12.3 | 4.2   | 0.77  | —    | mb        | 4.6 42115741 |
| KVAR  | 76.18 | 46.0  | SP    | 06:54:40.125 | 3.6  | 148.0 |       | 4.5  |      |     | 6.2  | 34.7  | 1.27  | —    |           | 42115742     |
| PV10  | 76.49 | 308.2 | P     | 06:54:34.300 | 0.4  |       |       |      | T    |     |      |       |       | e    |           | 40722945     |
| GNI   | 77.29 | 50.1  | P     | 06:54:40.120 | 1.8  |       |       |      | T    |     |      | 45.8  | 1.60  | e    | mb        | 5.4 40722847 |
| GNI   | 77.29 | 50.1  | LR    |              |      |       |       |      |      |     |      | 160.0 | 21.00 | —    | MS        | 4.3 40722848 |
| TUC   | 77.54 | 301.9 | PFAKE | 06:54:50.000 | 10.1 |       |       |      |      |     |      |       |       | —    |           | 40722981     |
| TUC   | 77.54 | 301.9 | LR    |              |      |       |       |      |      |     |      | 250.0 | 20.00 | —    | MS        | 4.5 40722982 |
| WUAZ  | 78.17 | 305.2 | P     | 06:54:44.110 | 0.9  |       |       |      | T    |     |      | 11.4  | 0.90  | e    | mb        | 5.0 40722995 |
| WUAZ  | 78.17 | 305.2 | LR    |              |      |       |       |      |      |     |      | 520.0 | 22.00 | —    | MS        | 4.8 40722996 |
| STEW  | 78.22 | 313.9 | P     | 06:54:43.330 | 0.0  |       |       |      | T    |     |      |       |       | e    |           | 40722966     |
| DAU   | 78.36 | 310.1 | P     | 06:54:44.780 | 0.6  |       |       |      | T    |     |      |       |       | e    |           | 40722818     |
| SNAA  | 78.64 | 170.8 | P     | 06:54:44.900 | -0.2 |       |       |      | T    |     |      |       |       | de   |           | 42017674     |
| HWUT  | 78.67 | 311.3 | P     | 06:54:45.500 | -0.3 |       |       |      | T    |     |      | 15.1  | 0.70  | e    | mb        | 5.1 40722868 |
| HWUT  | 78.67 | 311.3 | LR    |              |      |       |       |      |      |     |      | 310.0 | 22.00 | —    | MS        | 4.6 40722869 |
| QLMT  | 78.80 | 314.6 | P     | 06:54:47.350 | 0.8  |       |       |      | T    |     |      |       |       | e    |           | 40722947     |
| MSU   | 78.96 | 308.2 | P     | 06:54:48.190 | 0.7  |       |       |      | T    |     |      |       |       | e    |           | 40722917     |
| TMI   | 79.04 | 313.1 | P     | 06:54:47.830 | -0.0 |       |       |      | T    |     |      |       |       | —    |           | 40722975     |
| HRV   | 79.23 | 316.5 | P     | 06:54:48.520 | -0.3 |       |       |      | T    |     |      |       |       | e    |           | 40722867     |
| KNB   | 79.41 | 306.6 | P     | 06:54:50.810 | 0.8  |       |       |      | T    |     |      | 2.5   | 0.80  | e    | mb        | 4.3 40722878 |
| KNB   | 79.41 | 306.6 | LR    |              |      |       |       |      |      |     |      | 500.0 | 21.00 | —    | MS        | 4.8 40722879 |
| DUG   | 79.53 | 309.9 | P     | 06:54:50.910 | 0.3  |       |       |      | T    |     |      | 29.4  | 1.40  | e    | mb        | 5.0 40722823 |
| DUG   | 79.53 | 309.9 | LR    |              |      |       |       |      |      |     |      | 320.0 | 20.00 | —    | MS        | 4.7 40722824 |
| LRM   | 79.60 | 315.6 | P     | 06:54:51.400 | 0.6  |       |       |      | T    |     |      |       |       | e    |           | 40722901     |
| MCMT  | 79.81 | 314.6 | P     | 06:54:52.390 | 0.4  |       |       |      | T    |     |      |       |       | —    |           | 40722909     |
| HLID  | 80.87 | 313.2 | P     | 06:54:57.900 | 0.3  |       |       |      | T    |     |      | 12.3  | 1.30  | e    | mb        | 4.8 40722863 |
| HLID  | 80.87 | 313.2 | LR    |              |      |       |       |      |      |     |      | 420.0 | 19.00 | —    | MS        | 4.8 40722864 |
| ELK   | 81.40 | 310.3 | P     | 06:55:00.560 | 0.1  |       |       |      | T    |     |      | 21.2  | 1.50  | e    | mb        | 5.0 40722828 |

| Sta   | Dist   | EVaz  | Phase | Time         | TRes | Azim  | AzRes | Slow | SRes | Def | SNR | Amp   | Per   | Qual | Magnitude | ArrID        |
|-------|--------|-------|-------|--------------|------|-------|-------|------|------|-----|-----|-------|-------|------|-----------|--------------|
| YKA   | 82.18  | 332.3 | P     | 06:55:08.200 | 4.1  | 94.0  |       | 5.3  |      | T   |     | 2.4   | 1.00  | _e   | mb        | 4.3 40723005 |
| PFO   | 82.33  | 303.1 | P     | 06:55:08.615 | 3.2  | 308.7 |       | 2.3  |      | T   | 4.2 | 2.0   | 0.97  | —    | mb        | 4.2 42115748 |
| NEW   | 82.89  | 317.9 | P     | 06:55:08.560 | 0.5  |       |       |      |      | T   |     |       |       | —    |           | 40722920     |
| NEW   | 82.89  | 317.9 | PcP   | 06:55:20.000 | 6.5  |       |       |      |      |     |     |       |       | —    |           | 40722921     |
| NEW   | 82.89  | 317.9 | LR    | 06:55:20.000 | 11.7 |       |       |      |      |     |     | 380.0 | 21.00 | —    | MS        | 4.7 40722922 |
| BNW   | 82.91  | 310.0 | PFAKE | 06:55:20.000 | 11.7 |       |       |      |      |     |     |       |       | —    |           | 40722788     |
| BNW   | 82.91  | 310.0 | LR    | 06:55:20.000 | 11.7 |       |       |      |      |     |     | 240.0 | 20.00 | —    | MS        | 4.6 40722789 |
| TPH   | 82.92  | 307.6 | P     | 06:55:09.030 | 0.6  |       |       |      |      | T   |     | 17.6  | 1.10  | _e   | mb        | 5.2 40722979 |
| TPH   | 82.92  | 307.6 | LR    | 06:55:20.000 | 10.8 |       |       |      |      |     |     | 320.0 | 22.00 | —    | MS        | 4.7 40722980 |
| VTV   | 83.05  | 304.1 | PFAKE | 06:55:20.000 | 10.8 |       |       |      |      |     |     | 560.0 | 20.00 | —    | MS        | 4.9 40722991 |
| VTV   | 83.05  | 304.1 | LR    | 06:55:20.000 | 10.8 |       |       |      |      |     |     |       |       | —    |           | 40722816     |
| DAC   | 83.23  | 305.8 | PFAKE | 06:55:20.000 | 9.9  |       |       |      |      |     |     | 320.0 | 22.00 | —    | MS        | 4.7 40722817 |
| DAC   | 83.23  | 305.8 | LR    | 06:55:20.000 | 9.9  |       |       |      |      |     |     |       |       | —    |           | 40722914     |
| MNV   | 83.65  | 308.0 | PFAKE | 06:55:20.000 | 7.8  |       |       |      |      |     |     | 300.0 | 20.00 | —    | MS        | 4.7 40722915 |
| MNV   | 83.65  | 308.0 | LR    | 06:55:20.000 | 7.8  |       |       |      |      |     |     |       |       | —    |           | 40722935     |
| PAS   | 83.75  | 303.7 | PFAKE | 06:55:20.000 | 7.2  |       |       |      |      |     |     | 530.0 | 19.00 | —    | MS        | 4.9 40722936 |
| PAS   | 83.75  | 303.7 | LR    | 06:55:20.000 | 7.2  |       |       |      |      |     |     | 30.6  | 1.80  | _e   | mb        | 5.2 40722997 |
| WVOR  | 83.96  | 312.0 | P     | 06:55:14.520 | 0.9  |       |       |      |      | T   |     | 310.0 | 19.00 | —    | MS        | 4.7 40722998 |
| WVOR  | 83.96  | 312.0 | LR    | 06:55:20.000 | 6.2  |       |       |      |      |     |     | 110.0 | 21.00 | —    | MS        | 4.2 40722871 |
| ISA   | 83.96  | 305.2 | PFAKE | 06:55:20.000 | 13.5 |       |       |      |      |     |     | 220.0 | 20.00 | —    | MS        | 4.5 40722857 |
| ISA   | 83.96  | 305.2 | LR    | 06:55:20.000 | 13.5 |       |       |      |      |     |     |       |       | —    |           | 40722856     |
| HAWA  | 84.53  | 316.0 | PFAKE | 06:55:30.000 | 0.5  |       |       |      |      | T   |     |       |       | _e   |           | 40722782     |
| HAWA  | 84.53  | 316.0 | LR    | 06:55:30.000 | 0.5  |       |       |      |      | T   |     |       |       | _e   |           | 40722987     |
| BEKR  | 85.32  | 309.5 | P     | 06:55:21.030 | 0.5  |       |       |      |      |     |     | 380.0 | 20.00 | —    | MS        | 4.8 40722809 |
| VIPTM | 85.35  | 314.2 | P     | 06:55:22.180 | 1.6  |       |       |      |      |     |     | 65.5  | 1.30  | _e   | mb        | 5.7 40722893 |
| CMB   | 85.42  | 307.7 | PFAKE | 06:55:30.000 | 8.9  |       |       |      |      |     |     |       |       | —    |           | 40723003     |
| CMB   | 85.42  | 307.7 | LR    | 06:55:30.000 | 8.9  |       |       |      |      |     |     | 290.0 | 22.00 | —    | MS        | 4.6 40723004 |
| LN    | 86.09  | 316.4 | P     | 06:55:24.480 | 0.2  |       |       |      |      | T   |     |       |       | _e   |           | 40722790     |
| ARU   | 86.78  | 33.8  | P     | 06:55:28.450 | 1.1  |       |       |      |      |     |     | 90.0  | 20.00 | —    | MS        | 4.2 40722812 |
| YBH   | 87.01  | 311.5 | PFAKE | 06:55:40.000 | 11.1 |       |       |      |      | T   |     |       |       | —    |           | 40722813     |
| YBH   | 87.01  | 311.5 | LR    | 06:55:40.000 | 11.1 |       |       |      |      |     |     | 470.0 | 19.00 | —    | MS        | 4.9 40722929 |
| BMW   | 87.08  | 316.2 | P     | 06:55:29.710 | 0.6  |       |       |      |      |     |     |       |       | —    |           | 40722930     |
| COR   | 87.26  | 314.3 | PFAKE | 06:55:40.000 | 10.0 |       |       |      |      | T   |     |       |       | de   |           | 35484913     |
| COR   | 87.26  | 314.3 | LR    | 06:55:40.000 | 10.0 |       |       |      |      | T   |     |       |       | _e   |           | 41194522     |
| OCWA  | 87.64  | 317.5 | PFAKE | 06:55:40.000 | 8.3  |       |       |      |      |     |     | 230.0 | 20.00 | —    | MS        | 4.7 40722810 |
| OCWA  | 87.64  | 317.5 | LR    | 06:55:40.000 | 8.3  |       |       |      |      |     |     | 900.0 | 20.00 | —    | MS        | 5.3 40722954 |
| SYO   | 88.03  | 159.9 | P     | 06:55:34.400 | 1.3  |       |       |      |      |     |     |       |       | —    |           | 40722955     |
| MAIO  | 88.92  | 53.8  | P     | 06:55:38.000 | -0.2 |       |       |      |      | T   |     |       |       | —    |           | 40723001     |
| COLA  | 96.22  | 337.1 | PFAKE | 06:56:20.000 | 8.9  |       |       |      |      | T   |     |       |       | —    |           | 40723002     |
| COLA  | 96.22  | 337.1 | LR    | 06:56:20.000 | 8.9  |       |       |      |      |     |     | 90.0  | 19.00 | —    | MS        | 4.4 40722906 |
| SBA   | 106.16 | 184.2 | PP    | 07:01:20.000 | -1.9 |       |       |      |      |     |     |       |       | —    |           | 4.4 40722907 |
| SBA   | 106.16 | 184.2 | LR    | 07:01:20.000 | -1.9 |       |       |      |      |     |     | 130.0 | 21.00 | —    | MS        | 4.5 40722984 |
| YAK   | 112.11 | 8.9   | PFAKE | 07:01:30.000 | 12.8 |       |       |      |      |     |     |       |       | —    |           | 40722985     |
| YAK   | 112.11 | 8.9   | LR    | 07:01:30.000 | 12.8 |       |       |      |      |     |     |       |       | —    |           |              |
| MA2   | 115.89 | 358.1 | PFAKE | 07:01:40.000 | 15.3 |       |       |      |      |     |     |       |       | —    |           |              |
| MA2   | 115.89 | 358.1 | LR    | 07:01:40.000 | 15.3 |       |       |      |      |     |     |       |       | —    |           |              |
| ULN   | 116.84 | 29.3  | PFAKE | 07:01:40.000 | 13.2 |       |       |      |      |     |     |       |       | —    |           |              |
| ULN   | 116.84 | 29.3  | LR    | 07:01:40.000 | 13.2 |       |       |      |      |     |     |       |       | —    |           |              |

| Sta  | Dist   | EVaz  | Phase | Time         | TRes | Azim  | AzRes | SRes | Def | SNR  | Amp    | Per   | Qual | Magnitude | ArrID        |
|------|--------|-------|-------|--------------|------|-------|-------|------|-----|------|--------|-------|------|-----------|--------------|
| SMY  | 118.51 | 341.9 | LR    |              |      |       |       |      |     |      | 1090.0 | 21.30 | —    | MS        | 5.5 40722963 |
| HON  | 120.61 | 298.0 | PFAKE | 07:01:50.000 | 15.2 |       |       |      |     |      |        |       |      |           | 40722865     |
| HON  | 120.61 | 298.0 | LR    |              |      |       |       |      |     |      | 1230.0 | 21.30 | —    | MS        | 5.5 40722866 |
| HIA  | 121.13 | 20.8  | PFAKE | 07:01:50.000 | 14.9 |       |       |      |     |      |        |       |      |           | 40722858     |
| HIA  | 121.13 | 20.8  | LR    |              |      |       |       |      |     |      | 100.0  | 22.00 | —    | MS        | 4.4 40722859 |
| PET  | 121.74 | 352.0 | PFAKE | 07:01:50.000 | 13.8 |       |       |      |     |      |        |       |      |           | 40722937     |
| PET  | 121.74 | 352.0 | LR    |              |      |       |       |      |     |      | 100.0  | 22.00 | —    | MS        | 4.4 40722938 |
| BJT  | 127.05 | 29.9  | PFAKE | 07:02:00.000 | 13.2 |       |       |      |     |      |        |       |      |           | 40722786     |
| BJT  | 127.05 | 29.9  | LR    |              |      |       |       |      |     |      | 150.0  | 19.00 | —    | MS        | 4.7 40722787 |
| TAT0 | 141.22 | 39.3  | PFAKE | 07:02:30.000 | 16.4 |       |       |      |     |      |        |       |      |           | 40722970     |
| TAT0 | 141.22 | 39.3  | LR    |              |      |       |       |      |     |      | 130.0  | 21.00 | —    | MS        | 4.7 40722971 |
| T00  | 147.22 | 177.2 | PKP   | 07:02:24.100 | 0.6  |       |       |      |     |      | 8.2    | 1.20  | e    |           | 40722977     |
| STKA | 152.40 | 169.3 | PKP   | 07:02:37.500 | 5.8  |       |       |      |     |      | 7.5    | 1.00  | e    |           | 40722967     |
| ASPA | 157.06 | 146.7 | PKP   | 07:02:39.700 | 1.4  |       |       |      |     |      | 8.7    | 1.20  | e    |           | 40722774     |
| ASAR | 157.06 | 146.7 | PKP   | 07:02:41.200 | 2.9  | 203.1 |       |      |     | 12.8 | 2.5    | 0.98  | —    |           | 42115729     |
| ASAR | 157.06 | 146.7 | PKP2  | 07:03:09.850 | -1.3 | 200.0 |       |      |     | 7.7  | 1.5    | 0.78  | —    |           | 42115730     |
| ASAR | 157.06 | 146.7 | SPKP2 | 07:03:16.950 | 1.5  | 209.2 |       |      |     | 4.5  | 3.1    | 1.12  | —    |           | 42115731     |

STOP

| Title   | Example of station parameter reports grouped according<br>IMS1.0 with ISF1.0 extensions                                                                     |
|---------|-------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Raymond J. Willemann</b> , IRIS, 1200 New York Ave, NW, Suite 800,<br>Washington DC 20005, U.S.A; E-mail: <a href="mailto:ray@iris.edu">ray@iris.edu</a> |
| Version | September 2001; DOI: <b>10.2312/GFZ.NMSOP-2_IS_10.2</b>                                                                                                     |

## 1 Introduction

On the following pages a sample is shown of parameter readings of seismic stations from unassociated arrivals as they were grouped by the National Earthquake Information Center (NEIC) of the USGS in Denver, USA, and reported to the International Data Center (ISC) in Thatcham, UK.

ASCII characters are used as part of the Group and Arrival IDs. According to the IMS1.0 documentation about their data-type "Grouped Arrivals" these characters should help the agency to recognize that several phase arrivals come from the same event but that the reporting station can not locate the event, so they are assigned to the same "group". Note that NEIC uses the groups more restrictively: they only let arrivals share the same group ID if they are at the same station and likely come from the same event. This is within the rules. Examples are, e.g., on page 2 for stations VRAC, TXAR and SLR.

The simple example given below fails to show an important ISF extension, namely the phase information sub-block which allows to give, e.g., for every phase reading, information about what filter was used to do the reading, and so on. This kind of information can now be put, in accordance with the ISF format (see IS10.1) into a "sub-block" that would follow the one shown below and share the same values of arrival ID (column heading ArrID).

The abbreviations used for the different columns are more or less self-explanatory. Column Sta contains the three or four letter station codes and in column Chan the components may be given, in which the parameter reading (time, amplitude, period etc.) was made ( Z – vertical, H – horizontal; N – north and E – east). The column for characterizing the arrival quality (Qual) gives either e (for emergent) or i (for impulsive) onsets and may additionally give the polarity of the first motion (c for compression and d for dilatation). Note that (regrettably) in most cases no channel information is provided by the stations and also amplitudes (Amp) and period readings (Per) are rarely given. Azimuth (Azim) and slowness (Slow) readings are usually reported from seismic arrays only (see, e.g., YKA – Yellowknife Array on page 2). In the column SNR measurements of the signal-to-noise ratio could be reported, which, however, would make sense only with accompanying information in the phase information sub-block about the filter bandwidth. Note, that in some cases the column Author (reporting station or network/array) may also be blank (see page 6).

| Net | Sta  | Chan | Aux | Date       | Time        | Phase | Azim | Slow | SNR | Amp   | Per  | Qual | Group    | C | Author | ArrID   |
|-----|------|------|-----|------------|-------------|-------|------|------|-----|-------|------|------|----------|---|--------|---------|
|     | VRAC | ???  |     | 2000/09/01 | 00:05:41.30 | Pn    |      |      |     |       |      |      | 00IAAAA  |   | VRAC   | AIAAAAA |
|     | VRAC | ???  |     | 2000/09/01 | 00:06:04.30 | Sg    |      |      |     |       |      |      | 00IAAAA  |   | VRAC   | AIAAAAA |
|     | CGP  | ???  |     | 2000/09/01 | 00:09:52.50 | P     |      |      |     |       |      | e    | 00IAAAB  |   | PIVS   | AIAAAAA |
|     | ITM  | ???  |     | 2000/09/01 | 00:20:22.50 | Pb    |      |      |     |       |      | e    | 00IAAAC  |   | ATH    | AIAAAAA |
|     | VLI  | ???  |     | 2000/09/01 | 00:20:41.00 | Pg    |      |      |     |       |      | e    | 00IAAAC  |   | ATH    | AIAAAAA |
|     | EVR  | ???  |     | 2000/09/01 | 00:20:43.00 | Pb    |      |      |     |       |      | e    | 00IAAAC  |   | ATH    | AIAAAAA |
|     | BKM  | ???  |     | 2000/09/01 | 00:21:28.50 | P     |      |      |     |       |      | i    | 00IAAAD  |   | NOU    | AIAAAAA |
|     | NDI  | ???  |     | 2000/09/01 | 00:36:17.00 | P     |      |      |     |       |      | e    | 00IAAAE  |   | NDI    | AIAAAAA |
|     | EZN  | ???  |     | 2000/09/01 | 00:56:01.00 | Pg    |      |      |     |       |      | e    | 00IAAAF  |   | ISK    | AIAAAAA |
|     | EZN  | ???  |     | 2000/09/01 | 00:56:07.50 | Sg    |      |      |     |       |      | e    | 00IAAAF  |   | ISK    | AIAAAAA |
|     | PGP  | ???  |     | 2000/09/01 | 01:00:09.80 | P     |      |      |     |       |      | e    | 00IAAAG  |   | PIVS   | AIAAAAA |
|     | CWAR | ???  |     | 2000/09/01 | 01:03:39.90 | P     |      |      |     |       |      |      | 00IAAAH  |   | IDC    | AIAAAAA |
|     | FITZ | ???  |     | 2000/09/01 | 01:04:04.80 | P     |      |      |     | 9.1   | 0.50 | c    | 00IAAAI  |   | AUST   | AIAAAAA |
|     | ASAR | ???  |     | 2000/09/01 | 01:05:27.30 | P     |      |      |     |       |      |      | 00IAAAJ  |   | IDC    | AIAAAAA |
|     | ASAR | ???  |     | 2000/09/01 | 01:05:37.30 | Pp    |      |      |     |       |      |      | 00IAAAJ  |   | IDC    | AIAAAAA |
|     | ASPA | ???  |     | 2000/09/01 | 01:05:27.30 | P     |      |      |     |       |      | e    | 00IAAAK  |   | AUST   | AIAAAAA |
|     | STKA | ???  |     | 2000/09/01 | 01:06:47.35 | P     |      |      |     |       |      |      | 00IAAAL  |   | IDC    | AIAAAAA |
|     | STKA | ???  |     | 2000/09/01 | 01:06:57.10 | Pp    |      |      |     |       |      |      | 00IAAAL  |   | IDC    | AIAAAAA |
|     | STKA | ???  |     | 2000/09/01 | 01:25:44.13 | Lr    |      |      |     |       |      |      | 00IAAAL  |   | IDC    | AIAAAAA |
|     | MAT  | ???  |     | 2000/09/01 | 01:16:52.00 | P     |      |      |     |       |      | e    | 00IAAAM  |   | JMA    | AIAAAAA |
|     | MAT  | ???  |     | 2000/09/01 | 01:17:26.00 | S     |      |      |     |       |      | e    | 00IAAAM  |   | JMA    | AIAAAAA |
|     | GJRN | ???  |     | 2000/09/01 | 01:17:32.36 | P     |      |      |     | 12.4  | 0.45 |      | 00IAAAN  |   | DMN    | AIAAAAA |
|     | GHAN | ???  |     | 2000/09/01 | 01:17:34.02 | P     |      |      |     |       |      |      | 00IAAAN  |   | DMN    | AIAAAAA |
|     | HARN | ???  |     | 2000/09/01 | 01:17:48.04 | P     |      |      |     |       |      |      | 00IAAAN  |   | DMN    | AIAAAAA |
|     | TXAR | ???  |     | 2000/09/01 | 01:18:06.75 | PKPbc |      |      |     |       |      |      | 00IAAAP  |   | IDC    | AIAAAAA |
|     | TXAR | ???  |     | 2000/09/01 | 01:18:17.63 | Pp'bc |      |      |     |       |      |      | 00IAAAP  |   | IDC    | AIAAAAA |
|     | TXAR | ???  |     | 2000/09/01 | 01:18:23.25 | sp'bc |      |      |     |       |      |      | 00IAAAP  |   | IDC    | AIAAAAA |
|     | MAT  | ???  |     | 2000/09/01 | 01:24:12.00 | P     |      |      |     |       |      | e    | 00IAAAQ  |   | JMA    | AIAAAAA |
|     | MAT  | ???  |     | 2000/09/01 | 01:24:21.00 | S     |      |      |     |       |      | e    | 00IAAAQ  |   | JMA    | AIAAAAA |
|     | NB2  | ???  |     | 2000/09/01 | 01:26:59.50 | P     |      |      |     | 2.1   | 0.80 |      | 00IAAAAR |   | NAO    | AIAAAAA |
|     | SLR  | ???  |     | 2000/09/01 | 01:43:25.90 | P     |      |      |     | 155.0 | 1.00 | e    | 00IAAAAS |   | PRE    | AIAAAAA |
|     | SLR  | ???  |     | 2000/09/01 | 01:43:40.10 | S     |      |      |     |       |      | e    | 00IAAAAS |   | PRE    | AIAAAAA |
|     | KSR  | ???  |     | 2000/09/01 | 01:43:47.90 | P     |      |      |     |       |      | e    | 00IAAAAS |   | PRE    | AIAAAAA |
|     | KSR  | ???  |     | 2000/09/01 | 01:43:54.90 | S     |      |      |     |       |      | e    | 00IAAAAS |   | PRE    | AIAAAAA |
|     | BLF  | ???  |     | 2000/09/01 | 01:44:21.30 | P     |      |      |     |       |      | e    | 00IAAAAS |   | PRE    | AIAAAAA |
|     | BLF  | ???  |     | 2000/09/01 | 01:44:55.90 | S     |      |      |     | 33.0  | 0.70 | e    | 00IAAAAS |   | PRE    | AIAAAAA |
|     | BOSA | ?HZ  |     | 2000/09/01 | 01:44:21.90 | P     |      |      |     |       |      | e    | 00IAAAAT |   | NEIC   | AIAAAAA |
|     | BOSA | ?HZ  |     | 2000/09/01 | 01:44:56.10 | S     |      |      |     |       |      | e    | 00IAAAAT |   | NEIC   | AIAAAAA |
|     | YKA  | ???  |     | 2000/09/01 | 01:50:20.60 | P     | 274. | 1.7  |     |       |      | e    | 00IAAAU  |   | OTT    | AIAAAAA |
|     | KYTH | ???  |     | 2000/09/01 | 01:56:56.50 | Pb    |      |      |     |       |      | e    | 00IAAAV  |   | ATH    | AIAAAAA |
|     | ITM  | ???  |     | 2000/09/01 | 01:57:02.70 | Pb    |      |      |     |       |      | e    | 00IAAAV  |   | ATH    | AIAAAAA |

| Net | Sta   | Chan | Aux | Date       | Time        | Phase | Azim | Slow | SNR | Amp  | Per  | Qual | Group    | C | Author | ArrID    |
|-----|-------|------|-----|------------|-------------|-------|------|------|-----|------|------|------|----------|---|--------|----------|
|     | PCI   | ???  |     | 2000/09/01 | 02:00:32.50 | P     |      |      |     |      |      | ce   | 00IAAAW  |   | DJA    | AIAAAA^A |
|     | VNA1  | ???  |     | 2000/09/01 | 02:06:18.90 | P     |      |      |     |      |      | ci   | 00IAAAAX |   | AWI    | AIAAAA A |
|     | VNA3  | ???  |     | 2000/09/01 | 02:06:20.20 | P     |      |      |     |      |      | di   | 00IAAAAX |   | AWI    | AIAAAA^A |
|     | VNA2  | ???  |     | 2000/09/01 | 02:06:24.10 | P     |      |      |     |      |      |      | 00IAAAAX |   | AWI    | AIAAAAaA |
|     | PLP   | ???  |     | 2000/09/01 | 02:11:39.50 | P     |      |      |     |      |      | i    | 00IAAAAY |   | PIVS   | AIAAAAaB |
|     | PLP   | ???  |     | 2000/09/01 | 02:11:47.00 | S     |      |      |     |      |      | e    | 00IAAAAY |   | PIVS   | AIAAAAbB |
|     | EZN   | ???  |     | 2000/09/01 | 02:33:28.00 | Pg    |      |      |     |      |      | e    | 00IAAAAZ |   | ISK    | AIAAAAcA |
|     | EZN   | ???  |     | 2000/09/01 | 02:33:35.00 | Sg    |      |      |     |      |      | e    | 00IAAAAZ |   | ISK    | AIAAAAcB |
|     | FINC  | ???  |     | 2000/09/01 | 02:39:19.70 | P     |      |      |     |      |      | e    | 00IAAABA |   | PMG    | AIAAAAdA |
|     | ILAR  | ???  |     | 2000/09/01 | 03:20:59.90 | P     |      |      |     |      |      |      | 00IAAABB |   | IDC    | AIAAAAeA |
|     | INK   | ???  |     | 2000/09/01 | 03:21:38.55 | P     |      |      |     |      |      |      | 00IAAABB |   | IDC    | AIAAAAfA |
|     | NVAR  | ???  |     | 2000/09/01 | 03:24:27.38 | P     |      |      |     |      |      |      | 00IAAABC |   | IDC    | AIAAAAgA |
|     | FINES | ???  |     | 2000/09/01 | 03:24:30.20 | P     |      |      |     |      |      |      | 00IAAABC |   | IDC    | AIAAAAhA |
|     | TXAR  | ???  |     | 2000/09/01 | 03:26:00.03 | P     | 296. | 13.8 |     |      |      |      | 00IAAABD |   | IDC    | AIAAAAiA |
|     | YKA   | ???  |     | 2000/09/01 | 03:48:38.60 | P     |      |      |     |      |      | e    | 00IAAABE |   | OTT    | AIAAAAjA |
|     | VRAC  | ???  |     | 2000/09/01 | 03:56:19.80 | Pg    |      |      |     |      |      |      | 00IAAABF |   | VRAC   | AIAAAAkA |
|     | VRAC  | ???  |     | 2000/09/01 | 03:56:38.20 | Sg    |      |      |     |      |      |      | 00IAAABF |   | VRAC   | AIAAAAkB |
|     | YKA   | ???  |     | 2000/09/01 | 04:29:11.40 | P     | 272. | 6.6  |     | 1.5  | 0.80 |      | 00IAAABG |   | OTT    | AIAAAAlA |
|     | GSPH  | ???  |     | 2000/09/01 | 04:32:34.00 | P     |      |      |     |      |      | ci   | 00IAAABH |   | PIVS   | AIAAAAmA |
|     | BIPH  | ???  |     | 2000/09/01 | 04:32:59.00 | P     |      |      |     |      |      | i    | 00IAAABH |   | PIVS   | AIAAAAnA |
|     | BIPH  | ???  |     | 2000/09/01 | 04:33:27.50 | S     |      |      |     |      |      | i    | 00IAAABH |   | PIVS   | AIAAAAnB |
|     | ASAR  | ???  |     | 2000/09/01 | 04:35:05.25 | P     |      |      |     |      |      |      | 00IAAABI |   | IDC    | AIAAAAoA |
|     | STKA  | ???  |     | 2000/09/01 | 04:36:29.25 | P     |      |      |     |      |      |      | 00IAAABJ |   | IDC    | AIAAAApA |
|     | TXAR  | ???  |     | 2000/09/01 | 04:48:19.77 | PKP   |      |      |     |      |      | di   | 00IAAABL |   | IDC    | AIAAAAqA |
|     | MAIO  | ???  |     | 2000/09/01 | 05:05:29.00 | P     |      |      |     |      |      |      | 00IAAABK |   | IDC    | AIAAAArA |
|     | MAIO  | ???  |     | 2000/09/01 | 05:06:04.00 | S     |      |      |     |      |      |      | 00IAAABL |   | TEH    | AIAAAArB |
|     | KER   | ???  |     | 2000/09/01 | 05:07:04.00 | P     |      |      |     |      |      | e    | 00IAAABM |   | TEH    | AIAAAAsA |
|     | BILL  | ???  |     | 2000/09/01 | 05:08:42.10 | P     |      |      |     |      |      | ci   | 00IAAABN |   | NERS   | AIAAAAtA |
|     | BILL  | ???  |     | 2000/09/01 | 05:11:57.80 | S     |      |      |     |      |      | e    | 00IAAABN |   | NERS   | AIAAAAtB |
|     | YKA   | ???  |     | 2000/09/01 | 05:08:56.10 | P     | 297. | 12.8 |     |      |      |      | 00IAAABP |   | OTT    | AIAAAAuA |
|     | ISP   | ???  |     | 2000/09/01 | 05:09:23.60 | Pn    |      |      |     |      |      | e    | 00IAAABQ |   | ISK    | AIAAAAvA |
|     | YKA   | ???  |     | 2000/09/01 | 05:09:40.10 | P     |      |      |     | 6.6  | 0.80 |      | 00IAAABR |   | OTT    | AIAAAAwA |
|     | ELL   | ???  |     | 2000/09/01 | 05:09:42.00 | P     |      |      |     |      |      | e    | 00IAAABS |   | ISK    | AIAAAAxA |
|     | WMQ   | ???  |     | 2000/09/01 | 05:09:57.00 | P     | 288. | 8.7  |     | 96.0 | 1.50 |      | 00IAAABT |   | BJI    | AIAAAAyA |
|     | MA2   | ???  |     | 2000/09/01 | 05:10:13.40 | P     |      |      |     |      |      | e    | 00IAAABU |   | NERS   | AIAAAAzA |
|     | MA2   | ???  |     | 2000/09/01 | 05:10:43.20 | Pp    |      |      |     |      |      | e    | 00IAAABU |   | NERS   | AIAAAAzB |
|     | MA2   | ???  |     | 2000/09/01 | 05:14:42.50 | S     |      |      |     |      |      | e    | 00IAAABU |   | NERS   | AIAAAAzC |
|     | KAF   | ???  |     | 2000/09/01 | 05:10:46.10 | P     |      |      |     |      |      |      | 00IAAABV |   | HEL    | AIAAAA{A |
|     | YAK   | ???  |     | 2000/09/01 | 05:11:14.40 | P     |      |      |     |      |      |      | 00IAAABW |   | YARS   | AIAAAA A |
|     | MOX   | ???  |     | 2000/09/01 | 05:11:37.40 | P     |      |      |     |      |      | e    | 00IAAABX |   | JEN    | AIAAAA}A |
|     | DAC   | ?HZ  |     | 2000/09/01 | 05:11:40.30 | P     |      |      |     |      |      |      | 00IAAABY |   | NEIC   | AIAAAA~A |

| Net | Sta  | Chan | Aux | Date       | Time        | Phase | Azim | Slow | SNR | Amp    | Per   | Qual | Group    | C | Author | ArrID      |
|-----|------|------|-----|------------|-------------|-------|------|------|-----|--------|-------|------|----------|---|--------|------------|
|     | NB2  | ???  |     | 2000/09/01 | 05:11:52.70 | P     |      |      |     | 2.8    | 0.80  | —    | 00IAAABZ |   | NAO    | AI4AAAB!A  |
|     | YSS  | ???  |     | 2000/09/01 | 05:11:56.70 | P     |      |      |     | 50.0   | 1.00  | _ci  | 00IAAACA |   | OBN    | AI4AAAB"A  |
|     | CLNS | ???  |     | 2000/09/01 | 05:12:00.00 | P     |      |      |     |        |       | —    | 00IAAACB |   | YARS   | AI4AAAB#A  |
|     | CLNS | ???  |     | 2000/09/01 | 05:12:31.00 | SP    |      |      |     |        |       | —    | 00IAAACB |   | YARS   | AI4AAAB#B  |
|     | NRIS | ???  |     | 2000/09/01 | 05:12:15.00 | P     |      |      |     |        |       | —    | 00IAAACB |   | OBN    | AI4AAAB\$A |
|     | KLR  | ???  |     | 2000/09/01 | 05:12:22.00 | P     |      |      |     |        |       | _e   | 00IAAACB |   | OBN    | AI4AAAB% A |
|     | TIY  | ???  |     | 2000/09/01 | 05:12:45.60 | P     |      |      |     |        |       | _e   | 00IAAACD |   | BJI    | AI4AAAB&A  |
|     | TIY  | ??Z  |     |            |             | LR    |      |      |     | 540.0  | 20.00 | —    | 00IAAACD |   | BJI    | AI4AAAB&B  |
|     | TIY  | ??N  |     |            |             | LR    |      |      |     | 470.0  | 15.00 | —    | 00IAAACD |   | BJI    | AI4AAAB&C  |
|     | TRO  | ???  |     | 2000/09/01 | 05:13:00.12 | P     |      |      |     |        |       | _e   | 00IAAAAE |   | BER    | AI4AAAB' A |
|     | KTK1 | ???  |     | 2000/09/01 | 05:13:06.12 | P     |      |      |     |        |       | _e   | 00IAAAAE |   | BER    | AI4AAAB (A |
|     | LVZ  | ???  |     | 2000/09/01 | 05:13:13.70 | P     |      |      |     |        |       | _e   | 00IAAACF |   | OBN    | AI4AAAB) A |
|     | APA  | ???  |     | 2000/09/01 | 05:13:15.90 | P     |      |      |     |        |       | _ci  | 00IAAACF |   | OBN    | AI4AAAB* A |
|     | TLY  | ???  |     | 2000/09/01 | 05:13:33.30 | P     |      |      |     |        |       | _ci  | 00IAAACF |   | OBN    | AI4AAAB+ A |
|     | MOY  | ???  |     | 2000/09/01 | 05:13:41.00 | P     |      |      |     | 32.0   | 1.20  | _e   | 00IAAACF |   | OBN    | AI4AAAB, A |
|     | NVS  | ???  |     | 2000/09/01 | 05:13:56.00 | P     |      |      |     | 54.0   | 0.70  | _i   | 00IAAACG |   | ASRS   | AI4AAAB- A |
|     | NVS  | ???  |     | 2000/09/01 | 05:14:26.00 | P     |      |      |     |        |       | _e   | 00IAAACG |   | ASRS   | AI4AAAB- B |
|     | NB2  | ???  |     | 2000/09/01 | 05:13:58.90 | P     |      |      |     |        |       | —    | 00IAAACG |   | BER    | AI4AAAB. A |
|     | BJI  | ???  |     | 2000/09/01 | 05:14:06.00 | P     |      |      |     | 10.0   | 1.00  | _e   | 00IAAACI |   | BJI    | AI4AAAB/ A |
|     | BJI  | ??Z  |     |            |             | LR    |      |      |     | 590.0  | 18.00 | —    | 00IAAACI |   | BJI    | AI4AAAB/ B |
|     | HHC  | ???  |     | 2000/09/01 | 05:14:17.20 | P     |      |      |     | 30.0   | 1.20  | _e   | 00IAAACI |   | BJI    | AI4AAAB0A  |
|     | HHC  | ??Z  |     |            |             | LR    |      |      |     | 760.0  | 10.00 | —    | 00IAAACI |   | BJI    | AI4AAAB0B  |
|     | HHC  | ??N  |     |            |             | LR    |      |      |     | 900.0  | 18.00 | —    | 00IAAACI |   | BJI    | AI4AAAB0C  |
|     | HHC  | ??E  |     | 2000/09/01 | 05:14:22.60 | P     |      |      |     | 880.0  | 14.00 | —    | 00IAAACI |   | BJI    | AI4AAAB0D  |
|     | ARU  | ???  |     | 2000/09/01 | 05:14:22.60 | P     |      |      |     | 58.0   | 0.70  | _ci  | 00IAAACJ |   | OBN    | AI4AAAB1A  |
|     | OBN  | ???  |     | 2000/09/01 | 05:14:45.10 | P     |      |      |     |        |       | _e   | 00IAAACJ |   | OBN    | AI4AAAB2A  |
|     | OBN  | ???  |     | 2000/09/01 | 05:15:19.40 | PP    |      |      |     |        |       | —    | 00IAAACJ |   | OBN    | AI4AAAB2B  |
|     | NJ2  | ???  |     | 2000/09/01 | 05:14:46.80 | P     |      |      |     | 25.0   | 1.00  | _e   | 00IAAACK |   | BJI    | AI4AAAB3A  |
|     | NJ2  | ??Z  |     |            |             | LR    |      |      |     | 660.0  | 13.00 | —    | 00IAAACK |   | BJI    | AI4AAAB3B  |
|     | GTA  | ???  |     | 2000/09/01 | 05:14:56.50 | P     |      |      |     | 17.0   | 1.00  | —    | 00IAAACK |   | BJI    | AI4AAAB4A  |
|     | GTA  | ??Z  |     |            |             | LR    |      |      |     | 440.0  | 14.00 | —    | 00IAAACK |   | BJI    | AI4AAAB4B  |
|     | GTA  | ??N  |     |            |             | LR    |      |      |     | 710.0  | 16.00 | —    | 00IAAACK |   | BJI    | AI4AAAB4C  |
|     | XAN  | ???  |     | 2000/09/01 | 05:15:03.00 | P     |      |      |     | 9.4    | 0.80  | —    | 00IAAACK |   | BJI    | AI4AAAB5A  |
|     | XAN  | ??Z  |     |            |             | LR    |      |      |     | 610.0  | 20.00 | —    | 00IAAACK |   | BJI    | AI4AAAB5B  |
|     | XAN  | ??E  |     | 2000/09/01 | 05:15:05.50 | P     |      |      |     | 600.0  | 20.00 | —    | 00IAAACK |   | BJI    | AI4AAAB5C  |
|     | LZH  | ???  |     | 2000/09/01 | 05:15:32.50 | PP    |      |      |     | 67.0   | 1.00  | _c   | 00IAAACK |   | BJI    | AI4AAAB6A  |
|     | LZH  | ???  |     | 2000/09/01 | 05:15:32.50 | PP    |      |      |     |        |       | —    | 00IAAACK |   | BJI    | AI4AAAB6B  |
|     | LZH  | ???  |     | 2000/09/01 | 05:15:54.00 | SP    |      |      |     |        |       | —    | 00IAAACK |   | BJI    | AI4AAAB6C  |
|     | LZH  | ??Z  |     |            |             | LR    |      |      |     | 630.0  | 16.00 | —    | 00IAAACK |   | BJI    | AI4AAAB6D  |
|     | LZH  | ??N  |     |            |             | LR    |      |      |     | 3870.0 | 22.00 | —    | 00IAAACK |   | BJI    | AI4AAAB6E  |
|     | WHN  | ???  |     | 2000/09/01 | 05:15:08.00 | P     |      |      |     |        |       | _e   | 00IAAACK |   | BJI    | AI4AAAB7A  |

| Net | Sta   | Chan | Aux | Date       | Time        | Phase | Azim | Slow | SNR | Amp  | Per  | Qual | Group    | C | Author | ArrID    |
|-----|-------|------|-----|------------|-------------|-------|------|------|-----|------|------|------|----------|---|--------|----------|
|     | CD2   | ???  |     | 2000/09/01 | 05:15:33.80 | P     |      |      |     | 19.0 | 1.00 | c    | 00IAAACK |   | BJI    | AI8AAB8A |
|     | GVA   | ???  |     | 2000/09/01 | 05:15:51.60 | P     |      |      |     | 40.0 | 1.00 | d    | 00IAAACK |   | BJI    | AI8AAB9A |
|     | KIV   | ???  |     | 2000/09/01 | 05:15:53.20 | P     |      |      |     | 46.0 | 1.00 | d    | 00IAAACL |   | OBN    | AI8AAB:A |
|     | KIV   | ???  |     | 2000/09/01 | 05:16:05.80 |       |      |      |     |      |      | i    | 00IAAACL |   | OBN    | AI8AAB:B |
|     | KIV   | ???  |     | 2000/09/01 | 05:25:05.50 | S     |      |      |     |      |      | q    | 00IAAACL |   | OBN    | AI8AAB:C |
|     | ZEI   | ???  |     | 2000/09/01 | 05:16:02.00 | P     |      |      |     |      |      | e    | 00IAAACL |   | OBN    | AI8AAB:A |
|     | ZEI   | ???  |     | 2000/09/01 | 05:16:48.00 |       |      |      |     |      |      | e    | 00IAAACL |   | OBN    | AI8AAB:B |
|     | LSA   | ???  |     | 2000/09/01 | 05:16:10.60 | P     |      |      |     |      |      | e    | 00IAAACK |   | BJI    | AI8AAB<A |
|     | YKA   | ???  |     | 2000/09/01 | 05:16:12.80 | P     |      |      |     | 1.3  | 0.70 | e    | 00IAAACK |   | OTT    | AI8AAB=A |
|     | MAIO  | ???  |     | 2000/09/01 | 05:16:24.00 | P     | 237. | 4.8  |     |      |      | e    | 00IAAACP |   | TEH    | AI8AAB>A |
|     | MAIO  | ???  |     | 2000/09/01 | 05:17:44.00 | P     |      |      |     |      |      | e    | 00IAAACK |   | TEH    | AI8AAB?A |
|     | MAIO  | ???  |     | 2000/09/01 | 05:18:21.00 | S     |      |      |     |      |      | e    | 00IAAACK |   | TEH    | AI8AAB?B |
|     | KONO  | ???  |     | 2000/09/01 | 05:30:14.03 | P     |      |      |     |      |      | e    | 00IAAACK |   | BER    | AI8AAB@A |
|     | MAT   | ???  |     | 2000/09/01 | 05:34:07.00 | P     |      |      |     |      |      | e    | 00IAAACK |   | JMA    | AI8AABAA |
|     | MAT   | ???  |     | 2000/09/01 | 05:34:45.00 | S     |      |      |     |      |      | e    | 00IAAACK |   | JMA    | AI8AABAB |
|     | KBS   | ???  |     | 2000/09/01 | 05:35:24.93 | P     |      |      |     |      |      | e    | 00IAAACK |   | BER    | AI8AABBA |
|     | KSAR  | ???  |     | 2000/09/01 | 05:46:16.30 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABCA |
|     | ASAR  | ???  |     | 2000/09/01 | 05:50:01.50 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABDA |
|     | STKA  | ???  |     | 2000/09/01 | 05:50:55.20 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABEA |
|     | FINES | ???  |     | 2000/09/01 | 05:53:55.05 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABFA |
|     | NVAR  | ???  |     | 2000/09/01 | 05:53:55.20 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABGA |
|     | BKM   | ???  |     | 2000/09/01 | 06:00:48.00 | P     |      |      |     |      |      | i    | 00IAAACK |   | NOU    | AI8AABHA |
|     | PLCA  | ???  |     | 2000/09/01 | 06:01:10.06 | PKPbc |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABIA |
|     | LPAZ  | ???  |     | 2000/09/01 | 06:01:21.84 | PKPbc |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABJA |
|     | STKA  | ???  |     | 2000/09/01 | 06:05:54.00 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABKA |
|     | MAIO  | ???  |     | 2000/09/01 | 06:06:05.00 | P     |      |      |     |      |      | e    | 00IAAACK |   | TEH    | AI8AABLA |
|     | MAIO  | ???  |     | 2000/09/01 | 06:06:42.00 | S     |      |      |     |      |      | e    | 00IAAACK |   | TEH    | AI8AABL  |
|     | ASAR  | ???  |     | 2000/09/01 | 06:06:35.45 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABMA |
|     | VNDA  | ???  |     | 2000/09/01 | 06:10:04.74 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABNA |
|     | ILAR  | ???  |     | 2000/09/01 | 06:12:53.90 | P     |      |      |     |      |      |      | 00IAAACK |   | IDC    | AI8AABOA |
|     | MRA   | ???  |     | 2000/09/01 | 06:20:20.50 | P     |      |      |     |      |      | c    | 00IAAACK |   | SJA    | AI8AABPA |
|     | MRA   | ???  |     | 2000/09/01 | 06:20:23.00 | S     |      |      |     |      |      |      | 00IAAACK |   | SJA    | AI8AABPB |
|     | WATA  | ???  |     | 2000/09/01 | 06:20:43.60 | Pg    |      |      |     | 39.6 | 0.10 | c    | 00IAAACK |   | VIE    | AI8AABQB |
|     | WATA  | ???  |     | 2000/09/01 | 06:20:45.70 | Sg    |      |      |     |      |      | i    | 00IAAACK |   | VIE    | AI8AABQA |
|     | WTTA  | ???  |     | 2000/09/01 | 06:20:43.70 | Pg    |      |      |     | 18.2 | 0.20 | c    | 00IAAACK |   | VIE    | AI8AABRA |
|     | WTTA  | ???  |     | 2000/09/01 | 06:20:46.00 | Sg    |      |      |     |      |      | i    | 00IAAACK |   | VIE    | AI8AABRB |
|     | SQTA  | ???  |     | 2000/09/01 | 06:20:45.10 | Pg    |      |      |     | 5.4  | 0.10 | c    | 00IAAACK |   | VIE    | AI8AABSA |
|     | SQTA  | ???  |     | 2000/09/01 | 06:20:48.60 | Sg    |      |      |     |      |      | i    | 00IAAACK |   | VIE    | AI8AABSB |
|     | MOTA  | ???  |     | 2000/09/01 | 06:20:46.30 | Pg    |      |      |     | 6.0  | 0.20 | d    | 00IAAACK |   | VIE    | AI8AABTB |
|     | MOTA  | ???  |     | 2000/09/01 | 06:20:50.70 | Sg    |      |      |     |      |      | i    | 00IAAACK |   | VIE    | AI8AABTB |
|     | DPC   | ???  |     | 2000/09/01 | 06:23:55.30 | Pg    |      |      |     |      |      | e    | 00IAAACK |   | PRU    | AI8AABUG |



| Net | Sta  | Chan | Aux | Date       | Time        | Phase | Azim | Slow | SNR | Amp | Per  | Qual | Group    | C | Author | ArrID    |
|-----|------|------|-----|------------|-------------|-------|------|------|-----|-----|------|------|----------|---|--------|----------|
|     | DPC  | ???  |     | 2000/09/01 | 06:24:10.60 | Sg    |      |      |     |     |      | e    | 00IAAADG |   | PRU    | AI000000 |
|     | BRG  | ???  |     | 2000/09/01 | 06:23:59.90 | Pg    |      |      |     |     |      | i    | 00IAAADH |   |        | AI000001 |
|     | BRG  | ???  |     | 2000/09/01 | 06:24:19.30 | Sg    |      |      |     |     |      | i    | 00IAAADH |   |        | AI000002 |
|     | IZI  | ???  |     | 2000/09/01 | 06:52:39.00 | Pn    |      |      |     |     |      | e    | 00IAAADJ |   | ISK    | AI000003 |
|     | BAG  | ???  |     | 2000/09/01 | 06:54:02.00 | P     |      |      |     |     |      | i    | 00IAAADJ |   | QCP    | AI000004 |
|     | BCPH | ???  |     | 2000/09/01 | 06:54:06.60 | P     |      |      |     |     |      | ci   | 00IAAADK |   | PIVS   | AI000005 |
|     | RIY  | ???  |     | 2000/09/01 | 06:57:23.40 | Pg    |      |      |     |     |      | i    | 00IAAADL |   | ZAG    | AI000006 |
|     | RIY  | ???  |     | 2000/09/01 | 06:57:27.60 | Sg    |      |      |     |     |      | i    | 00IAAADL |   | ZAG    | AI000007 |
|     | ASAR | ???  |     | 2000/09/01 | 06:58:23.80 | P     |      |      |     |     |      |      | 00IAAADM |   | IDC    | AI000008 |
|     | PHNC | ???  |     | 2000/09/01 | 06:59:58.10 | P     |      |      |     |     |      |      | 00IAAADN |   | NIC    | AI000009 |
|     | PHNC | ???  |     | 2000/09/01 | 07:00:05.20 | S     |      |      |     |     |      |      | 00IAAADN |   | NIC    | AI000010 |
|     | CSS  | ???  |     | 2000/09/01 | 07:00:01.00 | P     |      |      |     | 2.7 | 0.24 |      | 00IAAADN |   | NIC    | AI000011 |
|     | CSS  | ???  |     | 2000/09/01 | 07:00:09.50 | S     |      |      |     |     |      |      | 00IAAADN |   | NIC    | AI000012 |
|     | ALFC | ???  |     | 2000/09/01 | 07:00:09.80 | P     |      |      |     |     |      |      | 00IAAADN |   | NIC    | AI000013 |
|     | ALFC | ???  |     | 2000/09/01 | 07:00:25.80 | S     |      |      |     |     |      |      | 00IAAADN |   | NIC    | AI000014 |
|     | AKMC | ???  |     | 2000/09/01 | 07:00:12.90 | P     |      |      |     |     |      |      | 00IAAADN |   | NIC    | AI000015 |
|     | AKMC | ???  |     | 2000/09/01 | 07:00:30.30 | S     |      |      |     |     |      |      | 00IAAADN |   | NIC    | AI000016 |
|     | NVAR | ???  |     | 2000/09/01 | 07:01:32.28 | P     |      |      |     |     |      |      | 00IAAADP |   | IDC    | AI000017 |
|     | TXAR | ???  |     | 2000/09/01 | 07:02:01.23 | P     |      |      |     |     |      |      | 00IAAADP |   | IDC    | AI000018 |
|     | ILAR | ???  |     | 2000/09/01 | 07:02:18.50 | P     |      |      |     |     |      |      | 00IAAADP |   | IDC    | AI000019 |
|     | YKA  | ???  |     | 2000/09/01 | 07:02:54.60 | P     | 29.  | 12.7 |     |     |      | e    | 00IAAADQ |   | OTT    | AI000020 |
|     | YLV  | ???  |     | 2000/09/01 | 07:23:09.20 | Pg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000021 |
|     | YLV  | ???  |     | 2000/09/01 | 07:23:12.30 | Sg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000022 |
|     | IZI  | ???  |     | 2000/09/01 | 07:23:13.00 | Pg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000023 |
|     | ISK  | ???  |     | 2000/09/01 | 07:23:13.70 | Pg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000024 |
|     | HRT  | ???  |     | 2000/09/01 | 07:23:14.00 | Pg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000025 |
|     | KCT  | ???  |     | 2000/09/01 | 07:23:18.60 | Pg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000026 |
|     | EYL  | ???  |     | 2000/09/01 | 07:23:19.50 | Pg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000027 |
|     | EYL  | ???  |     | 2000/09/01 | 07:23:31.00 | Sg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000028 |
|     | EDC  | ???  |     | 2000/09/01 | 07:23:23.00 | Pg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000029 |
|     | DST  | ???  |     | 2000/09/01 | 07:23:26.00 | Pg    |      |      |     |     |      | i    | 00IAAADR |   | ISK    | AI000030 |
|     | DST  | ???  |     | 2000/09/01 | 07:23:41.00 | Sg    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000031 |
|     | MDU  | ???  |     | 2000/09/01 | 07:23:34.00 | Pn    |      |      |     |     |      | e    | 00IAAADR |   | ISK    | AI000032 |
|     | FIA0 | ???  |     | 2000/09/01 | 07:28:41.35 | Pn    |      |      |     |     |      |      | 00IAAADS |   | HEL    | AI000033 |
|     | FIA0 | ???  |     | 2000/09/01 | 07:29:09.70 | Lg    |      |      |     |     |      |      | 00IAAADS |   | HEL    | AI000034 |
|     | HFS  | ???  |     | 2000/09/01 | 07:29:05.45 | Pn    |      |      |     |     |      |      | 00IAAADT |   | HFS    | AI000035 |
|     | HFS  | ???  |     | 2000/09/01 | 07:29:51.05 | Sn    |      |      |     |     |      |      | 00IAAADT |   | HFS    | AI000036 |
|     | HFS  | ???  |     | 2000/09/01 | 07:30:05.78 | Lg    |      |      |     |     |      |      | 00IAAADT |   | HFS    | AI000037 |
|     | OBKA | ???  |     | 2000/09/01 | 07:29:09.50 | Pg    |      |      |     |     |      | ci   | 00IAAADU |   | VIE    | AI000038 |
|     | OBKA | ???  |     | 2000/09/01 | 07:29:16.10 | Sg    |      |      |     |     |      | i    | 00IAAADU |   | VIE    | AI000039 |

| Topic   | Seismic station codes – new coding standards                                     |
|---------|----------------------------------------------------------------------------------|
| Authors | IASPEI WORKING GROUP ON STATION CODES<br>Chairman: Ray Buland, retired from USGS |
| Version | August 2008; DOI: 10.2312/GFZ.NMSOP-2_IS_10.3                                    |

## Foreword by the Editor: The IASPEI resolution No. 3 - 2009

At its Scientific Assembly 2009 in Cape Town, South Africa, the International Association of Seismology and Physics of the Earth's Interior (IASPEI) adopted the following resolution:

*“Recognizing that the International Registry of Seismic Stations no longer meets the needs of the seismological community,*

*Recommends adoption of the newly developed seismic network and station coding standards developed by IASPEI to promote compatibility between waveform and parameter data exchange, attribution of parameter data, and flexibility for seismic network operators to better support earthquake monitoring and hazard assessment.*

The resolution is based on the August 2008 report of the IASPEI Working Group on Station Codes, which is reproduced below. The new registry is likely to become operational after some further discussion and testing around 2013.

## 1 Introduction

At the 2006 IASPEI meeting in Santiago, Chile, the Commission on Seismic Observation and Interpretation (CoSOI) Working Group on Seismic Network and Station Codes (WG-SNSC) began discussing how to update the International Registry (IR) of seismic station codes. Although the IR has served the international seismological community well for more than 40 years, it is no longer adequate to meet the evolving needs of network operators and the international data centers. At the 2007 IASPEI meeting in Perugia, Italy, the WG-SNSC approved the main points of the recommendation made in Santiago. On November 11-12, 2007, an *ad hoc* working group with representatives from ISC, NEIC, and EMSC met to discuss the implementation of the recommendation.

The following draft coding standards have been constructed from the recommendations as understood by the WG-SNSC, including comments on the recommendations from the International Federation of Digital Seismograph Networks (FDSN) as well as a number of network operators. In order to achieve the specificity required by the coding standards, the recommendations have been extended as needed. Note that in the following the words “shall” or “must” denotes proposed mandatory standards, while the word “should” denotes strongly recommended (non-mandatory) standards. The word “may” denotes acceptable options for usage.

This document describes how the coding standard would be used to register stations, submit parameter data to data centers, and attribute data summarized in catalogs to the correct agency. General steps and institutional responsibilities for implementing the new coding

standard are also discussed. However, details of the associated software implementation are beyond the scope of this document.

## 2 Coding standards

The location of seismic instrumentation shall be designated by the four fields: *Agency.Deployment.Station.Location*. The specification of a channel in connection with parameter data (e.g., to specify the channel from which a measurement was derived) shall be designated by the five fields: *Agency.Deployment.Station.Location.Channel*. Derived parameters (e.g., picks, amplitudes, hypocenters, magnitudes, etc.) shall be attributed to *Agency* or *Agency.Deployment*.

All fields shall be coded in alphabetic Latin characters (without diacritical marks) and/or Arabic numerals (i.e., characters represented by printable ASCII characters with numerical equivalents less than 128). Non-blank characters in any field shall be left justified and have no imbedded blanks. Non-blank characters in fixed field formats shall be blank padded to the right. All fields shall be case insensitive and have no other restrictions or extensions except as noted in the following. The fields shall be coded as follows:

- Agency: 1-5 alphanumeric characters.
- Deployment: 1-8 alphanumeric characters.
  - Special case: FDSN network codes used as deployment codes shall be case sensitive as required by the SEED v2.4 standard (note that in this case, the agency must be “FDSN”).
- Station: 1-5 alphanumeric characters (this relaxes current International Registry standards, but remains consistent with the SEED v2.4 standard).
- Location: 0-2 alphanumeric characters (e.g., SEED v2.4 compatible).
- Channel: 3 alphanumeric characters following the SEED v2.4 coding standard.

Agencies, deployments, stations, locations, and channels shall be defined in discrete time periods called epochs in the following. Each epoch shall be defined by start and end date-times. The end date-time of the current epoch shall be coded as “undetermined”. The relevant epoch shall be deduced by context (e.g., the date-time of parameter data) and looked up by date-time range rather than explicitly coded as an instance.

## 3 Usage standards

The following usage standards are required or recommended:

- The new coding standards will necessitate a significant revision in the role and operation of the International Registry. However, the ISC and the NEIC, in close cooperation, will continue to act as the central authority for collecting, organizing, and distributing basic seismic station information (e.g., coordinates, owners, operator(s), etc.). This will include tools supporting the simultaneous use of parametric and waveform data (e.g., providing the correspondence between the new coding standard and FDSN nomenclature).

- Scope rules:
  - Agency codes shall be centrally assigned (i.e., guaranteed to be globally unique by the International Registry).
  - Deployment codes shall be created by agencies as needed, but *Agency.Deployment* must be unique at any one time (i.e., duplicated *Agency.Deployment* codes must belong to non-overlapping epochs; see Example 8).
  - Station codes shall be created by agencies as needed, but *Agency.Deployment.Station* must be unique at any one time.
  - Location codes shall be created by agencies as needed, but *Agency.Deployment.Station.Location* must be unique at any one time.
  - The *Agency.Deployment.Station.Location.Channel* designation is intended to be applicable to all types of seismic installations (e.g., fixed and portable seismic networks, seismic arrays, strong motion networks, and strong motion arrays; see Examples 6-9).
- Reporting rules:
  - Deployment, station, location, and channel codes shall be reported to the International Registry for tracking purposes (e.g., to uniquely identify stations and channels and ensure correct attribution in the global bulletins).
    - Channel codes that are unavailable shall be assumed to be SHZ (i.e., short period vertical).
  - Channel codes should be reported along with parametric data to specify the waveform from which the parameters were derived.
  - Latitude, longitude, and elevation associated with each *Agency.Deployment.Station.Location* shall be reported to the International Registry for tracking purposes. Coordinates shall be reported relative to the WGS84 datum (or the actual datum shall be specified). Elevation shall be the elevation at the sensor.
  - Epochs defining the validity of an *Agency.Deployment.Station.Location* designation or changes in associated latitude, longitude, and elevation shall be reported to the International Registry for tracking purposes.
  - Stations, locations, channels, and epochs already documented in a dataless-SEED volume need not be reported separately.
- Usage rules:
  - *Agency.Deployment.Station.Location* shall uniquely identify the location of one or more seismic sensors or, alternatively, one or more seismic channels. This implies that each *Agency.Deployment.Station.Location* shall be associated with one latitude, longitude, and elevation from which parametric data is measured.
    - One *Agency.Deployment.Station.Location* may be used for more than one set of co-located seismic sensors (e.g., seismometers and accelerometers operated by one agency).
    - More than one *Agency.Deployment.Station.Location* may be associated with the same latitude, longitude, and elevation (e.g., to designate co-

located sensors operated by different agencies or to avoid the duplication of channel codes).

- o *Agency.Deployment.Station.Location.Channel* shall uniquely identify one channel associated with a particular sensor. Channels from co-located sensors that have the same names under SEED v2.4 rules shall have different *Agency.Deployment.Station.Location* designations (e.g., typically the location is made different).
- o The following guidelines for assigning station and location codes are recommended:
  - Station and/or location codes may be changed even if sensors have not been moved.
  - Elements of an array may be assigned the same station code and different location codes regardless of the scale of the array (see Example 6).
  - For the convenience of seismic analysts, it is strongly recommended that existing station codes should not be changed more often than absolutely necessary.
  - It is strongly recommended that station and/or location codes should be changed if the associated sensors are moved far enough to result in a significant teleseismic travel-time residual discrepancy (e.g., more than 0.2 s or about 1.2 km).

## 4 Alias standards

Although each *Agency.Deployment.Station.Location* must be associated with a unique latitude, longitude, and elevation, the inverse is not true (or desirable):

- Different *Agency.Deployment.Station.Location* codes may be associated with the same set of sensors (i.e., they are equivalent or aliased):
  - o For compatibility:
    - “FDSN” shall be assigned as an agency code.
      - FDSN network codes are legal deployment codes (although they are a special case because they are case sensitive).
      - When distributing seismic waveform data in FDSN standard SEED or mini-SEED, the FDSN centrally assigned network code shall be used.
        - o Note that there is no intent for *Agency.Deployment* to replace the FDSN network code now or in the future.
      - For correct attribution of parameter data, FDSN.*Network* designations can and should be aliased with the *Agency.Deployment* of the owner/operator (see Example 2).
      - All stations currently registered in the IR shall have an FDSN *Agency.Deployment* alias of FDSN.IR (as well as International Registry *Agency.Deployment* codes of ISC.IR and NEIC.IR; see Example 2).

- o To designate joint ownership/operation:
  - Many stations are owned/operated cooperatively by two or more agencies. Aliasing multiple *Agency.Deployment.Station.Location* codes provides a means for sharing attribution (see Example 4).
- o To designate logical participation:
  - Many stations may be thought of as being part of one or more networks (e.g., IMS) that have no role in owning or operating them (see Example 4).
- Completely separate (i.e., non-aliased) *Agency.Deployment.Station.Location* codes:
  - o Shall be associated with the same latitude, longitude, and elevation to designate co-located sensors operated by different agencies or as part of different deployments (see usage rules above and Example 3).
  - o May be associated with the same latitude, longitude, and elevation to designate co-located sensors operated by the same agency (see usage rules above and Example 7).
  - o Shall be associated with channels from co-located sensors that have the same names under SEED v2.4 rules (see usage rules above).

It is clear that in practice it will be desirable to create aliases at all levels in the *Agency.Deployment.Station.Location.Channel* hierarchy. For example, it may be desirable to alias all stations and locations associated with two or more equivalent *Agency.Deployment* identifiers or it may be desirable to alias all locations associated with two or more equivalent *Agency.Deployment.Station* identifiers. In addition, because aliases may be used for backwards compatibility, to show joint ownership, or to show affiliation with more than one network, it will be necessary to flag the type of each aliases.

## 5 Epoch standards

The following epoch standards are required or recommended:

- Epochs shall be associated with agencies, deployments, stations, locations, channels, and aliases.
- Agency and deployment epochs:
  - o If possible, the start date-time of the earliest epoch should be associated with the creation of the agency or deployment.
  - o Agency and deployment epochs shall be closed when the agency or deployment is terminated or its name is changed.
  - o If an agency or deployment is renamed, a new epoch with the start date-time of the renaming shall be created for the new *Agency.Deployment*.
- Station and location epochs:
  - o If possible, the start date-time of the earliest epoch should be associated with the opening of the station or the installation of equipment at a location.
  - o Station and location epochs shall be closed when the station or location is closed, moved, or renamed.

- If stations or locations have been moved or renamed, new epochs with the new coordinates or names shall be created.
  - For deployments adhering to FDSN rules, station, location, and channel epochs shall be coincident with station and channel epochs documented in the network SEED volumes.
- Alias epochs:
  - Aliases may change with time (e.g., if the agency operating a station changes or the agency changes names). These changes shall be documented in epochs associated with the aliases (see Example 5).
  - Alias epochs shall align with *Agency.Deployment.Station.Location* epochs when possible.

## 6 Display standards

It is expected that the *Agency.Deployment.Station.Location{.Channel}* nomenclature may be used in a variety of ways. The following display standards are recommended:

- Fixed format:
  - Each string should be left justified and blank padded within its field. A null location code should be represented as two blanks.
- Variable format:
  - Represented as a dotted string of the form *Agency.Deployment.Station.Location{.Channel}* with all blanks and trailing dots omitted (see Example 1).

## 7 Agency deployment style guide

The coding standard provides considerable flexibility in assigning agency and deployment codes. However, in creating and discussing the examples (below), it has become clear that some care should be taken in assigning agency codes in particular. For example, it is tempting to think of agencies within a larger organization as deployments (e.g., USGS.NEIC). However, since NEIC can also be considered an agency, this creates confusion and reduces flexibility (because it does not allow for deployments within NEIC).

For this reason, it is strongly suggested that agencies should be organizational units that operate networks or create parameter data. This allows agencies to have multiple network deployments. The fact that an agency may be part of a larger organization will be captured in the description of each agency (e.g., NEIC can be described as the National Earthquake Information Center, US Geological Survey, Department of Interior, US Federal Government).

## 8 Implementation

The ISC and NEIC will work together to implement the revised International Registry. This will include:

- Developing the underlying database structures to support the new coding, epochs, and aliases.
- Populating the new database with the currently registered stations (using the ISC.IR, NEIC.IR, and FDSN.IR deployments).
- Populating the new database with channels currently documented in dataless SEED volumes.
- Working with the seismological agencies and the FDSN to create aliases to existing FDSN network codes.
- Revising web forms to register agencies and report deployments, stations, locations, channels, epochs, and aliases.
- Updating the database from the web form input.
- Making the contents of the new database accessible by means of an interactive (i.e., human) web interface and a software accessible (i.e., program-to-program) network service.
- Developing strategies to attribute parameter data contributed in existing formats to the correct agencies.
- Working with seismic processing system developers to add *Agency.Deployment.Station.Location.Channel* information to contributed parameter data.

## 9 Examples

In the following, the symbol == is used to designate an alias. Note that in the following there are examples of *Agency.Deployment* aliasing, *Agency.Deployment.Station* aliasing, *Agency.Deployment.Station.Location* aliasing, and

*Agency.Deployment.Station.Location.Channel* aliasing. While some attempt has been made to find real-world examples, these examples have been modified as needed to illustrate particular points.

**Example 1 – displaying a null location code:** The US National Earthquake Information Center (NEIC) operates the Advanced National Seismic System (ANSS) Backbone network, which includes station Dugway (DUG). Dugway has only one location, which has a null location code. The station could be designated as NEIC.ANSSBN.DUG. The broadband, high-gain, vertical channel could be designated as NEIC.ANSSBN.DUG..BHZ.

**Example 2 – default aliasing:** The Geological Survey of Canada (GSC) operates the Canadian National Seismograph Network (CNSN), which includes the station Whitehorse (WHY). By default this station might be known as GSC.CNSN.WHY == ISC.IR.WHY ==



NEIC.IR.WHY. If the network had also been assigned an FDSN network code (CN in this case), it would also have aliases of FDSN.CN.WHY and FDSN.IR.WHY

**Example 3 – designating co-located stations:** The Geophysical Institute of Israel (GII) operates the Israeli Seismic Network (ISN), the Israeli Seismic Network Backup (ISNB) and the Israeli Strong Motion Array (ISMA), all of which have sensors located at Eilat (EIL). These sensors might be registered separately as GII.ISN.EIL (== FDSN.IS.EIL), GII.ISNB.EIL, and GII.ISMA.EIL respectively. Note that only one of these stations would be aliased with ISC.IR.EIL, NEIC.IR.EIL, and FDSN.IR.EIL (e.g., with GII.ISN.EIL, but not with GII.ISNB.EIL or GII.ISMA.EIL).

**Example 4 – designating a cooperative station:** The Geophysical Institute of Israel (GII) operates the Israeli Seismic Network (ISN), which includes station Eilat (EIL). EIL is also part of the GeoForschungsZentrum (GFZ) GEOFON network and is station AS48 of the Comprehensive Test Ban Treaty Organization (CTBTO) International Monitoring Network (IMS). These relationships might be shown by aliasing GII.ISN.EIL == GFZ.GEOFON.EIL == CTBTO.IMS.AS48. Note that these aliases may not imply the same sort of relationship (e.g., GII and GFZ may have cooperated in establishing the station, but the GII may only be contributing data to the CTBTO).

**Example 5 – alias epochs:** The University of Utah at Salt Lake City (UUSLC) operates a statewide network, which includes the station San Rafael Swell (SRU). After July 1, 2007, this station became affiliated with the ANSS Backbone network. Therefore, prior to July 1, 2007 the aliasing might have been UUSLC.UU.SRU == FDSN.UU.SRU. After July 1, 2007, the aliasing might be UUSLC.UU.SRU == FDSN.UU.SRU == NEIC.ANSSBN.SRU. That is, a new alias epoch would show that the station was not part of the ANSSBN deployment before July 1, 2007.

**Example 6 – designating an array:** The US National Data Center (USNDC) operates a number of arrays as the US contribution to the Comprehensive Test Ban Treaty Organization (CTBTO). The thirteen elements of the Pinedale (PDAR) array could be designated as CTBTO.USNDC.PD01, CTBTO.USNDC.PD02,... or as CTBTO.USNDC.PDAR.01, CTBTO.USNDC.PDAR.02,... Array elements can be treated as separate stations or as locations of one station, regardless of the separation.

**Example 7 – designating co-located sensors:** The US National Data Center (USNDC) operates a number of arrays as the US contribution to the Comprehensive Test Ban Treaty Organization (CTBTO). The central array element includes co-located broadband and short period sensors. These could be shown as CTBTO.USNDC.PD31..BHZ and CTBTO.USNDC.PD32..SHZ or as CTBTO.USNDC.PD31..BHZ and CTBTO.USNDC.PD31..SHZ. That is, the station codes could follow either the USNDC convention (different station codes for co-located sensors) or the current IR convention.

**Example 8 – designating portable deployments:** Suppose US National Earthquake Information Center (NEIC) portable networks are deployed in Marble, Colorado, USA at two different times. These might both be designated as NEIC.MARBLE in two non-overlapping

epochs (e.g., epoch 1 from 1997-06-15 to 1997-07-02 and epoch 2 from 2001-08-22 to 2002-06-10).

**Example 9 – designating a strong motion array:** Suppose the US National Strong Motion Program (NSMP) deployed a strong motion array in the (fictional) Benz building. The basement sub-array might be designated NSMP.BENZ.BSMT.NE, NSMP.BENZ.BSMT.NW,....

# Chapter 11

## Data Analysis and Seismogram Interpretation

(Version March 2014; DOI:10.2312/GFZ.NMSOP-2\_ch11)

Peter Bormann<sup>1)</sup>, Klaus Klinge<sup>2)</sup>, and Siegfried Wendt<sup>3)</sup>

- <sup>1)</sup> Formerly GFZ German Research Centre for Geosciences, Department 2: Physics of the Earth, Telegrafenberg, 14473 Potsdam, Germany; E-mail: [pb65@gmx.net](mailto:pb65@gmx.net);  
<sup>2)</sup> Formerly Bundesanstalt für Geowissenschaften und Rohstoffe, Stilleweg 2, 30655 Hannover Germany; E-Mail: [klaus.klinge@gmail.com](mailto:klaus.klinge@gmail.com)  
<sup>3)</sup> Geophysical Observatory Collm, University of Leipzig, 04779 Wermsdorf, Germany; E-mail: [wendt@rz.uni-leipzig.de](mailto:wendt@rz.uni-leipzig.de)

**Note:** When reference is made to figures then the first figure number is the respective Chapter number.

|                                                                                                  | <b>Page</b> |
|--------------------------------------------------------------------------------------------------|-------------|
| <b>11.1 Introduction</b>                                                                         | 2           |
| <b>11.2 Criteria and parameters for routine seismogram analysis</b>                              | 9           |
| 11.2.1 Record duration and dispersion                                                            | 9           |
| 11.2.2 Key parameters to be measured                                                             | 10          |
| 11.2.2.1 Phase onset time                                                                        | 10          |
| 11.2.2.2 Amplitudes and period                                                                   | 12          |
| 11.2.2.3 First-motion polarity                                                                   | 14          |
| 11.2.3 Advanced wavelet parameter reporting from digital records                                 | 15          |
| 11.2.4 Criteria to be used for phase identification                                              | 16          |
| 11.2.4.1 Travel-time and slowness                                                                | 16          |
| 11.2.4.2 Amplitude-distance relations, dominating periods and waveforms                          | 19          |
| 11.2.4.3 Polarization                                                                            | 26          |
| 11.2.4.4 Example for documentation and reporting of seismogram readings                          | 29          |
| 11.2.5 Criteria to be used in event identification and discrimination                            | 31          |
| 11.2.5.1 Discrimination between shallow and deep earthquakes                                     | 31          |
| 11.2.5.2 Discrimination between natural earthquakes and man-made seismic events                  | 34          |
| 11.2.6 Quick event identification and location by means of single-station 3-component recordings | 38          |
| 11.2.6.1 What to do with single station seismograms?                                             | 39          |
| 11.2.6.2 3-component hypocenter location                                                         | 40          |
| 11.2.7 Hypocenter location by means of network and array recordings                              | 46          |
| 11.2.8 Magnitude determination                                                                   | 50          |

|             |                                                                                                             |         |
|-------------|-------------------------------------------------------------------------------------------------------------|---------|
| <b>11.3</b> | <b>Routine signal processing of digital seismograms</b>                                                     | 52      |
| 11.3.1      | Signal detection                                                                                            | 52      |
| 11.3.2      | Signal filtering, restitution and simulation                                                                | 53      |
| 11.3.3      | Signal coherence at networks and arrays                                                                     | 64      |
| 11.3.4      | f-k and vespagram analysis                                                                                  | 66      |
| 11.3.5      | Beamforming                                                                                                 | 67      |
| 11.3.6      | Polarization analysis                                                                                       | 69      |
| <b>11.4</b> | <b>Software for routine analysis</b>                                                                        | 70      |
| 11.4.1      | Brief overview and software library links                                                                   | 70      |
| 11.4.2      | SHM                                                                                                         | 72      |
| 11.4.3      | Seisan                                                                                                      | 73      |
| 11.4.4      | SeisComP3                                                                                                   | 75      |
| 11.4.5      | Earthworm                                                                                                   | 77      |
| 11.4.6      | ObsPy                                                                                                       | 79      |
| <b>11.5</b> | <b>Examples of seismogram analysis</b>                                                                      | 79      |
| 11.5.1      | Introductory remarks                                                                                        | 79      |
| 11.5.2      | Seismograms from near and regional sources ( $0^\circ < D < \approx 20^\circ$ )                             | 83      |
| 11.5.3      | Teleseismic earthquakes ( $\approx 20^\circ < D < 180^\circ$ )                                              | 91      |
| 11.5.3.1    | Distance range $\approx 20^\circ < D < 100^\circ$                                                           | 91      |
| 11.5.3.2    | Distance range $100^\circ < D < 144^\circ$                                                                  | 96      |
| 11.5.3.3    | Core distance range beyond $144^\circ$                                                                      | 100     |
| 11.5.4      | Late and very late core phases                                                                              | 102     |
| 11.5.5      | Final remarks on the recording and analysis of teleseismic events                                           | 108     |
| <b>11.6</b> | <b>Is fully automatic seismogram analysis already possible and feasible?</b>                                | 110     |
| 11.6.1      | Introduction                                                                                                | 110     |
| 11.6.2      | Amplitude/energy event triggers, phase identification and event location                                    | 112     |
| 11.6.3      | Machine learning approaches for improving event detection and classification at single 3-component stations | 117     |
|             | <b>Acknowledgments</b>                                                                                      | 118     |
|             | <b>Recommended overview readings</b>                                                                        | 119     |
|             | <b>References</b>                                                                                           | 119-126 |

## 11.1 Introduction

Chapter 11 deals with seismogram analysis and extraction of seismic parameter values for data exchange with national and international data centers, for use in research and last, but not least, for informing the public about seismic events. It is mainly written for training purposes and for use as a reference source for seismologists working at observatories, network and analysis centers or with mobile networks for monitoring and analyzing the seismotectonic activity. Complementary, procedures for volcano-seismic monitoring are treated in detail in Chapter 13. However, this Chapter deliberately refers also to classical analog or simple computer assisted interactive procedures and products of seismogram analysis. They are easier to grasp and educationally often more illustrative and appealing. Accordingly we recall with this Chapter also historical aspects of seismic recording and seismogram analysis and provide information and procedures applicable also in the context of education and outreach programs. Thus we hope to address, as the related IRIS program ([http://www.iris.edu/hq/programs/education\\_and\\_outreach](http://www.iris.edu/hq/programs/education_and_outreach)), at least with some sections a wider than specialist audience, including also grade 6-12 students and teachers involved in

“Seismology and Seismographs at Schools”, college students and faculty members, researchers from neighboring disciplines and the public.

Accordingly, the Chapter describes the basic requirements, tools and products of both traditional analog and nowadays almost exclusively digital routine observatory practice, i.e.:

- recognize the occurrence of an earthquake in a record;
- identify and annotate the seismic phases;
- determine onset time and polarity correctly;
- measure the maximum ground amplitude and related period;
- calculate slowness and azimuth;
- determine source parameters such as the hypocenter, origin time, magnitude, etc.

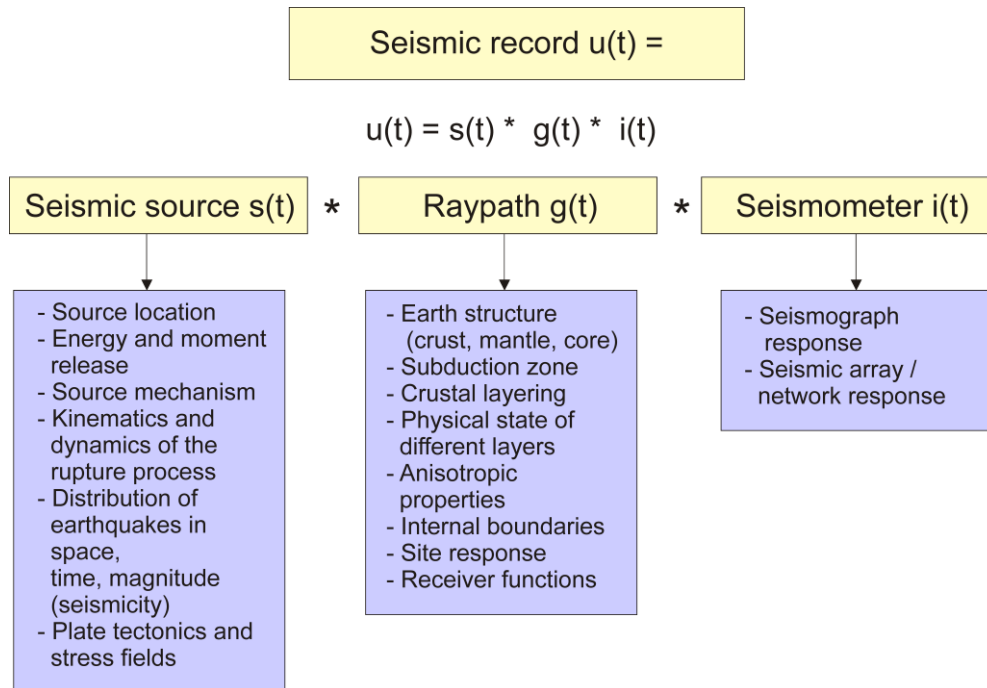
In modern digital observatory practice these procedures are implemented in computer programs. Experience, a basic knowledge of elastic wave propagation (Chapter 2), and the available software can guide a seismologist to analyze large amounts of data and interpret seismograms correctly. Nowadays, in most observatories this work is partially or completely automatized. Earthquake detection, initial P-phase picking, initial localization and magnitude determination are normally carried out by the automatic procedure. Seismologists, using graphical processing and visualization tools, then interactively revise and correct these initial results based on scientific reasoning, ample knowledge and – very important - much practical experience. The aim of this Chapter is to introduce the basic knowledge, data, procedures and tools required for proper seismogram analysis and phase interpretation and to illustrate this with selected seismogram examples. Very rapidly developing new procedures and data products will have to be described in specialized future amendments to this Chapter.

Seismograms reflect the combined influence of the seismic source (Chapter 3), the propagation path (Chapter 2), the frequency response of the recording instrument (Chapter 4, section 4.5 and Chapter 5, section 5.2.8), and the ambient noise at the recording site (Chapter 4, section 4.6). Fig. 11.1 summarizes some essential fields of study and kinds of information which benefit or are largely based on the analysis and interpretation of seismograms. The more completely we interpret the seismograms and the better we quantify the parameters that can be derived therefrom, the better we reveal and understand the structure and properties of the Earth as well as the geometry, kinematics and dynamics of seismic sources and the underlying processes.

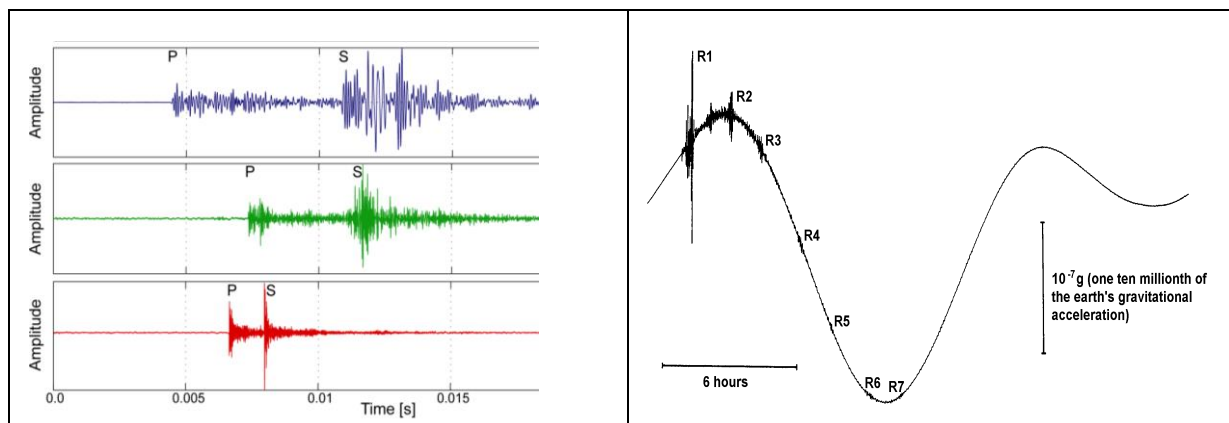
Seismograms carry basic information about earthquakes, chemical and nuclear explosions, mining-induced earthquakes, rock bursts and other events generating seismic waves in a very wide range of frequencies from kHz to mHz recorded at distances from a few meters to several 10,000 kilometers from the source (Fig. 11.2).

Seismological data analysis for single stations is nowadays increasingly replaced by network (Chapter 8) and array analysis (Chapter 9). Array-processing techniques have been developed for more than 50 years. Networks and arrays, in contrast to single stations, enable better signal detection and source location. Also, arrays can be used to estimate slowness and azimuth, which allow better phase identification. Further, more accurate magnitude values can be expected by averaging single station magnitudes and for distant sources the signal coherency can be used to determine onset times more reliably. Tab. 11.1 summarizes basic characteristics of single stations, station networks and arrays. In principle, an array can be used as a network and in special cases a network can be used as an array. The most important

differences between networks and arrays are in the degree of signal coherence and the data analysis techniques used.



**Fig. 11.1** Different sub-systems (without seismic noise) which influence a seismic record (yellow boxes) and the information that can be derived from the analysis of seismic records (blue boxes). The ‘\*’ represents the convolution operator.



**Fig. 11.2 Left:** Very weak mining induced seismo-acoustic events with frequencies between about 1 and 100 kHz recorded at distances from meters to decameters in a South African goldmine (cut-out from Figure 2A of Plenkers et al. (2010), Seism. Res. Lett., vol. 81, no. 3, p. 470; © Seismological Society of America; **Right:** Very broadband STS1 record with seismic wave groups from a magnitude 8.2 earthquake in the Kermadec Islands (October 20, 1986) that are superimposed to solid Earth's tides. The mantle Rayleigh wave R7 has traveled almost 130000 km before it was recorded at the station of Nagoya University, Japan (modified from a pamphlet of the Japanese Global Seismology Subcommittee for the POSEIDON project).

**Tab. 11.1** Short characteristic of single stations, station networks and arrays.

|                        |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Single station</b>  | Classical type of seismic station with its own data processing. Event location only possible by means of three-component records.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>Seismic network</b> | Local, regional or global distribution of stations, physical or ad hoc virtuell, that are as identical as possible, with a common data center and time reference (Chapter 8). Event source parameter estimation (hypocenter, magnitude, mechanism...) is one of the main tasks.                                                                                                                                                                                                                                                                                                                                                        |
| <b>Seismic array</b>   | Cluster of seismic stations with a common time reference and uniform instrumentation. The stations are located close enough to each other in space for the signal waveforms to be correlated between adjacent sensors (Chapter 9). Benefits are: <ul style="list-style-type: none"> <li>• extraction of coherent signals from random noise;</li> <li>• determination of directional information of approaching wavefronts (determination of backazimuth of the source);</li> <li>• determination of slowness and thus of epicentral distance of the source;</li> <li>• event localization by using backazimuth and distance</li> </ul> |

Like single stations, also narrow band-limited seismometer systems loose nowadays increasingly in importance for global monitoring. However, the much easier to handle and cheaper short-period seismographs, which are particularly suitable for local monitoring of weaker events, still outnumber by far the more costly and with respect to environmental effects much more sensitive broadband systems (see <http://www.iris.edu/data/DCProfiles.htm> as well as Chapter 5 and IS 5.4). In any event, bandwidth limitation filters the ground motion, thus distorting the signal shape, the more the narrower the bandwidth. In the extreme it may even shift the apparent onset time and reverse the apparent first-motion polarity (Chapter 4, section 4.5). Most seismological observatories, and especially regional and global networks, are now equipped with broadband seismometers that are able to record signal frequencies between about 0.001 Hz and 50 Hz. The frequency and dynamic range covered by broadband recordings are shown in Figs. 1.9, 4.7 and 11.38 in comparison with classical band-limited analog recordings of the Worldwide Standard Seismograph Network (WWSSN).

A number of classical analog seismograph systems are still in operation at autonomous single stations in some countries. Also, archives are filled with analog recordings of these systems, which were collected over many decades. These data constitute a wealth of information most of which has yet to be fully analyzed and scientifically exploited. Although digital data are superior in many respects, both for advanced routine analysis and even more for scientific research, it will be many years or even decades of digital data acquisition before one may consider the bulk of these old data as no longer needed. However, for the rare big and thus unique earthquakes, and for earthquakes in areas with low seismicity rates, the preservation and comprehensive analysis of these classical and historic seismograms will remain of utmost importance for many years. Around 300,000 paper seismograms, especially those containing large earthquakes, were already scanned and digitized in the IASPEI SeismoArchives Project ([http://www.iaspei.org/downloads/IASPEI\\_SeismoArchives\\_Project\\_summary\\_2011.pdf](http://www.iaspei.org/downloads/IASPEI_SeismoArchives_Project_summary_2011.pdf),

<http://www.iris.edu/seismo/>) but this is still a small fraction of the total amount of analog recordings.

More and more old analog data will be reanalyzed only after being digitized and by using similar procedures and analysis programs as for recent original digital data. Nevertheless, station operators and analysts should still be in a position to handle, understand and properly analyze analog seismograms or plotted digital recordings without computer support and with only modest auxiliary means. Digital seismograms are analyzed in much the same way as classical seismograms, except that the digital analysis uses interactive software which makes the analysis quicker and easier, allows comfortable modification of the time and amplitude resolution of the record plots, modification of the record response and component representation (filtering, record simulation, axes rotation) and to use many other useful operations. But the correct seismogram interpretation requires much of the same knowledge of the appearance of seismic records and of individual seismic phases as the interpretation of classical analog records. The analyst needs to know the typical features in seismic records as a function of distance, depth and source process of the seismic events, their dependence on the polarization of the different types of seismic waves and thus on the azimuth of the source and the orientation of sensor components with respect to it. He/she also needs to be aware of the influence of the seismograph response on the appearance of the record. Without this solid background knowledge, phase identifications and parameter readings may be rather incomplete, systematically biased or even wrong, no matter what kind of sophisticated computer programs for seismogram analysis are used.

Therefore, in this Chapter we will deal first with an introduction to the fundamentals of seismogram analysis at single stations and seismic networks, based on waveform data. Although many decision makers in seismology see nowadays less operational need for this kind of instruction and training, from an educational point of view its importance cannot be overemphasized. An analyst trained in comprehensive and competent analysis of traditional analog seismic recordings, when given access to advanced tools of computer-assisted analysis, will by far outperform any computer specialist with only limited seismological background knowledge.

Automated phase identification and parameter determination have advanced significantly in recent years (Chapter 16) and enjoy massive application at seismological observatories, physical and virtual network centers (Chapter 8) and for seismic arrays (Chapter 9). Only with automation the data from a steadily increasing number of seismic stations can be processed, especially during seismic swarms and aftershock activities. Still, the quality of automatic results might be inferior to those achievable by well-trained analysts (section 11.4). Therefore, an adequate combination of automatic and manual data processing should be assured. Both processing phases and their interaction must be optimized. In many observatories the automatic procedures are only used for fast information to the authorities and the public and run simply in parallel to the manual data processing which intends to produce high quality results for later scientific applications. Therefore, this Manual focusses on providing competent guidance and advice to station operators and seismologists with limited experience on the complexity of seismic waveforms, their comprehensive analysis and interpretation. Moreover, specialists in program development and the automation of algorithms usually also lack the required seismological background knowledge and own rich experience in seismogram analysis. This, however, is an indispensable requirement for further qualification of currently available computer procedures for interactive and automatic data analysis, parameter determination and source location. They have to be in tune both with



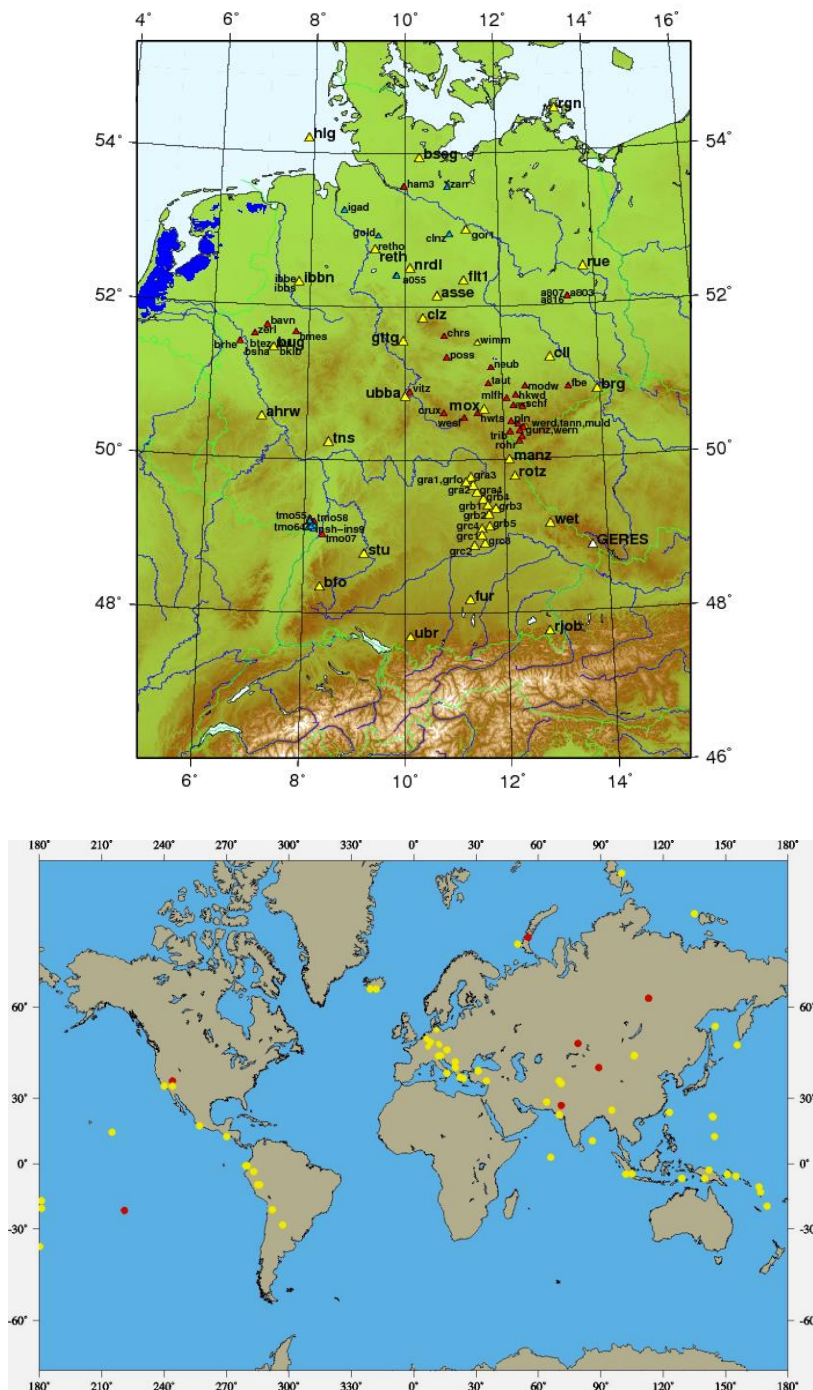
older data and established standards to assure long-term data continuity and compatibility and at the same time also assure a broader range of quality observatory products which meet modern research requirements. In this sense, the chapter also addresses the needs of this advanced user community.

Digital recording has the big advantage that the seismic recordings can be reprocessed easily when new processing methods are available. It should also be noted that the falling prices of digital storage media make continuous recording and backup economically feasible. So, new detection algorithms can be applied to the recordings, or the parameters of the algorithms can be tuned by comparing their results with those of human interpretation.

Accordingly, we first give a general introduction to “classical” routine seismogram interpretation which was developed using analog records, respectively waveform plots of digital records at single stations and local or regional seismic networks. Then we discuss both similarities and differences when processing analog or digital data. The basic requirements for parameter extraction, bulletin production as well as parameter and waveform data exchange are also outlined. In the sub-Chapter on digital seismogram analysis we discuss in more detail problems of signal coherence, the related different procedures of data processing and analysis as well as available software for it. The majority of record examples from Germany have been processed with the program Seismic Handler (SHM) developed by K. Stammer which is used for seismic waveform retrieval and data analysis. This program and descriptions are available via <http://www.seismic-handler.org>. Reference is made and examples are given, however, also to/from other analysis software that is widely used internationally, e.g. SEISAN, SeisComp3, Earthworm, SAC and many others (see 11.4). Typical examples of seismic records from different single stations, networks and arrays in different distance ranges (local, regional and teleseismic) and at different source depth are presented, mostly broadband data or filtered records derived therefrom. Special sections are dedicated to the interpretation of seismic core phases (see 11.5.3.3 and 11.5.4).

The treatment of the detailed analysis of teleseismic recordings might provide an interesting, even “exotic” view for readers working at observatories with a high local seismicity where the work is often uniquely focused on the processing of near seismic events. At many networks in areas with complicated crustal structures (e.g., in subduction environments) the locations are normally only carried out with the first P- and S-waves and the high workload due to a strong seismicity does not permit to look for secondary or teleseismic arrivals. But, it should be noted that an understanding of teleseismic records is sometimes very beneficial for local network seismologist too. It happens, for instance, that strong relatively high-frequency teleseismic P or PKP arrivals, under the bad luck of low residuals due to some casual station geometry, are misinterpreted by automatic location procedure as phases from a local or regional event and then a “ghost” earthquake is reported. Another difficulty for automatic and even human interpretation is the recording of two or more seismic events at nearly the same time. This is not so seldom, especially for extended regional networks, and during aftershock or swarm activities in the monitored area. The automatic procedures, which assign the detected phases to earthquake origins, fail occasionally in this situation and then human interaction is necessary.

Since all authors of Chapter 11 come from Germany, the majority of records shown have been collected at stations of the German Regional Seismic Network (GRSN), the Gräfenberg array (GRF) or at other local station networks (Fig. 11.3a). Fig. 11.3b depicts the global distribution of epicenters of seismic events from which record examples are presented.



**Fig. 11.3** **a)** Broadband stations in Germany. Yellow triangles: stations of the German Regional Seismic Network (GRSN) and of the Gräfenberg-Array (GRF). Other triangles: other stations (for details Chapter 8, section 8.7.5). **b)** Global distribution of epicenters of seismic events (red dots: underground nuclear explosions; yellow dots: earthquakes) for which records from the above stations will be presented in Chapter 11 and DS 11.1-11.4.

All GRSN/GRF stations record originally velocity-broadband (BB-velocity) data. All record examples from these stations of short-period (SP), long-period (LP) or BB-displacement seismograms are digitally simulated ones, corresponding to those which would have been

recorded by classical Wood-Anderson, WWSSN-SP, WWSSN-LP, SRO-LP or Kirnos SKD seismographs. Users of this Chapter may feel that the seismograms presented are too much biased towards Europe. Indeed, we may have overlooked some important aspects or typical seismic phases which are well observed in other parts of the world. Therefore, we invite anybody to send valuable complementary data and information to the Editor for integration into revised or amended editions of this electronic Manual.

For routine analysis and international data exchange a standard nomenclature of seismic phases is required. The latest version of the IASPEI Standard Nomenclature of Seismic Phases is given in IS 2.1, together with ray diagrams for most phases. This nomenclature (published earlier in Storchak et al., 2002, 2003 and 2011) partially modifies and complements earlier ones, e.g., that in the last edition of the Willmore (1979) Manual of Seismological Observatory Practice (and in pre-2010 issues of the seismic Bulletins of the International Seismological Centre (ISC). It is more in tune than the earlier versions with the phase definitions of modern Earth and travel-time models (Chapter 2, section 2.7).

The scientific fundamentals of some essential subroutines in seismological analysis software, e.g., on event location and travel-time calculations, are separately treated in annexed material such as IS 11.1, IS 11.3, PD 11.1 and PD 11.2. Complementary tutorials, exercises and information sheets, also suitable for demonstration and practicals in undergraduate courses and “Seismology in Schools” explain in detail basic steps in the interpretation process (e.g., EX 11.1-11.3 and IS 11.6). More related information may be added in the course of future upgrading of this electronic Manual.

## **11.2 Criteria and parameters for routine seismogram analysis**

### **11.2.1 Record duration and dispersion**

The first thing an experienced seismologist looks for when assessing a seismic record is the duration of the signal. Due to the different nature and propagation velocity of seismic waves and the different propagation paths taken by them to a station, travel-time differences between the main wave groups usually grow with distance. Accordingly, the record spreads out in time. The various body-wave groups show no dispersion, so their individual duration remains more or less constant, only the time difference between them changes with distance. The time difference between the main body-wave onsets is roughly  $< 3$  minutes for events at distances  $D < 10^\circ$ ,  $< 16$  min for  $D < 60^\circ$ ,  $< 30$  min for  $D < 100^\circ$  and  $< 45$  min for  $D < 180^\circ$  (Figs. 1.4b, 1.6 and 2.60).

In contrast to body waves, the velocity of surface waves is frequency dependent and thus surface waves are dispersed. Accordingly, depending on the crustal/mantle structure along the propagation path, the duration of Love- and Rayleigh-wave trains increases with distance. At  $D > 100^\circ$  surface wave seismograms may last for an hour or more (Fig. 1.6), and for really strong events, when surface waves may circle the Earth several times, their oscillations on sensitive long-period (LP) or broadband (BB) records may be recognizable over 6 to 12 hours (Fig. 11.2 right). Even for reasonably strong regional earthquakes, e.g.,  $M_s \approx 6$  and  $D \approx 10^\circ$ , the oscillations may be recognizable above the noise level for about an hour although the time difference between the P and S onset is only about 2 min and between P and the maximum amplitude in the surface wave group only 5-6 min. The magnitude dependence of the total earthquake signal duration led the Hungarian seismologist Bisztricsany (1958) to propose a

simple teleseismic procedure for estimating the magnitude of earthquakes that is independent on the knowledge of the distance-dependent amplitude attenuation of different seismic phases and the risk of amplitude clipping of analog records in the case of very strong events.

Finally, besides proper dispersion, scattering may also spread wave energy. This is particularly true for the more high-frequency waves traveling in the usually heterogeneous crust. This gives rise to signal-generated noise and coda waves. Coda waves follow the main generating phases with exponentially decaying amplitudes. The coda duration depends mainly on the event magnitude (Figure 1b in DS 11.1) and only weakly on epicentral distance (Figure 2 in EX 11.1). Thus, record duration can be used for calculating also local coda or duration magnitudes  $M_c$  or  $M_d$  (Chapter 3, section 3.2.4.5).

In summary, signal duration, the time difference between the Rayleigh-wave maximum and the first body-wave arrival (Table 5 in DS 3.1) and in particular the time span between the first and the last recognized body-wave onsets before the arrival of surface waves allow a first rough estimate, whether the earthquake is a local, regional or teleseismic one. This rough classification is a great help in choosing the proper approach, criteria and tools for further more detailed seismogram analysis, source location, and magnitude determination.

## 11.2.2 Key parameters to be measured

### 11.2.2.1 Phase onset time

Onset times of seismic wave groups, first and foremost of the P-wave first arrival, when determined at many seismic stations at different azimuth and at different distance, but also of later arriving phases, are the key input parameter for the location of seismic events (IS 11.1). Travel times published in travel-time tables (such as Jeffreys and Bullen, 1940; Herrin, 1968; Kennett, 1991) and travel-time curves, such as those shown in Figs. 2.48, 2.56 or 2.60 have been derived either from observations or Earth models. They give, as a function of epicentral distance  $D$  and hypocentral depth  $h$ , the differences between *onset times*  $t_{ox}$  (also called *arrival times*) of the respective seismic phases  $x$  and the *origin time*  $OT$  of the seismic source. This time difference is called *travel time*. Onset times mark the first energy arrival of a seismic wave group. The process of recognizing and marking a wave onset and of measuring its onset time is termed *onset time picking*. IS 11.4 provides an elaborated tutorial for both manual and computer assisted consistent picking of phases at local and teleseismic distances and IS 11.5 looks into the problem of picking errors of local phases depending on the signal-to-noise ratio (SNR) as well as on the training and years of experience of analysts and provides a training data set to check analyst performance and the effect of training and gained experience.

The recognition of a wave onset largely depends on the SNR at various frequencies for the given waveform as a whole and the steepness and amplitude of its leading edge. Both are modulated by the shape and bandwidth of the recording seismograph or filter (Figs. 4.9 to 4.13). It is a classical convention in seismological practice to classify onsets, as a qualitative measure for the reliability of their time-picking, as either impulsive (i) or emergent (e). These lower case letters i or e are put in front of the phase symbol. Nowadays, however, with the time resolution and amplitudes and thus the steepness of the waveform flanks being manipulated at will with modern analysis systems, it has become more common to attribute instead a different weight to the picked onset time and polarity readings (see also IS 11.4).

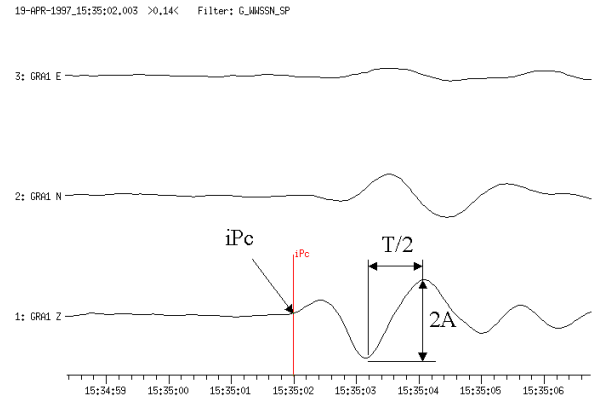
Generally, it is easier to recognize and precisely pick the very first arrival (usually a P wave) on a seismogram than later phases that arrive within the signal-generated coda of earlier waves, often termed “signal-generated noise”.

The relative precision with which an onset can be picked largely depends on the factors discussed above, but the absolute accuracy of onset-time measurement is controlled by the available time reference. Seismic body-wave phases travel rather fast. Their apparent velocities at the surface typically range between about 3 km/s and over 100 km/s (at the antipode the apparent velocity is effectively infinite). Therefore, an absolute accuracy of onset-time picking of less than a second and ideally less than 0.1 s is needed for estimating reliable epicenters (IS 11.1) and determining good Earth models from travel-time data. This was difficult to achieve in earlier decades when only mechanical pendulum clocks or marine chronometers were available at most stations. They have unavoidable drifts and could rarely be checked by comparison with radio time signals more frequently than twice a day. Also, the time resolution of classical paper or film records is usually between 0.25 to 2 mm per second, thus hardly permitting an accuracy of time-picking better than parts of a second. In combination with the limited timing accuracy, the reading errors at many stations of the classical world-wide network, depending also on distance and region, were often two to three seconds (Hwang and Clayton, 1991). However, this improved since the late 1970s with the availability of very-low frequency and widely received time signals, e.g., from the DCF and Omega time services, and recorders driven with exactly 50 Hz stabilized alternating current.

Yet, onset-time reading by human eye from analog records with minute marks led to sometimes even larger errors, a common one being the  $\pm 1$  min for the P-wave first arrival. This is clearly seen in Fig. 2.56 (left), which shows the travel-time picks collected by the ISC from the world-wide seismic station reports between 1964 and 1987. Nowadays, atomic clock time from the satellite-borne Global Positioning System (GPS) is readily available in nearly every corner of the globe. Low-cost GPS receivers are easy to install at both permanent and temporary seismic stations and generally affordable. Therefore the problem of unreliable absolute timing should no longer exist. However, technical problems do occur and one needs to be aware of potential timing problems. In fact, there still are many GPS driven digitizers with wrong timing. Nevertheless, also with high resolution digital data and exact timing now being available it is difficult to decide on the real signal onset, even for sharp P from explosions. Douglas et al. (1997) showed that the reading errors have at best a standard deviation between 0.1 and 0.2 s. However, human reading errors no longer play a major role when digital data are evaluated by means of seismogram analysis software which automatically records with high precision the time at the positions where onsets have been marked with a cursor. Moreover, the recognition of signal onsets and the precision of time picks can be modified easily in the case of digital records within the limits set by the sampling rate and the dynamic range of recording. Both the time and amplitude scales of a record can be compressed or expanded as needed, and task-dependent optimal filters for best phase recognition can be applied as well.

Fig. 11.4 shows such a digital record with the time scale expanded to 12 mm/s. The onset time can be reliably picked with an accuracy of a few tenths of a second. This P-wave first arrival has been classified as an impulsive (i) onset, although it looks emergent in this time-stretched plot. But by expanding the amplitude scale also, the leading edge of the wave arrival becomes steeper and so the onset appears impulsive. This ease with which digital records can be manipulated reduces the representativeness and thus the value of qualitative characterization of onset sharpness by either i or e although there are in fact “e” type of phases where no

change in scaling allows a clear characterization. Alternatively, it has been proposed, therefore, to quantify the *onset-time reliability*. This could be done by reporting, besides the most probable or interpreter-preferred onset time, the estimated range of uncertainty by picking the earliest ( $t_{ox-}$ ) and latest possible onset time ( $t_{ox+}$ ) for each reported phase  $x$ , and of the first arrival in particular (Fig. 11.6 ).



**Fig. 11.4** First motion onset times, phase and polarity readings (c – compression; d – dilatation), maximum amplitude  $A$  and period  $T$  measurements for a sharp (i - impulsive) onset of a P wave from a Severnaya Zemlya event of April 19, 1997, recorded by a broadband three-component single station of the Gräfenberg Array (GRF), Germany but filtered so as to simulate a short-period record of the World-Wide Standard Seismograph Network (WWSSN\_SP).

### 11.2.2.2 Amplitude and period

Whereas the quality, quantity and spatial distribution of reported arrival-time picks largely controls the precision of *source locations* (see IS 11.1), the quality and quantity of amplitude readings for identified specific seismic phases determine the representativeness of amplitude-based event *magnitudes*. The latter are usually based on readings of maximum ground-displacement amplitudes  $A_{\max}$  and related periods for body- and surface-wave groups but for the newly proposed IASPEI broadband magnitude standards  $mB\_BB$  and  $Ms\_BB$  on direct measurement of the maximum ground motion velocity amplitudes  $V_{\max}$  (Chapter 3, sections 3.2.4 and 3.2.5 and IS 3.3). For symmetric oscillations and as general IASPEI measurement standard amplitudes should be given as half peak-to-trough (double) amplitudes. Some computer programs mark the record cycle from which the maximum amplitude  $A_{\max}$  and the related  $T$  have been measured (Figures 3 and 4 in EX 3.1).

However, some algorithms for automatic amplitude measurement prefer maximum zero-to-peak (or trough) measurements instead. In the case of strongly asymmetric phase wavelets the difference may be significant (Fig. 3.19b) but remains always less than 0.3 magnitude units. The related periods were in analog records with limited time resolution traditionally measured as the time between the largest neighboring peaks or troughs, respectively (Fig. 3.19). However, digital data with high time resolution and the use of precise vertical line cursors for time picking in modern analysis tools make it preferable to measure as period twice the time difference between the adjacent maximum peak and maximum trough amplitude, as in Fig. 11.4. This is now the IASPEI recommended measurement standard for

periods (see IASPEI 2013). However, some automatic digital procedures measure zero-to-peak (or trough) amplitude instead and associate with it as period the double time difference between the two adjacent zero crossings related to the maximum positive or negative swing. For a comparison of  $\log(A/T)_{\max}$  and  $\log(V_{\max}/2\pi)$  measurements according to IASPEI standards or to modified automatic procedures see Chapter 3, section 3.2.3.2 and the discussion therein on possible discrepancies.

Note that the measured maximum trace amplitudes in a seismic record have to be corrected for the frequency-dependent magnification of the seismograph to find the “true” ground-motion amplitude at the given period. According to the new IASPEI standards amplitudes should nowadays be measured and reported in nanometers ( $1 \text{ nm} = 10^{-9} \text{ m}$ ), in contrast to the traditional units of micrometers ( $1 \mu\text{m} = 10^{-6} \text{ m}$ ), as in the original amplitude-based magnitude formulas defined in analog days when the seismographs recorded with relatively low magnification. Fig. 3.20 shows a few typical displacement amplification curves of standard analog seismographs used with paper or film records. For digital seismographs, instead of displacement or velocity magnification, the frequency dependent *nominal resolution* is usually given in units of nm/counts, or in  $\text{nm s}^{-1}/\text{count}$  for ground velocity measurements.

Both record amplitudes and related dominating periods do not only depend on the spectrum of the arriving waves but are mainly modulated by the shape, peak magnification and bandwidth of the seismograph or record filter response (e.g., Fig. 4.14). Moreover, the magnifications given in the seismograph response curves are strictly valid only for steady-state harmonic oscillations without any *transient response* (Chapter 4, section 4.5.2). The latter, however, might be significant when narrow-band seismographs record short wavelets of body waves. Signal shape, amplitudes and signal duration are then heavily distorted (Figs. 4.9-4.14 and 4.16-4.19). Therefore, we have written “true” ground motion in quotation marks. Scherbaum (2001 and 2007) gives a detailed discussion of signal distortion which is not taken into account in standard magnitude determinations from band-limited records. However, signal distortion must be corrected for in more advanced digital signal analysis for source parameter estimation by taking into account the full *transfer function* which comprises both the amplitude and the phase response of the seismograph (Chapter 5 and IS 5.2). The distortions are largest for the very first oscillation(s) and they are stronger and longer lasting the narrower the recording bandwidth (Chapter 4, sections 4.5.1 and 4.5.2). Transient response decays with time, depending also on the damping of the seismometer. It is usually negligible for amplitude measurements on dispersed teleseismic surface wave trains.

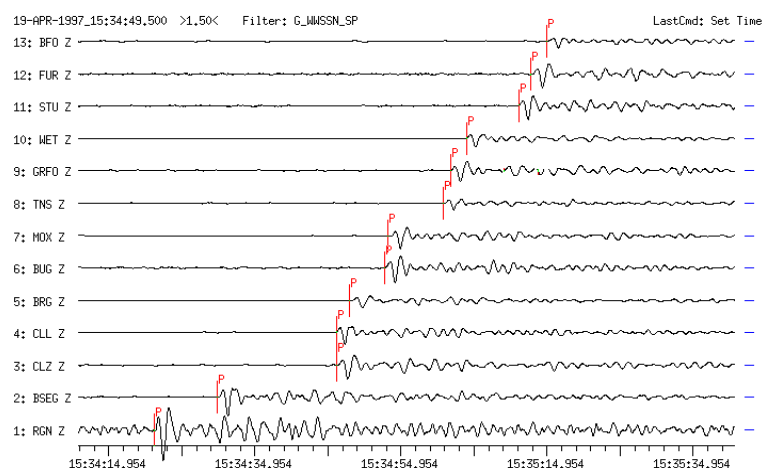
To calculate real ground motion amplitudes from recorded amplitudes, the frequency-dependent seismometer phase and amplitude response as well as the instrument gain, have to be known from careful calibration (Chapter 5, section 5.6 and IS 5.2). Analog seismograms should be clearly annotated and relate each record to a seismometer with known amplitude response. For digital data, the instrument response is usually included in the header information of each seismogram file or has to be inserted, e.g. in the case of SEED data, by the user or to be given in a separate file that is automatically linked when analyzing data files. As soon as amplitudes and associated periods are picked in digital records, most software tools for seismogram analysis calculate instantaneously the ground displacement or ground velocity amplitudes and write them in related parameter files.



### 11.2.2.3 First-motion polarity

Another parameter which has to be determined (if the SNR permits) and reported routinely is the polarity of the P-wave first motion in vertical component records. Reliable observations of the first motion polarity at stations surrounding the seismic source in different directions allow deriving seismic fault-plane solutions (Chapter 3, section 3.4 and EX 3.2). The wiring of seismometer components has to be carefully checked to assure that compressional (c) first arrivals appear on vertical-component records as an upward motion (+ or U) while dilatational (d) first arrivals are recorded as a downward first half-cycle (- or D). The conventions for horizontal component recordings are + (U) for first motions towards N and E, and – (D) for motions towards S and W, respectively. These need to be taken into account when determining the *backazimuth* of the seismic source from amplitude and polarity readings on 3-component records (see EX 11.2, Figure 1). However, horizontal component polarities are usually not considered in polarity-based fault-plane solutions and therefore not routinely reported to data centers. Fig. 11.4 shows a compressional first arrival.

One should be aware, however, that narrow-band signal filtering may reduce the first-motion amplitude by such a degree that its polarity may no longer be reliably recognized or may even get lost completely in the noise (Figs. 4.9 - 4.11). This may result in wrong polarity reporting and hence erroneous fault-plane solutions. Since short-period (SP) records usually have a narrower bandwidth than medium- to long-period or even broadband records, one should differentiate between first-motion polarity readings from SP and LP/BB records. Also, long-period waves integrate over much of the detailed rupture process and so should show more clearly the overall direction of motion which may not be the same as the first-motion arrival in SP records which may be very small. Therefore, according to recommendations in 1985 of the WG on Telegrafic Formats of the IASPEI Commission on Practice, polarities reported to international data centers should unambiguously differentiate between readings on SP (c and d) and those on LP and BB records (u for “up” = compression and r for “rarefaction” = dilatation). Although event-size and SNR-dependent, polarity readings on BB records tend to be more reliable (Chapter 4, Figs. 4.10 and 4.11). Fig. 11.5 shows short-period P-wave onsets from a Severnaya Zemlya event of April 19, 1997, as recorded in WWSSN\_SP filtered broadband stations of the German Regional Seismic Network (GRSN). The P-wave amplitudes vary significantly within the network, the first-motion polarity remains the same.



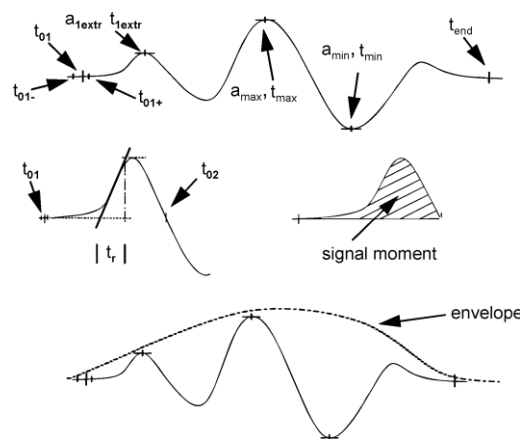
**Fig. 11.5** WWSSN\_SP filtered vertical-component broadband records at GRSN stations for the same event as in Fig. 11.4.



### 11.2.3 Advanced wavelet parameter reporting from digital records

The parameters discussed in 11.2.2 have been routinely reported over the decades of analog recording. Digital records, however, allow versatile signal processing so that additional wavelet parameters can be measured routinely. Such parameters may provide a much deeper insight into the seismic source processes and the seismic moment release. Not only can onset times be picked but their range of uncertainty can also be marked. Further, for a given wave group, several amplitudes and related times may be quickly measured and these allow inferences to be drawn on how the rupture process may have developed in space and time. Moreover, the duration of a true ground displacement pulse  $t_w$  and the rise time  $t_r$  to its maximum amplitude contain information about size of the source, the stress drop and the attenuation of the pulse while propagating through the Earth. Integrating over the area underneath a displacement pulse allows to determine its signal moment  $m_s$  which is, depending on the bandwidth and corner period of the recording, related to the seismic moment  $M_0$  (Seidl and Hellweg, 1988). Finally, inferences on the attenuation and scattering properties along the wave path can be drawn from the analysis of wavelet envelopes.

Fig. 11.6 depicts various parameters in relation to different seismic waveforms. One has to be aware, however, that each of these parameters can be severely affected by the properties of the seismic recording system (Fig. 4.17 and Scherbaum 2001). Additionally, one may analyze the SNR and report it as a quantitative parameter for characterizing signal strength and thus of the reliability of phase and parameter readings. This is routinely done when producing the Reviewed Event Bulletin (REB) of the International Data Centre (IDC) in the framework of the CTBTO (Chapter 15). The SNR may be either given as the ratio between the maximum amplitude of a considered seismic phase to that of the preceding ambient noise or signal-generated coda without taking differences in the frequencies into account, by the respective STA/LTA or rms ratio (see IS 8.1) or by determining the spectral SNR (Fig. 11.60). Regrettably, no standard procedure is yet available for best assessment of the quality of phase amplitude and onset-time readings (see also IS 11.4 and section 11.6).



**Fig. 11.6** Complementary signal parameters such as multiple wavelet amplitudes and related times, rise-time  $t_r$  of the displacement pulse, signal moment  $m_s$  and wavelet envelope (with modification from Scherbaum, Of Poles and Zeros, Fig. 1.9, p. 10, © 2001; with permission of Kluwer Academic Publishers).

Although these complementary signal parameters could be determined rather easily and quickly by using appropriate software for signal processing and seismogram analysis, their measurement and reporting to data centers is not yet common practice. It is expected, however, that the recently introduced more flexible formats such as the IASPEI Seismic Format for parameter reporting and storage (see ISF, Chapter 10.2.5), or QuakeML (section 2 in IS 3.2), in conjunction with e-mail and internet data transfer, will pave the way for their routine reporting.

## 11.2.4 Criteria to be used for phase identification

### 11.2.4.1 Travel time and slowness

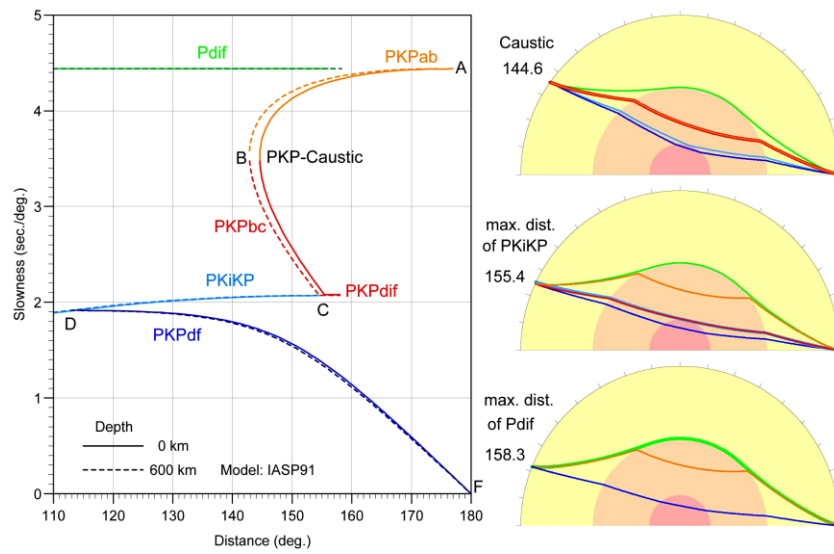
As outlined in Chapter 2, travel times of identified seismic waves are not only the key information for event location but also for the identification of seismic wave arrivals and the determination of the velocity structure of the Earth along the paths which these waves have traveled. The same applies to the horizontal component  $s_x$  of the slowness vector  $s$ . The following relations hold:

$$s_x = dt/dD = p = 1/v_{app}$$

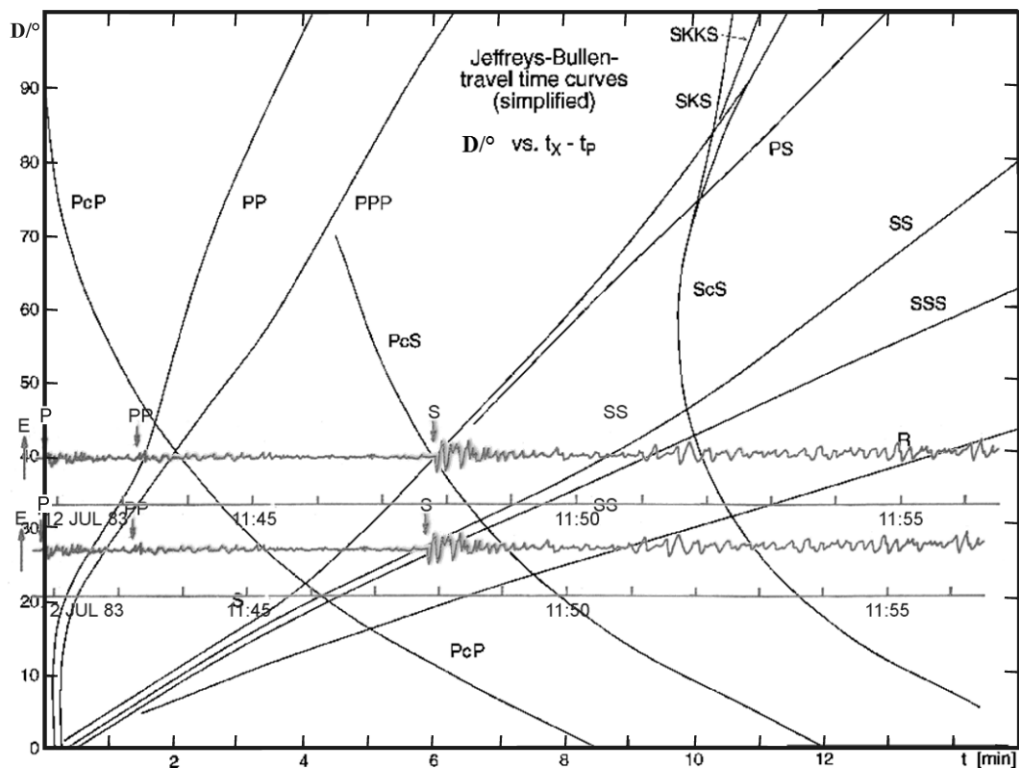
where  $v_{app}$  is the apparent horizontal velocity of wave propagation,  $dt/dD$  the gradient of the travel-time curve  $t(D)$  in the point of observation at distance  $D$ , and  $p$  is the ray parameter.  $v_{app} = \infty$  for vertical incidence at the surface ( $i = 0^\circ$ ) and  $v_{app} = v =$  true wave propagation velocity for  $i = 90^\circ$ , i.e., when the seismic ray runs along the Earth's surface. The animation linked to section 2.1 of IS 1.1 (New Zealand earthquake 2001) is an excellent example for the slowness differences of the different seismic phases traveling through the GRSN.

Due to the given structure of the Earth, the absolute as well as relative travel-time and slowness differences between various types of seismic waves vary with distance in a systematic way. Therefore, differential travel-time curves with respect to the first P-wave arrival (Figure 4 in EX 11.2) or absolute travel-time curves with respect to known origin time OT (Figs. 2.48 and 2.60) or known slowness-distance relationships (e.g., Fig. 11.7) are very suitable for identifying seismic waves in seismic records. This may be done by matching as many of the recognizable wave onsets in the record as possible with theoretical travel-time or slowness curves for various theoretically expected phases at epicentral distance  $D$ , depending also on source depth  $h$ .

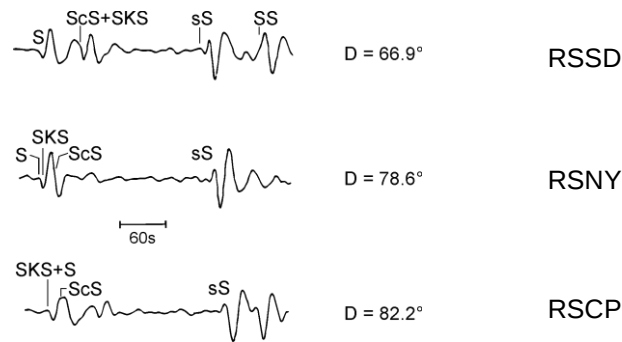
When using analog records with fixed time resolution one should make sure that the plotted  $t(D)$ -curves have the same time-resolution as your record and investigate the match between the theoretical phase travel-times and phase onsets on the records at different distances. Relative travel-time curves thus allow not only the identification of best matching phases but also the distance of the station from the epicenter of the source to be estimated (Fig. 11.8). Note, however, that in certain distance ranges the travel-time curves of different types of seismic waves (Figure 4 in EX 11.2) are close to each other, or even overlap, for example for PP and PcP between about  $40^\circ$  and  $50^\circ$  (see also Figure 6a in DS 11.2) and for S, SKS and ScS between  $75^\circ$  and  $90^\circ$  (Fig. 11.9). Proper phase identification then requires to take besides travel-time differences also additional criteria such as slowness, amplitude, polarity and dominating period into account (sections 11.2.4.2 to 11.2.4.4). The most probable distance should agree best also with those complementary criteria. If absolute travel-time curves or tables are available also the origin time can be estimated (EX 11.1 and EX 11.2).



**Fig. 11.7** The distance-variable horizontal component ray parameter (slowness) of different seismic core phases and the constant slowness of the Pdif phase that is diffracted around the core-mantle boundary (CMB).



**Fig. 11.8** Determination of epicentral distance by searching for the best fit when moving a seismic record from below upward over a differential travel-time curve until one reaches the best possible match of the dominating wave onsets with the travel-time branches of major seismic phases. In the given case the best match was achieved at about 40° distance (1° = 111.2 km).



**Fig. 11.9** Example of long-period horizontal component seismogram sections from a deep-focus earthquake in the Sea of Okhotsk (20.04.1984, mb = 5.9, h = 588 km), recorded at the stations RSSD, RSNY, and RSCP, respectively, in the critical distance range of overlapping travel-time branches of S, SKS and ScS. Because of the large focal depth the depth phase sS is clearly separated in time. Modified cut-out of Plate 40 in Kulhánek (1990), *Anatomy of Seismograms*, p.137-138; © 1990; with permission from Elsevier Science.

Note that the travel-time curves shown in Figs. 11.8 and 2.60 or those given in EX 11.1 and EX 11.2 are valid for near-surface sources only. Both absolute and (to a lesser extent) relative travel times between different phases change with source depth (see IASPEI 1991 Seismological Tables, Kennett, 1991) and, in addition, also depth phases may appear, as in Fig. 11.9 (see also section 2.6.3 in Chapter 2 and Table 1 in EX 11.2). Note also that far-regional travel-time curves ( $D > (17-20^\circ)$ ) vary little from region to region. Typically, the theoretical travel times of the main phases deviate less than 2 s from those observed (compare Figs. 2.57-2.60 in Chapter 2). In contrast, local/regional travel-time curves for crustal and uppermost mantle phases may differ much more from region to region. This is due to the pronounced lateral variations of crust and lithosphere thickness, structure, composition, age and related propagation velocities in continental and oceanic areas. For variations in crustal thickness see Fig. 2.11 in Chapter 2. Therefore, local/regional travel-time curves should be derived for each region in order to improve phase identification and estimates of source distance and depth and thus event location.

Nowadays, rapid epicenter and/or source depth estimates are normally already available at the data center prior to detailed record analysis. Then seismogram analysis software (section 11.4) such as SEISAN (Ottemöller, Voss and Havskov, 2013; Havskov and Ottemöller, 1999; <http://www.ifj.f.uib.no/seismo/software/seisan/seisan.html>), Seismic Handler (SHM) (author: Klaus Stammer, recent developments: Marcus Walther; <http://www.seismic-handler.org>) or SeisComp3 (<http://www.seiscomp3.org/>) allow the theoretically expected travel times for all main seismic phases to be marked on the record. This eases phase identification. An example is shown in Fig. 11.17 for a record analyzed with Seismic Handler. One should be aware, however, that arrival times calculated via ray theory and the appearance of the real wave onsets may differ. Therefore, onset-times of phases, calculated on the basis of a homogeneous 1D global average Earth model via ray theory can only guide the phase identification but should not be used for onset picking! A striking example of possible differences in the order of a few seconds between theoretically expected and really observed phase onsets is shown in Figure 9b of DS 11.3. Realizing this discrepancy led to the improvement of the velocity model for the Earth core in IASP91 (Kennett and Engdahl, 1991) which was taken into account in the AK135 model (Kennett et al., 1995).

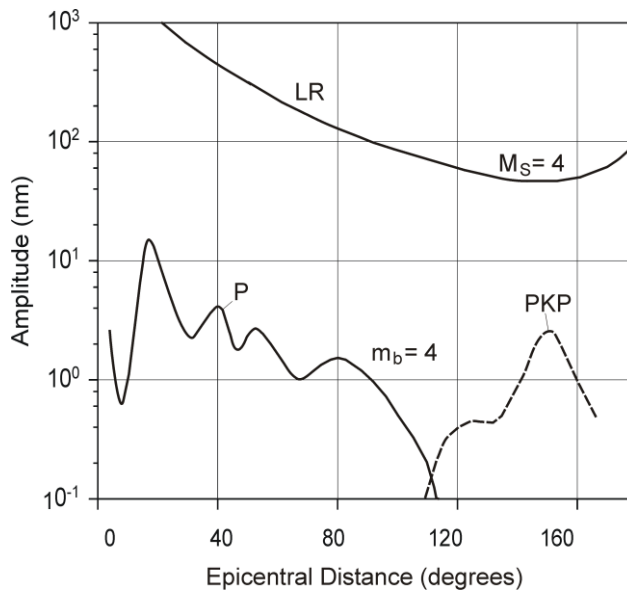
Be aware that one of the major challenges for modern global seismology is 3-D tomography of the Earth. We need to know better the location and size of deviations in wave velocity with respect to the global 1-D reference model. Only then will material flows in the mantle and core (which drive plate tectonics, the generation of the Earth's magnetic field and other processes) be better understood. Station analysts should never trust the computer generated theoretical onset times more than the ones that they can recognize in the record itself. For Hilbert transformed phases (see 2.5.4.3) onset times are best read after filtering to correct for the transformation. Without unbiased analyst readings we will never be able to derive improved models of the inhomogeneous Earth. Moreover, the first rapid epicenters, depths and origin-times published by the data centers are only preliminary estimates and are usually based on first arrivals only. Their improvement, especially with respect to source depth, requires more reliable onset-time picks, and the identification of secondary (later) arrivals (Figure 7 in IS 11.1).

At a local array or regional seismic network center both the task of phase identification and of source location of teleseismic events is easier than at a single station because local or regional slowness can be measured from the time differences of the respective wave arrivals at the various stations (Chapter 9 and sections 11.3.4 and 11.3.5 below). But even then, determining D from travel-time differences between P or PKP and later arrivals can significantly improve the location accuracy. This is best done by using three-component broadband recordings from at least one station in the array or network center. The reason is that travel-time differences between first and later arrivals vary more rapidly with distance than the slowness of first arrivals. But large arrays and regional networks usually give better control of the backazimuth of the source than 3-component recordings (compare 11.2.6 and 11.2.7), especially for low-magnitude events recorded at low SNR and thus uncertain first-motion polarity readings.

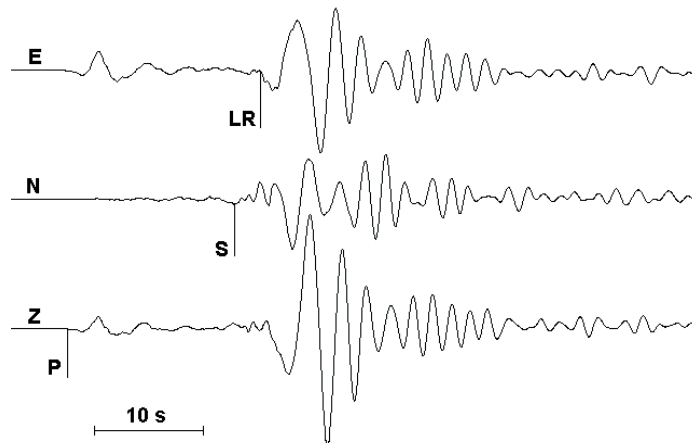
#### **11.2.4.2 Amplitude-distance relations, dominating periods and waveforms**

Amplitudes of seismic waves vary with distance due to geometric spreading, focusing and defocusing caused by variations of wave speed and attenuation in the Earth. To correctly identify body-wave phases one has first to be able to differentiate between body- and surface-wave groups and then estimate at least roughly, whether the source is at shallow, intermediate or rather large depth. At large distance, surface waves are only seen on LP and BB seismograms. Because of their 2D propagation, geometrical spreading for surface waves is less than for body waves that propagate 3-D. Also, because of their usually longer wavelength, surface waves are less attenuated and less affected by small-scale structural inhomogeneities than body waves. Therefore, on records of shallow seismic events, surface-wave amplitudes dominate over body-wave amplitudes and show less variability with distance (Figs. 11.10 to 11.12). This is also obvious when comparing the magnitude calibration functions for body and surface waves (Figures and tables in DS 3.1).

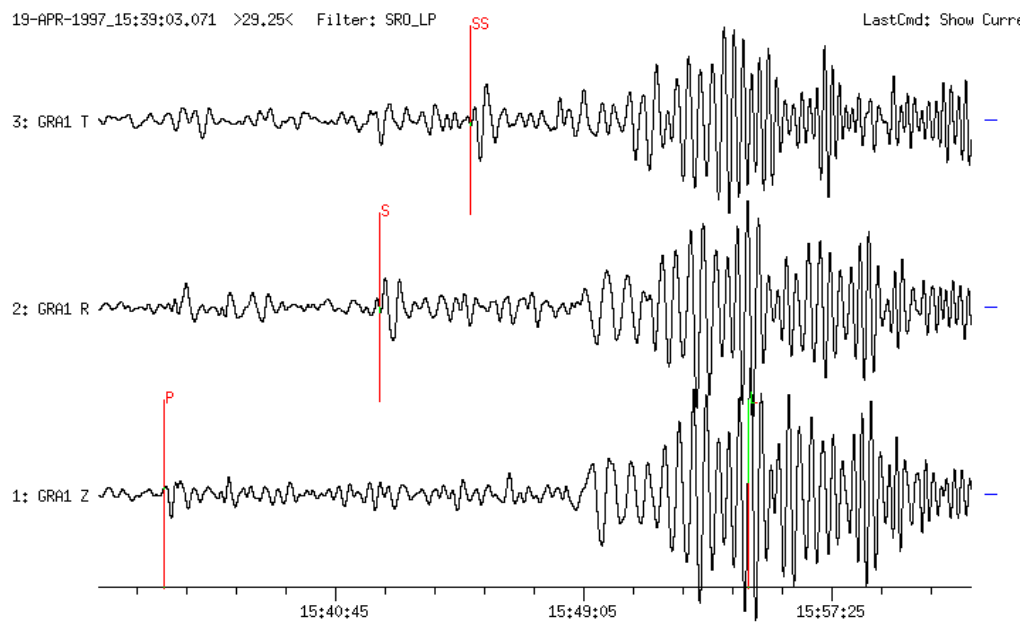
However, as source depth increases, surface-wave amplitudes decrease relative to those of body waves. The decrease is the larger the shorter the wavelengths. Thus, the surface waves from earthquakes at depth > about 70-100 km may have amplitudes smaller than those of body waves or may not even be detected at all on seismic records (Figure 2 in EX 11.2). This should alert seismogram analysts to look for depth phases, which are then usually well separated from their primary waves and more easily recognizable (Fig. 11.9 above and Figures 6a and b in DS 11.2).



**Fig. 11.10** Approximate smoothed amplitude-distance functions for P and PKP body waves (at about 1 Hz) and of long-period Rayleigh surface waves (LR,  $T \approx 20$  s) for an event of magnitude 4.



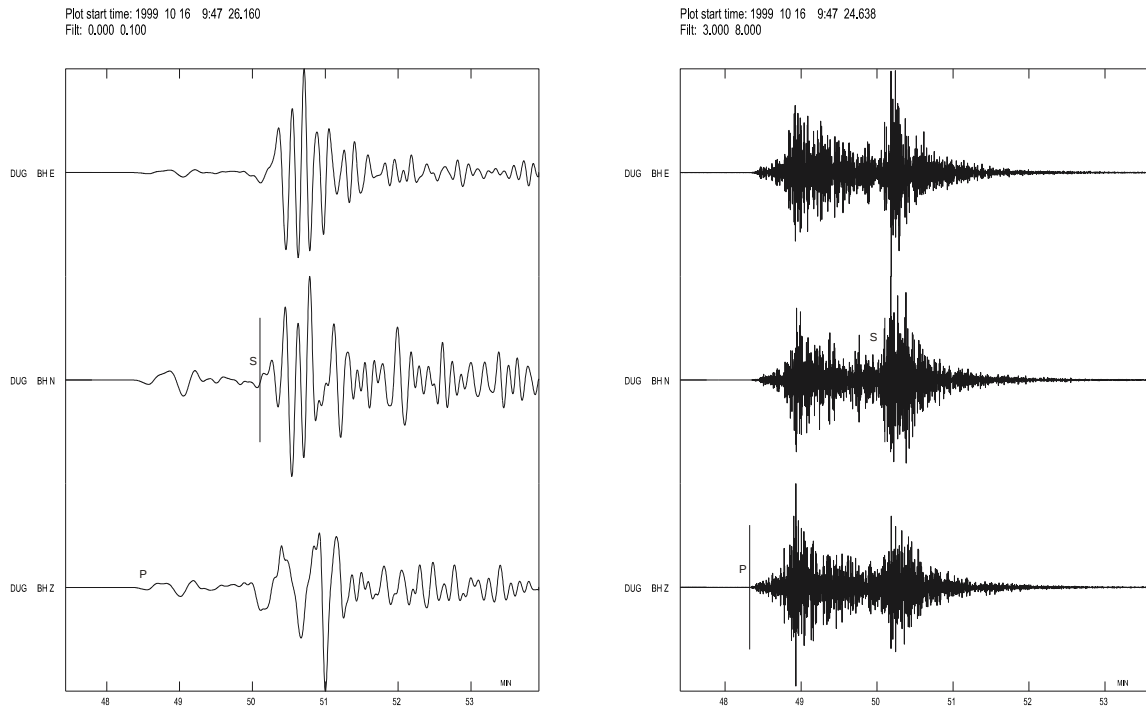
**Fig. 11.11** Three-component BB-velocity record at station MOX of a mine collapse in Germany; (13 March 1989;  $M_l = 5.5$ ) at a distance of 112 km and with a backazimuth of  $273^\circ$ . Note the Rayleigh surface-wave arrival LR with subsequent normal dispersion.



**Fig. 11.12** T-R-Z rotated three-component seismogram (SRO-LP filter) from an earthquake east of Severnaya Zemlya (19 April 1997,  $D = 46.4^\circ$ ,  $m_b = 5.8$ ,  $M_s = 5.0$ ). The record shows P, S, SS and strong Rayleigh surface waves with clear normal dispersion. The surface wave maximum has periods of about 20 s. It is called an Airy-phase and corresponds to a minimum in the dispersion curve for continental Rayleigh waves (Fig. 2.10).

Another feature that helps in phase identification is the waveform. Most striking is the difference in waveforms between body and surface waves. Dispersion in surface waves results in long wave trains of slowly increasing and then decreasing amplitudes, whereas non-dispersive body waves form short duration wavelets. Usually, the more long-period surface waves arrive first (“normal” or “positive” dispersion) (as in Figs. 11.11 and 11.12). However, very long-period waves ( $T > 60$  s to about 200 s), that penetrate into the upper mantle down to the asthenosphere (a zone of low wave speeds), may show inverse dispersion. The longest waves then arrive later in the wave train (Figs. 2.20 and 11.15).

For an earthquake of a given seismic moment, the maximum amplitude of the S wave is about five-times larger at source than that of the P waves (Fig. 11.13 left and Figs. 2.4, 2.25, and 2.50). This is a consequence of the different propagation velocities of P and S waves [see Eq. (3.2) in Chapter 3]. Also the spectrum is different for each wave type. Thus, P-wave source spectra have corner frequencies about  $\sqrt{3}$  times higher than those of S. In high-frequency filtered records this may increase P-wave amplitudes with respect to S-wave amplitudes (Fig. 11.13 right). Additionally, the frequency-dependent attenuation of S waves is significantly larger than for P waves (Chapter 2, section 2.5.4.2). Due to both effects, S waves and their multiple reflections and conversions are – within the teleseismic distance range – mainly observed on LP or BB records. On the other hand, the different P-wave phases, such as P, PcP, PKP, and PKKP, are well recorded, up to the largest epicentral distances, by SP seismographs with maximum magnification typically around 1 Hz.



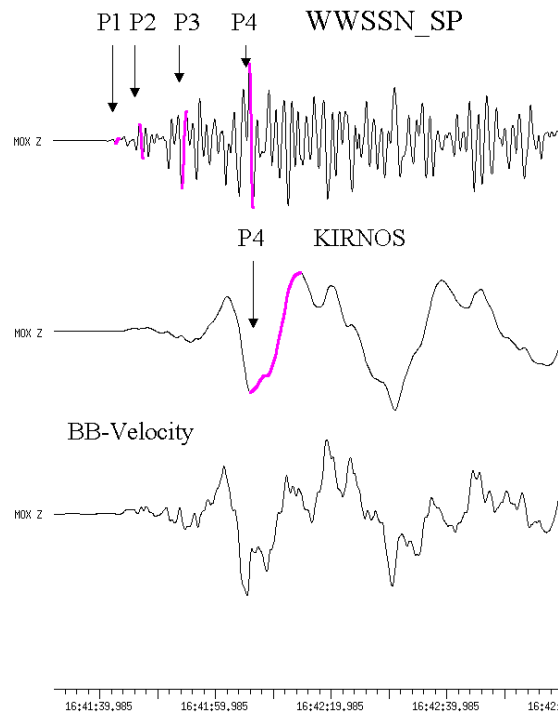
**Fig. 11.13** Left: Low-pass filtered ( $< 0.1$  Hz) and right: band-pass filtered (3.0-8.0 Hz) seismograms of the Oct. 16, 1999, earthquake in California ( $m_b = 6.6$ ,  $M_s = 7.9$ ) as recorded at the broadband station DUG at  $D = 6^\circ$  (courtesy of L. Ottemöller).

Generally, the rupture duration of earthquakes is much longer than the source process of explosions (milliseconds). It ranges from less than a second for small micro-earthquakes up to several minutes for the largest earthquakes with a source which is usually a complex multiple rupture process ( e.g., Figs. 11.14, 3.9 and 3.23 and the animations in sections 2.6 of IS 1.1).

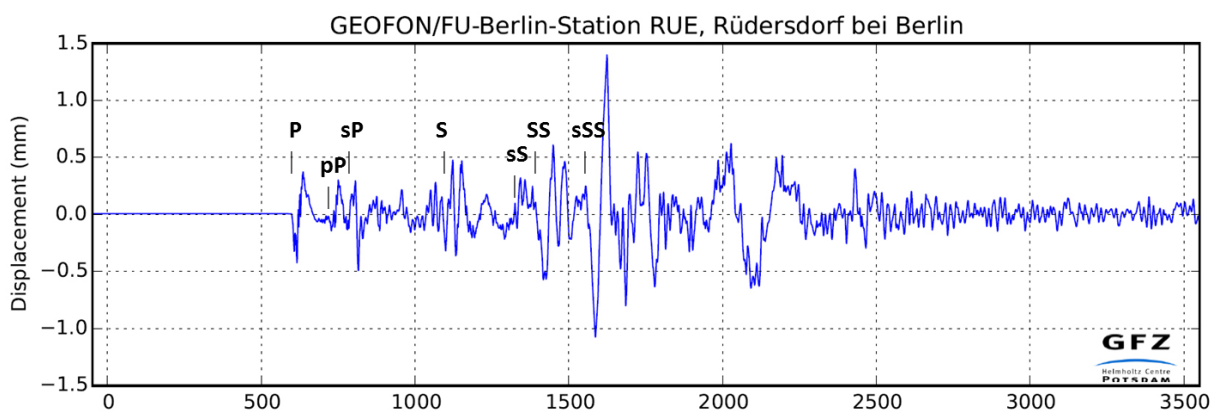
As compared to shallow crustal earthquakes, deep earthquakes of comparable magnitude are often associated with higher stress drop and smaller source dimension. This results in the strong excitation of higher frequencies and thus more simple and impulse-like waveforms (e.g. Figs. 11.15, 2.64, 4.14 and Figures 6a and b in DS 11.2). Therefore, S waves from deep earthquakes may be recognizable even in short-period records at teleseismic distances.

The waveforms of first arriving P waves from underground nuclear explosions (UNE) are in many cases rather impulsive too but associated with comparably small S waves. As compared to shallow earthquakes and when scaled to the same magnitude, the source dimension of an UNE is usually much smaller, their source process simpler and their source duration much shorter (typically in the range of milliseconds). Accordingly, explosions generate significantly more high-frequency energy than earthquakes and usually produce shorter and simpler waveforms of the P first arrivals. However, since their source depths is rather small, typically ranging between several hundred to less than 2000 m, these high frequency waves are prone to scattering and multi-pathing at upper crustal heterogeneities and the near-surface topography, thus producing strong signal-generated coda in local and regional records. Examples of regionally and (less complicated) teleseismically recorded waveforms of underground nuclear explosions are given in Fig. 2.17 and Figures 1 to 5 of DS 11.4, respectively.





**Fig. 11.14** Vertical component records of the P-wave group from a crustal earthquake in Sumatra (04 June 2000;  $m_b = 6.8$ ,  $M_s = 8.0$ ) at the GRSN station MOX at  $D = 93.8^\circ$ . **Top:** WWSSN-SP (type A); **middle:** medium-period Kirnos SKD BB-displacement record (type C), and **bottom:** original BB-velocity record. Clearly recognizable is the multiple rupture process with  $P_4 = P_{\max}$  arriving 25 s after the first arrival P1. The short-period magnitude  $m_b$  determined from P1 would be only 5.4,  $m_b = 6.3$  from P2 and  $m_b = 6.9$  when calculated from P4. When determining the medium-period broadband body-wave magnitude from P4 on the Kirnos record then one gets  $m_B = 7.4$ .



**Fig. 11.15** Very broadband vertical component record of the great  $M_w = 8.2$  deep earthquake ( $h = 609$  km) on 24 May 2013 underneath the Okhotsk Sea at station RUE ( $D = 67.9^\circ$ ), Germany. Note the rather impulsive broadband waveforms of the primary P and S waves and their related depth phases as well as the inverse dispersion of the very long-period mantle surface waves following sSS.

Note, however, that production explosions in large quarries or open cast mines, with yields ranging from several hundred to more than one kiloton TNT, are usually fired in sequences of time-delayed sub-explosions that are spread out over a larger area. Such explosions may generate rather complex wave fields, waveforms and unusual spectra, sometimes further complicated by the local geology and topography, and thus not easy to discriminate from local earthquakes. For discrimination criteria see section 11.2.5.2.

At some particular distances, body waves may have relatively large amplitudes (Fig. 11.10), especially near *caustics* (Figs. 2.30 to 2.33), e.g. for P waves in the distance range between  $15^\circ$  and  $30^\circ$  or around  $D = 145^\circ$  for PKP phases (Figs. 11.10 and 11.72). Due to the caustic around  $145^\circ$  PKP amplitudes are about as large as P-wave amplitudes around  $50^\circ$  epicentral distance (Fig. 11.10). The double triplication of the P-wave travel-time curve between  $15^\circ$  and  $30^\circ$  results in closely spaced successive onsets and consequently rather complex waveforms (Fig. 11.63). In contrast, amplitudes decay rapidly in *shadow zones* due to low-velocity layers (Fig. 2.33). In the teleseismic range this applies to direct P for distances beyond  $100^\circ$  in the so-called *core shadow* due to the drastic decay in P-wave velocity in the outer core as compared to the lower mantle (Fig. 2.79). P is then recorded only as diffracted wave Pdif (Figs. 11.7, 11.10, 2.55, 3.34 and example 2 in DS 11.3). At distances between about  $30^\circ$  and  $100^\circ$ , however, long-period waveforms of P and later phases may look rather simple and well separated (Figs. 11.12, 11.17, 11.20, 11.48, 11.52, and 11.63 as well as Figures 8a, 9c, 10c and 11b in DS 11.2). Beyond the PKP caustic, between  $145^\circ < D < 160^\circ$ , longitudinal core phases split into three travel-time branches with typical amplitude-distance patterns. This, together with their systematic relative travel-time differences, permits rather reliable phase identification and distance estimates, often better than  $1^\circ$  (Figs. 11.73 and 11.74 as well as exercise EX 11.3).

Fig. 11.16 is a simplified diagram showing the relative frequency of later body-wave arrivals with respect to the first P-wave arrival or the number  $n$  of analyzed earthquakes, as a function of epicentral distance  $D$  between  $36^\circ$  and  $166^\circ$ . They are based on observations in standard records (Fig. 3.20 left) of types A4 (SP -  $T < 1.5$  s), B3 (LP -  $T$  between 20 s and 80 s) and C (BB displacement for  $T = 0.1$  s - 20 s) at station MOX, Germany (Bormann, 1972a).

These diagrams show that in the teleseismic distance range one can mainly expect to observe in SP records the following longitudinal phases: P, PcP, ScP, PP, PKP (of branches ab, bc and df), P'P' (= PKPPKP), PKKP, PcPPKP, SKP and the depth phases of P, PP and PKP. In LP and BB records, however, additionally S, ScS, SS, SSS, SKS, SKSP, SKKS, SKKP, SKKKS, PS, PPS, SSP and their depth phases are frequently recorded. This early finding based on the visual analysis of traditional analog film recordings has been confirmed by stacking SP and LP filtered broadband records of the Global Digital Seismic Network (GDSN) (Astiz et al., 1996; Figs. 2.57 to 2.60) and most recently by a refined automatic near real-time envelop stacking procedure developed [at the IRIS Data Management Center \(DMC\) \(Trabant et al., 2012\)](#) (see Fig.11.83).

Since these diagrams or stacked seismogram sections reflect, in a condensed form, some systematic differences in waveforms, amplitudes, dominating periods and relative frequency of occurrence of seismic waves in different distance ranges, they may, when used in addition to travel-time curves and polarization criteria, give some guidance to seismogram analysts as to what kind of phases they may expect at which epicentral distances and in what kind of seismic records or band-pass filter range. Note, however, that the appearance of these phases is not “obligatory” in records of any station. Rather, it may vary from region to region,

depending also on the source mechanisms and the radiation pattern with respect to the recording station, the source depth, the area of reflection (e.g., underneath oceans, continental shield regions, young mountain ranges), and the distance of the given station from zones with frequent deep earthquakes. Therefore, no rigid and globally valid rules for phase identification can be given.

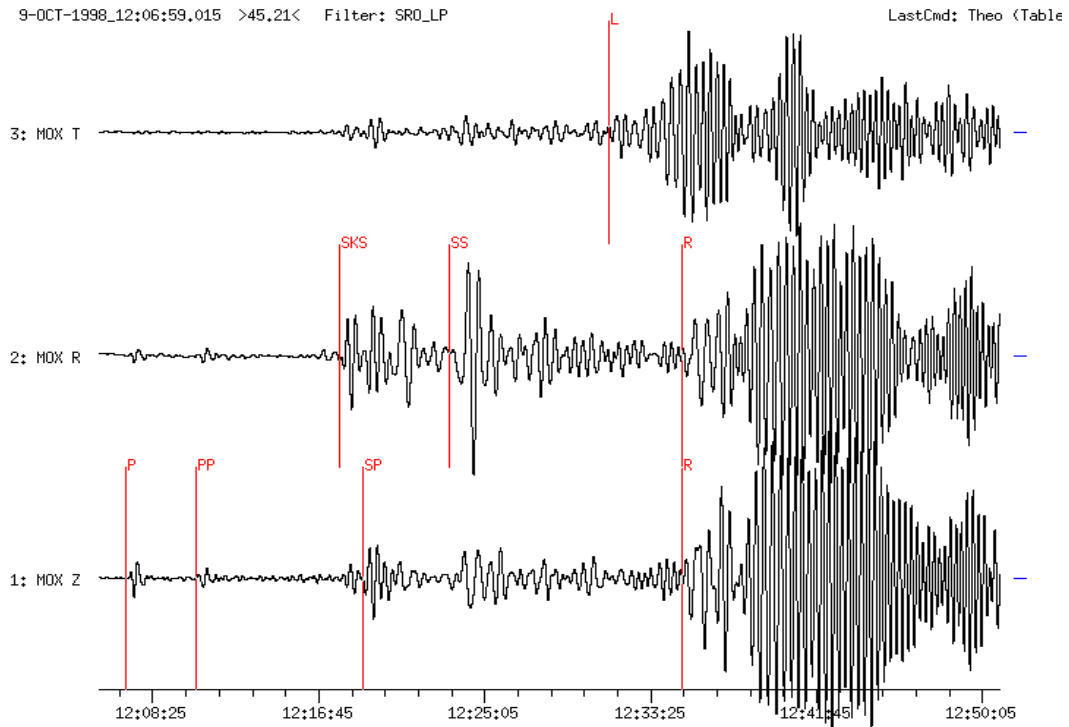
**Fig. 11.16** Relative frequency of occurrence of secondary phases in standard analog records at station MOX, Germany, within the teleseismic distance range 36° to 166°. The first column relates to 100% of analyzed P-wave first arrivals or of analyzed events (hatched column), respectively. In the boxes beneath the phase columns the type of standard records is indicated in which these phases have been observed best or less frequently/clear (then record symbols in brackets). A – short-period; B – long-period LP, C – Kirnos SKD BB-displacement according to the respective standard seismograph types in the Willmore (1979) Manual of Seismological Observatory Practice.

more long-period waves with large amplitudes are generated as well which then dominate also in BB records of local events, as illustrated in Figs. 11.11 and 11.13.

### 11.2.4.3 Polarization

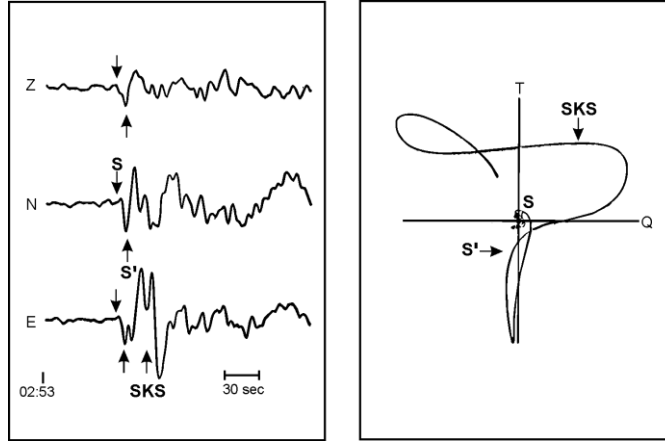
As outlined in sections 2.2 and 2.3 of Chapter 2, P and S waves are linearly polarized, with slight deviations from this ideal in the inhomogeneous and partially anisotropic real Earth (Figs. 2.7 and 2.8). In contrast, surface waves may either be linearly polarized in the horizontal plane perpendicular to the direction of wave propagation (transverse polarization; T direction; e.g., Love waves) or elliptically polarized in the vertical plane oriented in the radial (R) direction of wave propagation (Rayleigh waves; Figs. 2.9, 2.12, 2.15 and 2.16). P-wave particle motion is dominantly back and forth in the direction of wave propagation, whereas S-wave motion is perpendicular to it (Fig. 2.6). Accordingly, a P-wave motion can be split into two main components, one vertical (Z) and one horizontal (R) component (Fig. 2.7). The same applies to Rayleigh waves, but with a 90° phase shift between the Z and R components of motion. S waves, on the other hand, may show purely transverse motion, oscillating in the horizontal plane (SH; i.e., pure T component, as Love waves) or motion in the vertical propagation plane, at right angles to the ray direction (SV), or in any other combination of the two components SH and SV, depending on the source mechanisms and the related radiation pattern (Fig. 3.102). In the latter case S-wave particle motion has Z, R and T components, with SV wave split into a Z and an R component. But depending on the incidence angle and the  $v_p/v_s$  ratio at the station SV may even show elliptical polarization.

Fig. 11.17 shows the good separation of several main seismic phases on a Z-R-T-component plot. At such a large epicentral distance ( $D = 86.5^\circ$ ) the incidence angle (between the seismic ray and the vertical: Fig. 2.7) of P is small (about 15°; see EX 3.3). Therefore, the P-wave amplitude is largest on the Z component whereas for PP, which has a significantly larger incidence angle because of its shallower penetration into the mantle (Fig. 11.65), the amplitude on the R component is almost as large as on Z. For both P and PP no T component is recognizable above the noise. SKS is strong in R and has only a small T component (effect of anisotropy, Fig. 2.8). Depending on source depth and the  $v_p/v_s$  ratio the two phases PS and SP are more or less separated in time with the PS onset being SV polarized and SP as a P phase, both having a strong Z and R component. Love waves (LQ) appear as the first surface waves in T with very small amplitudes in R and Z. In contrast, Rayleigh waves (LR) are strongest in R and Z. SS in this example is also largest in R. From this one can conclude that the S waves generated by this earthquake and radiated into the direction of the recording station are almost purely of SV type. In other cases, however, it is only the difference in the R-T polarization which allows S to be distinguished from SKS (which is SV polarized; see below) in this distance range of around 80° where these two phases arrive closely to each other (Fig. 11.8 and Figure 11 in EX 11.2 where the E-W component agrees in fact with the R component because of the backazimuth of 269°).



**Fig. 11.17** Time-compressed long-period filtered three-component seismogram (SRO-LP simulation filter) of the October 9, 1998, Nicaragua earthquake recorded at station MOX ( $D = 86.5^\circ$ ). Horizontal components have been rotated (ZRT) with R (radial component) in source direction. The seismogram shows long-period phases P, PP, SKS, SP, SS and surface waves L (or LQ for Love wave) and R (or LR for Rayleigh wave).

Thus, when 3-component records are available, the particle motion of seismic waves in space can be reconstructed and used for the identification of seismic wave types (e.g., Vidale, 1986). However, usually the horizontal seismometers are oriented in geographic east (E) and north (N) direction. Then, first the backazimuth of the source has to be computed (see EX 11.2) and then the horizontal components have to be rotated into the horizontal R direction and the perpendicular T direction, respectively. This *axis rotation* is easily performed when digital 3-component data and suitable analysis software are available. It may even be carried one step further by rotating the R and Z component once more, R into the direction of the incident seismic ray (longitudinal L direction) and Z then being perpendicular to it in the Q direction of the SV component. The T component remains unchanged. Such a *ray-oriented co-ordinate system* separates and plots P, SH and SV waves in 3 different components L, T and Q, respectively (Fig. 11.18). These axes transformations are easily made given digital data from arbitrarily oriented orthogonal 3-component sensors such as the widely used triaxial sensors STS2 (Fig. 5.10 and DS 5.1). However, the principle types of polarization can often be quickly assessed even with manual measurement and elementary calculation of the backazimuth from the station to the source from analog 3-component records (see EX 11.2).



**Fig. 11.18** Ray-oriented broadband records (left: Z-N-E components; right: particle motion in the Q-T plane) of the S and SKS wave group from a Hokkaido  $M_s = 6.5$  earthquake on 21 March 1982, at station Kasperske Hory (KHC) at an epicentral distance of  $D = 78.5^\circ$ .

According to the above, all direct, reflected and refracted P waves and their multiples such as

Pg, Pn, PnPn, ...P, pP, PP, pPP, PPP, ...PcP, PKP, PKIKP, PKKP, ...PKPPKP(P'P'), ...

as well as conversions from P to S and vice versa, such as

pS, sP, SP, PS, SPP, ScP, PKS, SKP, SKPPKP, SKKP, SKS, SKKS, ...

have their dominant motion confined to, i.e., are polarized in the Z and R (or L and Q) plane. This applies to all core phases, also to SKS and its multiples, because no shear wave can travel through the liquid outer core and K stands for a longitudinal wave that has traveled through the Earth's outer core and I for a longitudinal wave that has traveled through the Earth's inner core. For ray paths of these waves see Fig. 11.65 and ray plots in IS 2.1.

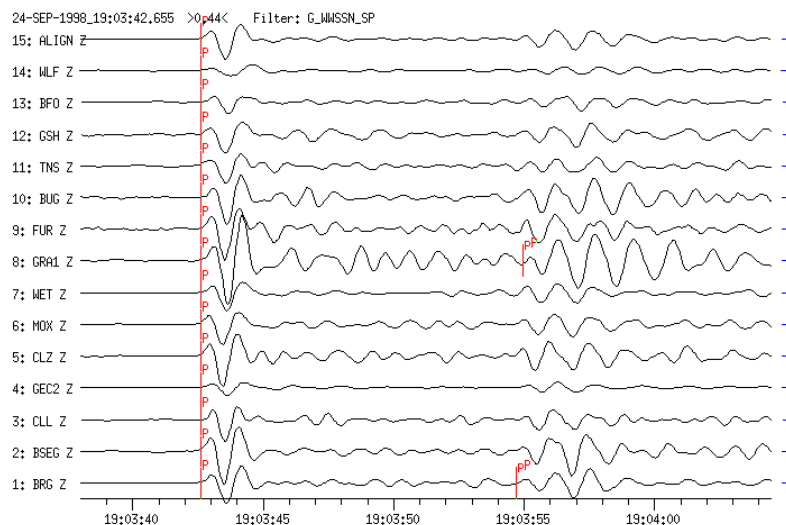
In contrast, S waves may have both SV and SH energy. However, discontinuities along the propagation path of S waves may act as selective SV/SH filters. Therefore, when an S wave arrives at the free surface, part of its SV energy may be converted into P, thus forming an SP phase. Consequently, the energy reflected as S has a larger SH component as compared to the incoming S. Thus, the more often a mixed SH/SV type of S wave is reflected at the surface, the more it becomes of SH type. Accordingly, SSS, SSSS etc. will show up most clearly or even exclusively on the T component (e.g., Fig. 11.48) unless the primary S wave is dominantly of SV-type (e.g. in Fig. 11.17). As a matter of fact, Love waves are formed through constructive interference of repeated reflections of SH waves at the free surface as well as crustal and upper mantle discontinuities. Similarly, when an S wave hits the core-mantle boundary, part of its SV energy is converted into P which is either refracted into the core with phase name letter K or reflected back into the mantle as P, thus forming the ScP phase. Consequently, multiple ScS is also usually best developed on the T component.

The empirical travel-time curves in Fig. 2.61 (Astiz et al., 1996) summarize rather well, which phases are expected to dominate the vertical, radial or transverse motion in rotated three-component records. If we supplement the use of travel-time curves with seismic recordings in different frequency bands, and take into account systematic differences in

amplitude, frequency content and polarization for P, S and surface waves, and when we know the distances, where caustics and shadow zones occur, then the identification of later seismic wave arrivals is entertaining and like a detective inquiry into the seismic record. It would be regrettable if this art dies out in the age of minimalistic automatic record analysis.

#### 11.2.4.4 Example for documenting and reporting of seismogram parameter readings

Fig. 11.19 shows a plot of the early part of a teleseismic earthquake recorded at stations of the GRSN. At all stations the first arriving P wave is clearly recognizable although the P-wave amplitudes vary strongly throughout the network. This is not a distance effect (the network aperture is less than 10% of the epicentral distance) but rather an effect of different local site conditions related to underground geology and crustal heterogeneity. As demonstrated with Figs. 4.37 and 4.38, the effect is not a constant for each station but depends both on azimuth and distance of the source. It is important to document this. Fig. 11.19 also shows for most stations a clear later arrival about 12 s after P. For the given epicentral distance, no other main phase such as PP, PPP or PcP can occur at such a time (see differential travel-time curves in Fig. 11.8). It is important to pick such later (so-called secondary) onsets which might be “depth phases” (see 11.2.5.1) as these allow a much better determination of source depth than from P-wave first arrivals alone (Figure 7 in IS 11.1).



**Fig. 11.19** WWSSN-SP filtered seismograms at 14 GRSN, GRF, GERES and GEOFON stations from an earthquake in Mongolia (24 Sept. 1998; depth (NEIC-QED) = 33 km; mb = 5.3, Ms = 5.4). Coherent traces have been time-shifted, aligned and sorted according to epicentral distance ( $D = 58.3^\circ$  to BRG,  $60.4^\circ$  to GRA1 and  $63.0^\circ$  to WLF). Note the strong variation in P-wave signal amplitudes and clear depth phases pP arriving about 12 s after P.

Tab. 11.2 gives for the Mongolia earthquake shown in Fig. 11.19 the whole set of parameter readings made at the analysis center of the former Central Seismological Observatory Gräfenberg (SZGRF) in Erlangen, Germany [now: Seismological Central Observatory (SZO) of the BGR in Hannover]:

- first line: date, event identifier, analyst;
- second and following lines: station, onset time, onset character (e or i), phase name (P, S, etc.), direction of first particle motion (c or d), analyzed component, period [s], amplitude [nm], magnitude (mb or Ms), epicentral distance [°]; and
- last two lines: source parameters as determined by the SZGRF (origin time OT, epicentre, average values of mb and Ms, source depth and name of Flinn-Engdahl-region).

Generally, these parameters are stored in a database, used for data exchange and published in lists, bulletins and made available through the Internet (IS 11.2). As an example of station parameter reports grouped according IMS1.0 with ISF1.0 extensions see IS 10.2. The onset characters i (impulsive) should be used only if the time accuracy is better than a few tens of a second, otherwise the onset will be described as e (emergent). Also, when the signal-to-noise-ratio (SNR) of onsets is small and especially, when narrow-band filters are used, the first particle motion should not be given because it might be distorted or lost in the noise. Broadband records are better suited for polarity readings (Figs. 4.10 and 4.11). Their polarities, however, should be reported as u (for “up” = compression) and r (for “rarefaction” = dilatation) so as to differentiate them from short-period polarity readings (c and d, respectively).

**Tab. 11.2** Parameter readings at the SZGRF analysis center for the Mongolia earthquake shown in Fig. 11.19 from records of the GRSN.

|            |               |          |         |   |   |      |   |           |           |     |   | ev_id 980924007 |  | KLI |
|------------|---------------|----------|---------|---|---|------|---|-----------|-----------|-----|---|-----------------|--|-----|
| 1998-09-24 |               |          |         |   |   |      |   |           |           |     |   |                 |  |     |
| BRG        | 19:03:27.2    | e P      |         | Z | T | 1.2  | A | 135.5     | mb        | 5.9 | D | 58.3            |  |     |
| ALIGN      | 19:03:27.2    | e P      |         | Z | T | 1.1  | A | 124.1     | mb        | 5.8 |   |                 |  |     |
| BSEG       | 19:03:28.2    | e P      |         | Z | T | 1.1  | A | 198.3     | mb        | 6.0 | D | 58.4            |  |     |
| CLL        | 19:03:28.6    | e P      |         | Z | T | 0.9  | A | 98.9      | mb        | 5.8 | D | 58.5            |  |     |
| GEC2       | 19:03:36.3    | i P      | c       | Z | T | 1.2  | A | 46.9      | mb        | 5.4 | D | 59.6            |  |     |
| CLZ        | 19:03:36.6    | e P      |         | Z | T | 1.1  | A | 177.7     | mb        | 6.0 | D | 59.6            |  |     |
| MOX        | 19:03:36.9    | e P      |         | Z | T | 1.2  | A | 122.4     | mb        | 5.8 | D | 59.6            |  |     |
| WET        | 19:03:38.5    | e P      |         | Z | T | 1.2  | A | 107.2     | mb        | 5.7 | D | 59.9            |  |     |
| BRG        | 19:03:39.3    | e pP     |         | Z |   |      |   |           |           |     |   |                 |  |     |
| GRA1       | 19:03:42.6    | i P      | c       | Z | T | 1.1  | A | 286.5     | mb        | 6.2 | D | 60.4            |  |     |
|            |               |          |         |   |   |      |   | b_slo 6.8 | b_az      | 54  |   |                 |  |     |
| BUG        | 19:03:48.5    | e P      |         | Z | T | 1.1  | A | 162.4     | mb        | 5.8 | D | 61.4            |  |     |
| FUR        | 19:03:48.6    | e P      |         | Z | T | 1.1  | A | 174.3     | mb        | 5.8 | D | 61.3            |  |     |
| TNS        | 19:03:49.9    | e P      |         | Z | T | 1.1  | A | 103.0     | mb        | 6.0 | D | 61.5            |  |     |
| GSH        | 19:03:54.5    | e P      |         | Z | T | 1.3  | A | 132.3     | mb        | 6.0 | D | 62.3            |  |     |
| GRA1       | 19:03:55.0    | e pP     |         | Z |   |      |   |           |           |     |   |                 |  |     |
| BFO        | 19:03:57.4    | e P      |         | Z | T | 1.1  | A | 55.2      | mb        | 5.6 | D | 62.8            |  |     |
| WLF        | 19:04:00.1    | e P      |         | Z | T | 1.7  | A | 80.3      | mb        | 5.6 | D | 63.0            |  |     |
| GRA1       | 19:11:52.2    | e S      |         | E |   |      |   |           |           |     |   | D 60.4          |  |     |
| GEC2       | 19:30:36.0    | e L      |         | Z | T | 19.9 | A | 3895.8    | MS        | 5.5 |   |                 |  |     |
| GRA1       | 19:31:03.6    | e L      |         | Z | T | 20.6 | A | 3398.7    | MS        | 5.5 |   |                 |  |     |
| SZGRF      | OT 18:53:39.3 | 45.30N   | 106.84E |   |   |      |   | mb_av 5.8 | MS_av 5.5 |     |   |                 |  |     |
| DEP        | 44km          | MONGOLIA |         |   |   |      |   |           |           |     |   |                 |  |     |

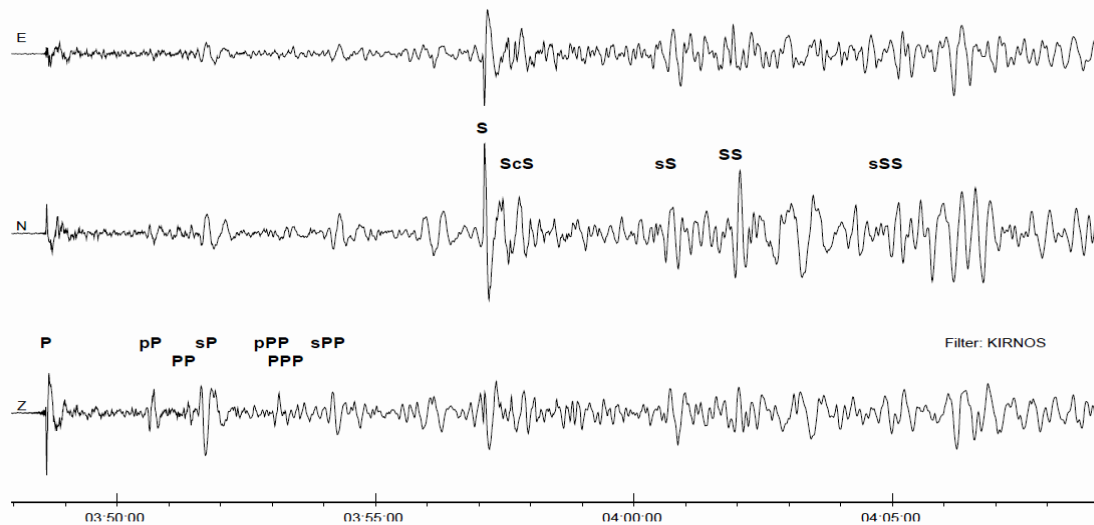
**Note:** For this event the international data center NEIC had “set” the source depth to 33 km because of the absence of reported depth phases. The depth-phase picks at the GRSN, however, with an average time difference of pP-P of about 12 s, give - in dependence of the velocity model in the source region - a focal depth of about 44 km. Also note in Tab. 11.2 the large differences in amplitudes (A) determined from the short-period records of individual stations. Accordingly, the station mb estimates vary between 5.4 and 6.2, which cannot be explained by the small differences in epicentral distance. These are mainly station site effects.



## 11.2.5 Criteria to be used in event identification and discrimination

### 11.2.5.1 Discrimination between shallow and deep earthquakes

Earthquakes are often classified on depth as: shallow focus (depth between 0 and 70 km), intermediate focus (depth between 70 and 300 km) and deep focus (depth between 300 and 700 km). However, the term "deep-focus earthquakes" is also often applied to all sub-crustal earthquakes deeper than 70 km. They are generally located in subducting slabs of the lithosphere which are subducted into the mantle. As noted above, the most obvious indication on a seismogram that a large earthquake has a deep focus is the small amplitude of the surface waves with respect to the body-wave amplitudes and the usually rather simple character of the P and S waveforms, which often have impulsive onsets (Fig. 4.15 and Fig. 11.20).

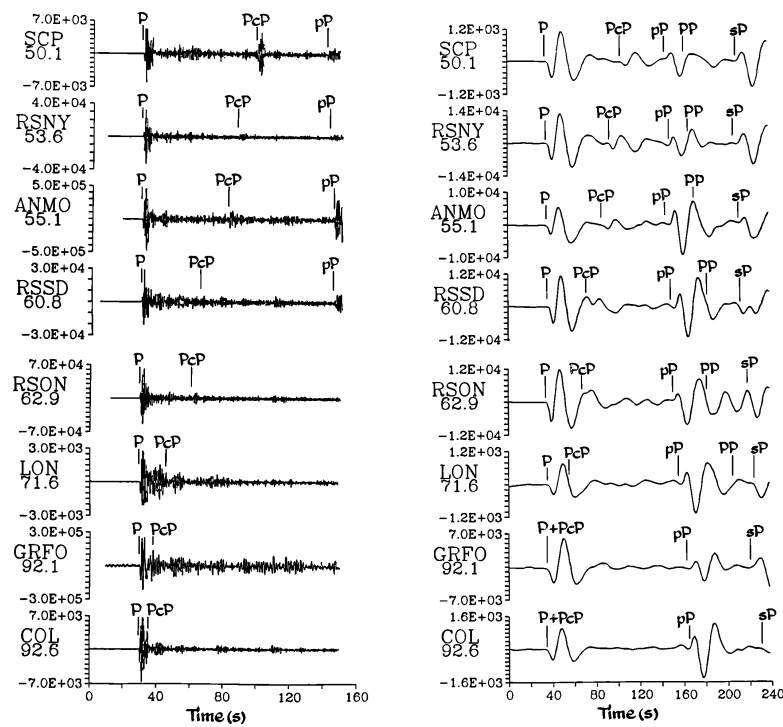


**Fig. 11.20** 3-component Kirnos SKD filtered record at station CLL ( $D = 70.6^\circ$ ) of the October 01, 2013  $M_w = 6.7$  deep earthquake ( $h = 592$  km) underneath the Okhotsk Sea . Note the sharp impulsive body-wave onsets with well-developed depth phases.

In contrast to shallow-focus earthquakes, S phases from deep earthquakes may sometimes be recognizable even in teleseismic short-period records. The body-wave/surface-wave ratio and the type of generated surface waves are also key criteria for discriminating between local earthquakes, which mostly occur at depth larger than 5 km, and quarry blasts, underground explosions or rockbursts in mines, which occur at shallower depth (see 11.2.5.2).

A more precise determination of the depth  $h$  of a seismic source, however, requires either the availability of a seismic network with at least one station being very near to the source, e.g., at an epicentral distance  $D < h$  (because only in the near range the travel time  $t(D, h)$  of the direct P wave varies strongly with source depth  $h$ ), or the identification of seismic *depth phases* in seismic records. The most accurate method of determining the focal depth of an earthquake in routine seismogram analysis, particularly when only single station or network records at teleseismic distances are available, is to identify and read the onset times of depth phases (Chapter 2, section 2.6.3). A depth phase is a characteristic phase of a wave reflected from the surface of the Earth at a point relatively near the hypocenter (Fig. 2.62). At distant seismograph stations, the depth phases pP or sP follow the direct P wave by a time interval

that changes only slowly with distance but rapidly with depth. The time difference between P and other primary seismic phases, however, such as PcP, PP, S, SS etc. changes much more with distance. When records of stations at different distances are available, the different travel-time behavior of primary and depth phases makes it easier to recognize and identify such phases. Because of the more or less fixed ratio between the velocities of P and S waves with  $v_p/v_s \approx \sqrt{3}$  (Chapter 2, section 2.2.3), pP and sP follow P with a more or less fixed ratio of travel-time difference  $t(sP-P) \approx 1.5 t(pP-P)$  (Figs. 11.20 and 11.21). Animations of seismic ray propagation and phase recordings from deep earthquakes are given in files 4, 6 and 9 of IS 11.3.



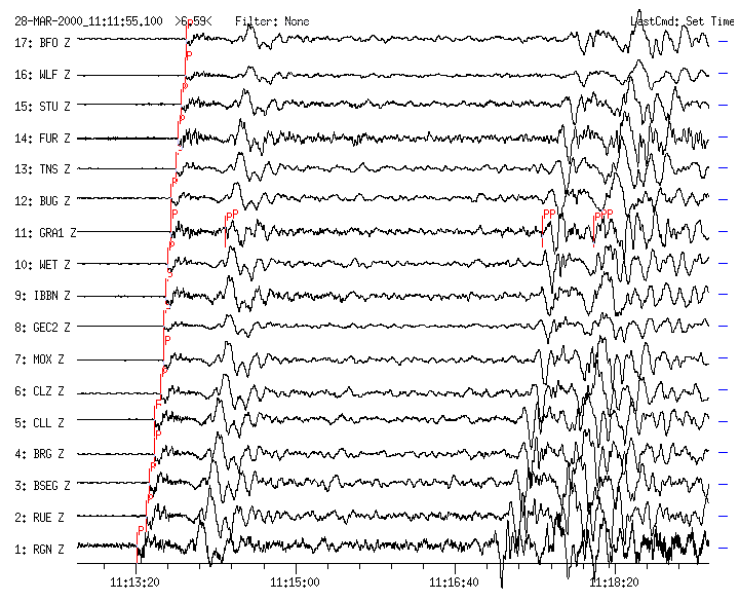
**Fig. 11.21** Short-period (left) and long-period (right) seismograms from a deep-focus Peru-Brazil border region earthquake on May 1, 1986 ( $m_b = 6.0$ ,  $h = 600$  km) recorded by stations in the distance range  $50.1^\circ$  to  $92.2^\circ$ . Note that the travel-time difference between P and its depth phases pP and sP, respectively, remains nearly unchanged. In contrast PcP comes closer to P with increasing distance and after merging with P on joint grazing incidence at the core-mantle boundary form the diffracted wave Pdif (reprinted from *Anatomy of Seismograms*, Kulhánek, Plate 41, p. 139-140; © 1990; with permission from Elsevier Science).

The time difference between pP and sP and other direct or multiple reflected P waves such as pPP, sPP, pPKP, sPKP, pPdif, sPdif, etc. are all roughly the same. S waves also generate depth phases, e.g., sS, sSKS, sSP etc. The time difference sS-S is slightly larger than sP-P (Figs. 1.10 and 11.20). The difference grows with distance to a maximum of 1.2 times the sP-P time. These additional depth phases may also be well recorded and can be used in a similar way for depth determination as pP and sP.

Given the rough distance between the epicenter and the station, the hypocenter depth ( $h$ ) can be estimated within  $\Delta h \approx \pm 10$  km from travel-time curves or determined by using time-

difference tables for depth-phases (e.g., from  $\Delta t(pP-P)$  or  $\Delta t(sP-P)$ ; see Kennett, 1991 or Table 1 in EX 11.2) or the “rule-of-thumb” in Eq. (11.4) in section 11.2.6.2. An example is given in Fig. 11.22. It depicts broadband records of the GRSN from a deep earthquake ( $h = 119$  km) in the Volcano Islands, West Pacific. The distance range is  $93^\circ$  to  $99^\circ$ . The depth phases  $pP$  and  $pPP$  are marked. From the time difference  $pP-P$  of 31.5 s and an average distance of  $96^\circ$ , it follows from Table 1 in EX 11.2 that the source depth is 122 km. When using Eq. (11.4) instead, we get  $h = 120$  km. This is very close to the source depth of  $h = 119$  km determined by NEIC from data of the global network.

Note that on the records in Fig. 11.22 the depth phases  $pP$  and  $pPP$  have larger amplitudes than the primary  $P$  wave. This may be the case also for  $sP$ ,  $sS$  etc., if the given source mechanism radiates more energy in the direction of the upgoing rays ( $p$  or  $s$ ; Fig. 2.62) than in the direction of the downgoing rays for the related primary phases  $P$ ,  $PP$  or  $S$ . Also, in Fig. 11.22,  $pP$ ,  $PP$  and  $pPP$  have longer periods than  $P$ . Accordingly, they are more coherent throughout the network than the more short-period  $P$  waves. Fig. 11.48 shows for the same earthquake the LP-filtered and rotated 3-component record at station RUE, Germany, with all identified major later arrivals being marked on the record traces. This figure is an example of the search for and comprehensive analysis of secondary phases.



**Fig. 11.22** Broadband vertical-component seismograms of a deep ( $h = 119$  km) earthquake from Volcano Islands region recorded at 17 GRSN, GRF and GEOFON stations. (Source data by NEIC: 2000-03-28 OT 11:00:21.7 UT;  $22.362^\circ\text{N}$ ,  $143.680^\circ\text{E}$ ; depth 119 km; mb 6.8;  $D = 96.8^\circ$  and  $BAZ = 43.5^\circ$  from GRA1). Traces are sorted according to distance. Amplitudes of  $P$  are smaller than  $pP$ . Phases with longer periods  $PP$ ,  $pP$  and  $pPP$  are much more coherent than  $P$ .

Crustal earthquakes usually have a source depth of less than 30 km, so the depth phases may follow their primary phases so closely that their waveforms overlap. Identification and onset-time picking of depth phases is then usually no longer possible by simple routine visual inspection of the record. Therefore, in the absence of depth phases reported by seismic stations, international data centers such as NEIC in its Monthly Listings of Preliminary (or

Quick) Determination of Epicenters often fix the source depth of (presumed) crustal events at 0 km, 10 km or 33 km, as has been the case for the event shown in Fig. 11.19. This is often further specified by adding the capital letter N (for “normal depth” = 33 km) or G (for depth fixed by a geophysicist/analyst). Waveform modeling, however (Fig. 2.85), may enable good depth estimates for shallow earthquakes to be obtained from the best fit of the observed waveforms to synthetic waveforms calculated for different source depth. Although this is not yet routine practice at individual stations, the NEIC has, since 1996, supplemented depth determinations from pP-P and sP-P by synthetic modeling of BB-seismograms. The depth determination is done simultaneously with the determination of fault-plane solutions. This has reduced significantly the number of earthquakes in the PDE listings with arbitrarily assigned source depth 10G or 33N.

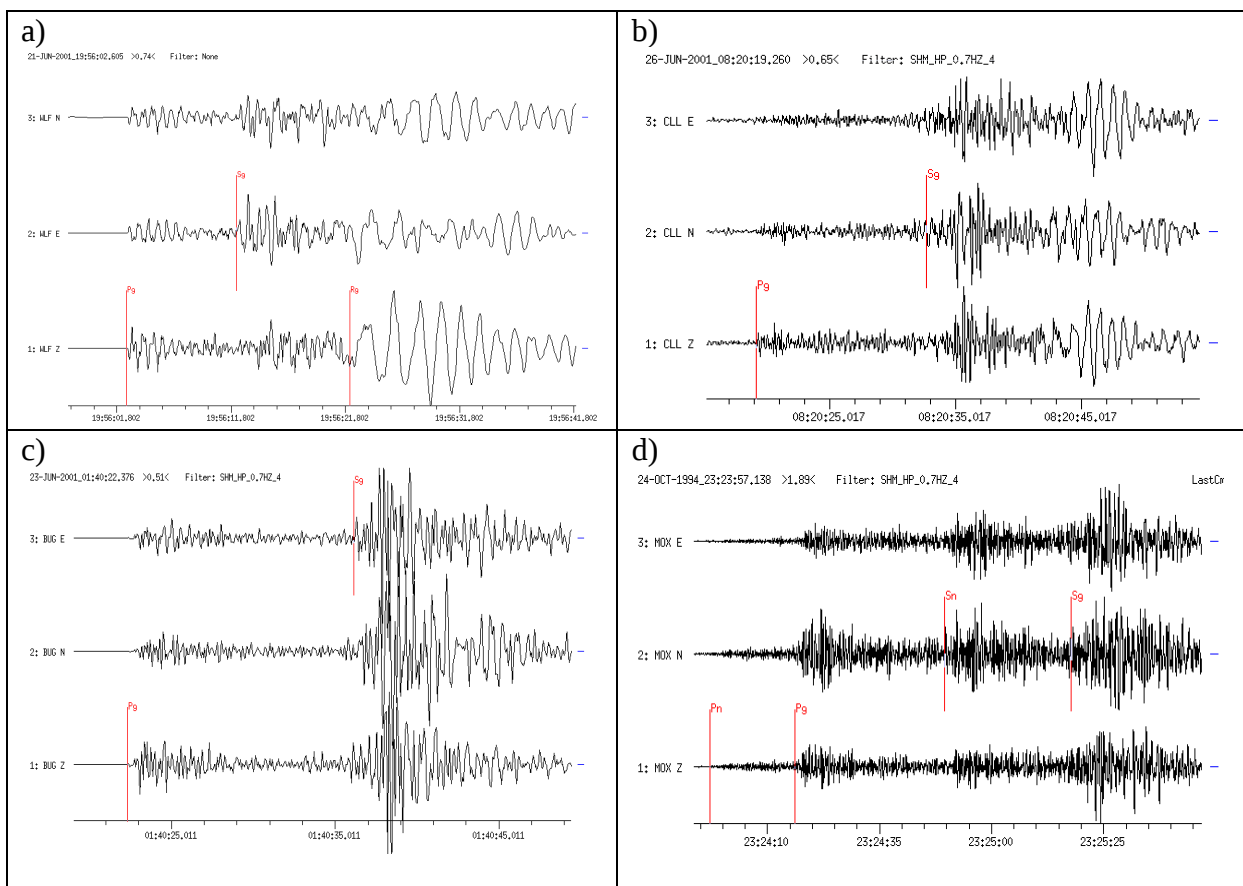
Note, however, that often there is no clear evidence of near-source surface reflections in seismic records, or they show apparent pP and sP but with times that are inconsistent from station to station. Douglas et al. (1974 and 1984) have looked into these complexities, particularly in short-period records. Some of these difficulties are avoided in BB and LP recordings. Also, for shallow sources, surface-wave spectra may give the best indication of depth but this method is not easy to apply routinely. Notably, Engdahl et al. (1998) developed an automatic phase identification algorithm, termed EHB for short. It aims at significantly improved hypocenter relocations. EHB is applied to the phase group which immediately follows the P phase at teleseismic distances, including pP, pwP, sP and PcP and allows also the unambiguous identification of later phases such as S. The EHB bulletin is available at the ISC website <http://www.isc.ac.uk/>. The EHB bulletin covers the period 1960-2008 and represented a significant improvement in routine hypocenter determinations made by PDE and ISC before data year 2009, when the ISC implemented a new location algorithm (Bondár and Storchak, 2011). The new ISC location algorithms matches the quality of the EHB bulletin and attempts to obtain a free-depth solution if enough depth resolution is available. Therefore, the ISC would benefit from reports of carefully picked depth phases.

In summary, observational seismologists should be aware that depth phases are vital for improving source locations and making progress in understanding earthquakes in relation to the rheological properties and stress conditions in the lithosphere and upper mantle. Therefore, they should do their utmost to recognize depth phases in seismograms despite the fact that they are not always present and that it may be difficult to identify them reliably. More examples of different kinds of depth phases are given in Figure 6b of DS 11.2 and Figures 1b, 2b, 5b and 7a+b in DS 11.3.

#### **11.2.5.2 Discrimination between natural earthquakes and man-made seismic events**

Quarry and mining blasts as well as controlled explosion charges may excite strong seismic waves. The largest of these events usually have local magnitudes in the range 2 to 4 and may be recorded over distances of several to several thousand kilometers. Rock bursts or collapses of large open galleries in underground mines may also generate seismic waves. The magnitude of these induced seismic events may range from around 2 to 5.5 and their waves may be recorded world-wide (as it was the case with the mining collapse shown in Fig. 11.11). In some countries with low to moderate natural seismicity but a lot of blasting and mining, anthropogenic (so-called “man-made” or “man-induced”) events may form a major fraction of all recorded seismic sources and even outnumber recordings of earthquakes. Then a major seismological challenge is the reliable discrimination of different source types.

According to Figs. 2.38 in Chapter 2 and Figures 4 and 5 in DS 11.4 the amplitude ratio P/S for chemical or underground nuclear explosions is generally larger than for tectonic earthquakes that are dominantly shear fractures. Fig.11.23 shows a comparison of seismograms from: (a) a mining-induced seismic event; (b) a quarry blast; (c) a local earthquake; (d) a regional earthquake. According to (a) and (b) the high-frequency body-wave arrivals are followed, after  $S_g$ , by well-developed lower-frequency and clearly dispersed crustal Rayleigh surface waves ( $R_g$ , with strong vertical components). This is not the case for the two earthquake records (c) and (d) because earthquakes, which are usually more than a few kilometers deep, do not generate short-period fundamental Rayleigh waves of  $R_g$  type. For even deeper (sub-crustal) earthquakes (e.g., Fig. 2.50) only the two high-frequency P- and S- wave phases are recorded within a few hundred kilometers from the epicenter.



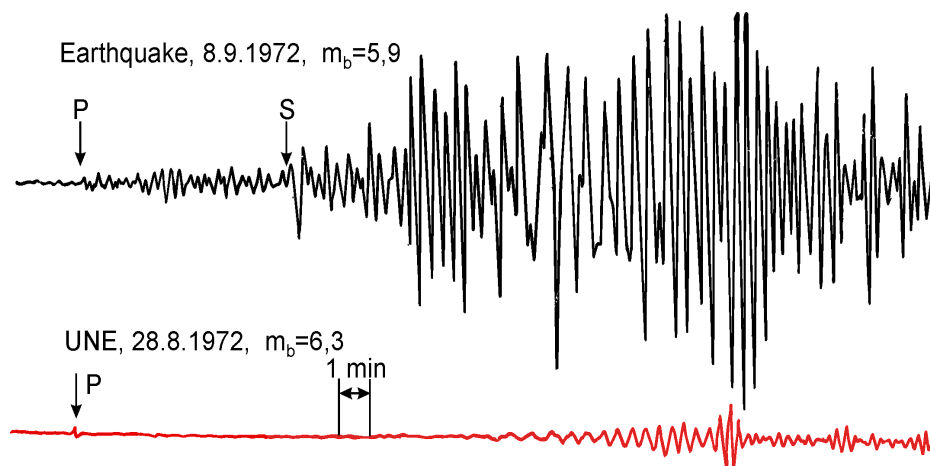
**Fig. 11.23** Examples of 3-component seismograms recorded from a: a) mining-induced seismic event ( $D = 80$  km); b) quarry blast explosion ( $D = 104$  km); c) local earthquake ( $D = 110$  km); d) regional earthquake ( $D = 504$  km). Note the different time scales.

Some observatories that record many quarry blasts and mining events, such as the SZGRF, have developed automatic discrimination filters on the basis of such systematic differences in frequency content and polarization. They allow routine separation of such man-made or induced seismic events from tectonic earthquakes. Chernobay and Gabsatarova (1999) give references to many other algorithms for (semi-) automatic source classification. These authors tested the efficiency of the spectrogram and the  $P_g/L_g$  spectral ratio method for routine

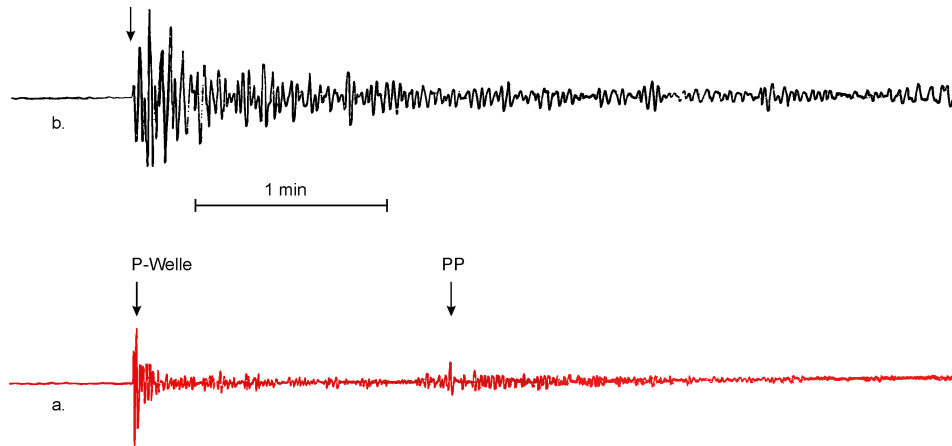
discrimination between earthquakes in the Northern Caucasus area of Russia with magnitudes  $< 4.5$  and chemical (quarry) explosions of comparable magnitudes. They showed that no single method can yet assure reliable discrimination between seismic signals from earthquakes and explosions in this region. However, by applying a self-training algorithm, based on hierarchical multi-parameter cluster analysis, almost 98% of the investigated events could be correctly classified and separated into 19 groups of different sources. However, local geology and topography as well as earthquake source mechanisms and applied explosion technologies may vary significantly from region to region. Therefore, there exists no straightforward and globally applicable set of criteria for reliable discrimination between man-made and natural earthquakes, thus requiring specific investigations for different regions.

In this context one should also discuss the discrimination between natural earthquakes (EQ) and underground nuclear explosions (UNE). The Comprehensive Nuclear-Test-Ban Treaty (CTBT) has been negotiated for decades as a matter of high political priority. A Preparatory Commission for the CTBT Organization (CTBTO) has been established with its headquarters in Vienna (<http://ctbto.org>) which is operating an International Monitoring System (IMS; Fig. 15.2 in Chapter 15). Agreement was reached only after many years of demonstrating the potential of seismic methods to discriminate underground explosions from earthquakes, down to rather small magnitudes  $m_b \approx 3.5$  to 4. Thus, by complementing seismic event detection and monitoring with hydro-acoustic, infrasound and radionuclide measurements it is now highly probable that test ban violations can be detected and verified.

The source process of UNEs is simpler and much shorter than for earthquake shear ruptures. Accordingly, P waveforms from explosions are shorter, have higher predominant frequencies, are usually more impulse-like than for earthquakes and expected to have compressional first motions in all directions. Also, UNEs generate lower amplitude S and surface waves than earthquakes of the same body-wave magnitude (Figs. 11.24 and 11.25).

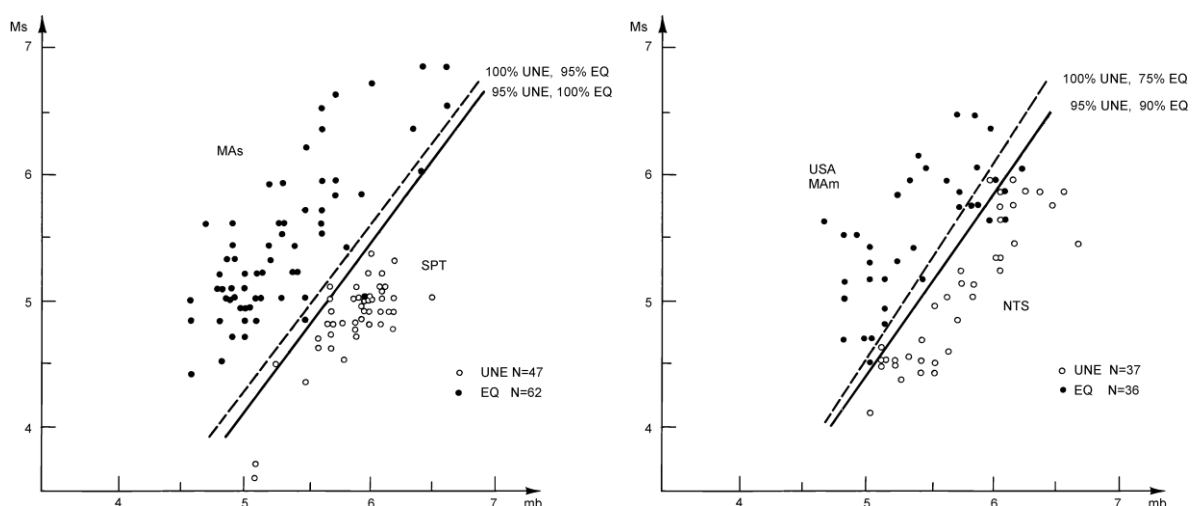


**Fig. 11.24** Broadband displacement records at station Moxa (MOX), Germany of a Jan Mayen earthquake and a Semipalatinsk underground nuclear explosion (UNE) of comparable short-period body-wave magnitude  $m_b$  and at nearly the same distance of about  $40^\circ$ .



**Fig. 11.25** Short-period records at station MOX a) of an underground nuclear explosion at the Semipalatinsk (SPT) test site in Kazakhstan ( $D = 41^\circ$ ) and b) of an earthquake in Afghanistan with comparable magnitude around 6 and at similar distance.

A powerful discriminant is the ratio between short-period P-wave magnitude  $m_b$  and long-period surface-wave magnitude  $M_s$ . The former samples energy around 1 Hz while the latter samples long-period energy around 0.05 Hz. Accordingly, much smaller  $M_s/m_b$  ratios are observed for explosions than for earthquakes (Fig. 11.26). Whereas for a global sample of EQs and UNEs the two populations overlap in an  $M_s/m_b$  diagram, the separation is good when earthquakes and explosions in the same region are considered (e.g., Bormann, 1972c). Early studies have shown that with data of  $m_b \geq 5$  from only one teleseismic station 100% of the observed UNEs with such magnitudes from the Semipalatinsk (SPT) test site could be separated from 95% of the EQs in Middle Asia (Fig. 11.26 left), whereas for the more distant test site in Nevada ( $D = 81^\circ$ ) 95% of the UNEs could be discriminated from 90% of the EQs in the Western USA and Middle America (Fig. 11.26 right).



**Fig. 11.26** Separation of EQs and UNEs by the  $M_s/m_b$  criterion according to data collected at station MOX, Germany. **Left:** for Middle Asia and the test site in Semipalatinsk (SPT); **Right:** for USA/Middle America and the Nevada test site (NTS).

Bowers and Walter (2002) published a similar mb-Ms diagram for US earthquakes and UNE with an average offset of about 0.8 mb units at  $M_s = 5$ , which is comparable with Fig. 11.26 and corresponds to about 6 times larger P-wave amplitudes around 1 Hz for explosions than for earthquakes with the same surface wave magnitude.

Other potential discrimination criteria, such as the different azimuthal distribution of P-wave first-motion polarities expected from UNEs (always +) and EQs (mixed + and -), have not proved to be reliable. One reason is, that due to narrowband filtering, which is applied to reach the lowest possible detection threshold, the P waveform, and particularly the first half cycle, is often so much distorted, that the real first-motion polarity is no longer recognizable in the presence of noise (Figs. 4.10 and 4.11). Detailed investigations also revealed that simplified initial model assumptions about the difference between explosion and earthquake sources do not hold. Surprisingly, the explosion source is poorly understood and source dimensions around magnitude mb  $\approx 4$  seem to be about the same for earthquakes and explosions. Also, many explosions do not approximate to a point-like expansion source in a half-space: significant Love waves may be generated (e.g., by Novaya Zemlya tests in a terrain of rough topography or by triggered tectonic stress release) and many P-wave seismograms show arrivals that cannot be explained (see, e.g., Douglas and Rivers, 1988). Further, it has become clear that much of the differences observed between records of UNEs and EQs are not due to source differences but rather to differences in the geology, topography and seismotectonics of the wider area around the test sites, and that this necessitates the calibration of individual regions (e.g., Douglas et al., 1974).

In summary, one can say that the key criteria to separate EQs and explosions usually work well for large events, however, difficulties come with trying to identify every EQ down to magnitudes around mb = 4 with an average of about 8000 earthquakes of this size per year. It is beyond the scope of this section to go into more detail on this issue. For CTBTO goals and procedures see Chapter 15. State-of-the-art summaries have been published by Bowers and Selby (2009), CTBTO PrepComm, (2009; <http://www.ctbto.org/verification-regime/>), Dahlman et al. (2009), Richards (2002) and Richards and Wu (2011).

### **11.2.6 Quick event identification and location by means of single-station three-component recordings**

Recognizing earthquakes in and locating them by means of seismic records, including also their size classification, are the most important tasks of observatory seismology. Increasingly seismograms are being analyzed at centers that receive the data in (near) real time from networks or arrays of seismometers (Chapters 8 and 9). The seismograms can then be analyzed jointly. This has clear advantages. Nevertheless, some seismic stations are still treated as single station for basic processing. Some of them still record with analog techniques. Yet much can be done even under these “old-fashioned” conditions by the station personnel, provided that at least some form of 3-component recording, either BB or both SP and LP, is available. If these records are in digital form, which is nowadays common, then modern software for digital seismogram analysis will make this task much easier and faster than in the “analog past”. But for most users of ready-made analysis software packages these tools are “black boxes”. Surely, to go through the various steps of analog and graphical location or other related seismological analysis procedures, is much more time consuming. This makes their application unacceptable in modern operational observatory practice. But the



classical analog procedures tell us much more about the underlying principles that are now implemented in algorithmic form in the analysis software. Therefore, those interested in the history of observatory practice but also students in undergraduate Earth science or “Seismology at Schools” courses should have taken note of and practiced them oneself, at least once, to get the necessary “feeling”. The same applies to the sections 11.2.7 on seismic network recordings and 11.2.8 on magnitude determinations, which also give reference to related basic exercises by hand. But all has to start with the single station seismogram analysis.

#### 11.2.6.1 What to do with single station seismograms?

3-component records allow to quickly assess the source type and to roughly estimate the source location and event magnitude. Moreover, later seismic phases can be identified in quite some detail, even if rapid epicenter determinations by international data centers are not available. There would be advantages if, in future, such readings of secondary phases, especially of depth phases, will be reported as early as possible to regional and global data centers so as to enable better depth determinations. Indeed, only recently both NEIC and the ISC introduced more flexible and sophisticated algorithms and the AK135 travel-time model in order to make better use also of secondary phase readings for more reliable hypocenter locations. Locations derived in near real-time usually have not yet well constrained source depths since they are based only on first arriving P and (eventually) S.

It has also been realized that epicentral distances estimated from three-component broadband readings of secondary phases may significantly improve location estimates by small seismic array stations that are based on measurements of the P-wave vector slowness only. The reason is that the travel-time differences between P and later arriving phases, especially of S and SS, vary much more with distance than the first P-wave arrival-time and thus the first P-wave slowness differences at the various stations. Other procedures of modern multi-station (but usually single-component) data analysis are mentioned later *en passant*. Array analysis is discussed in detail in Chapter 9.

How one may proceed when analyzing interactively single station 3C-seismograms without having any prior information about the source and aiming at event location is outlined in EX 11.2 which may be suitable also for demonstration or a practical in “Seismology at Schools” sessions. Additionally, one might look for criteria discussed in sub-section 11.2.5.2 for discriminating between explosions and earthquakes.

If modern broadband digital records are available, which are usually proportional to ground velocity, it is recommended to filter them to produce simulated standard analog Wood-Anderson (WA), WWSSN-SP and -LP seismograms (for responses see Fig. 11.38). Because these types of records are required for proper estimation of the most frequently measured classical types of magnitudes (ML, mb and Ms<sub>20</sub>) according to recently established IASPEI measurement standards (Chapter 3 and IS 3.3) and they may also ease phase identification by mutual inspection. One may also simulate Kirnos SKD BB-displacement and SRO-LP responses (Fig. 11.38) which are particularly suitable for later phase identification in the teleseismic range. This is documented in the following and in DS 11.2 and 11.3 with many record examples. The detailed seismogram analysis might then include phase identification, picking of onset times, amplitudes, periods and polarities, and, if required, also the application of special filters, such as the inverse Hilbert transformation, to phases that have been distorted

by traveling through internal caustics (Chapter 2, section 2.5.4.3), or for identifying phases on the basis of their different polarization.

Of course, in countries with many seismic events recorded every day it will not be possible, particularly for untrained analysts, to apply all these criteria to every recording. On the other hand, this kind of checking takes only a few seconds, minutes at most, for an experienced interpreter who has already trained himself/herself in recognizing immediately the different record patterns on seismograms from systems with standard responses. In addition, many data centers specialize in analyzing only seismograms from local, regional or teleseismic sources. Accordingly, either the number of questions to be asked to the record or the number of signals to be analyzed will be reduced significantly. Also, the task might be significantly eased at observatories or analysis centers which have advanced routines available for digital seismogram analysis such as SEISAN, Seismic Handler (SHM) or SeisComp3. Provided that first hypocenter estimates are already available from international data centers or from own automatic real-time analysis of array or network recordings (e.g., by using SeisComp3), these computer programs allow the theoretical onset times of expected seismic phases to be displayed on the seismogram. However, these theoretical times should not be followed blindly but considered only as assistance. The additional information on amplitudes, frequency content, and polarization has to be taken into account before giving a definite name to a recognizable onset (see 11.2.4 and 11.2.5).

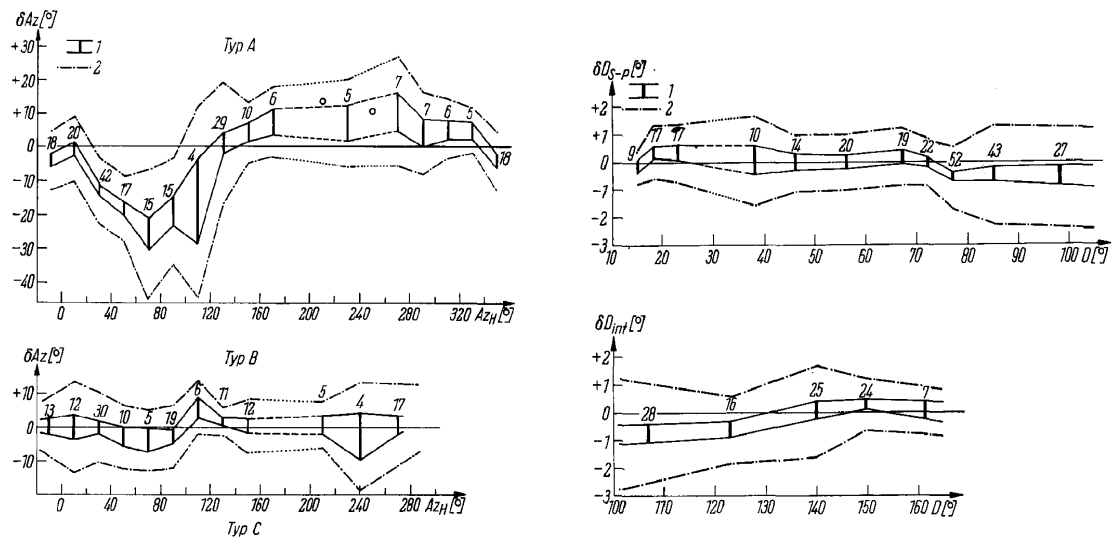
On the other hand, it is meaningless to list more detailed and strict criteria and rules about the appearance and identification of seismic phases in different distance ranges, because they vary from event to event and from source region to source region. They also depend on the specific conditions along the wave-propagation paths and at the recording site. Therefore, every station operator or network analyst has to develop, through experience and systematic data analysis, his/her own criteria for improved seismogram analysis, phase and source identification, and source location. In any event, however, the general approach to record analysis given above should be followed in order to avoid the analysis becoming thoughtless, boring and routine, which would inevitably result in the reporting of inhomogeneous and incomplete low-quality data of little value for research or to the general user.

### **11.2.6.2 3-component hypocenter location**

If well calibrated 3-component broadband and/or long-period recordings are available then it is possible to locate sufficiently strong local events ( $M_l > 3$ ) and teleseismic sources ( $m_b > 5$ ) with an accuracy comparable to or even better than those for un-calibrated arrays or station networks, which, however, should nowadays no longer exist. Yet, this could be demonstrated already some 40 years ago (Bormann, 1971a and b) by using standard film records of type A, B and C (for responses see Fig. 3.20). Amplitudes and onset times were at that time still measured by using an ordinary ruler or a sub-millimeter scaled magnification lens. Nevertheless, the mean square error of epicenters thus located within the distance range  $20^\circ < D < 145^\circ$  was less than 300 km when compared with the epicenter coordinates published by the seismological World Data Centers A and B. This result compares well with the  $4^\circ$  cap location constraint achieved by the fully automatic multi-parameter processing procedure developed by Tong and Kennett (1996) for single 3-component broadband seismograms.

Fig. 11.27 shows the statistical distribution of errors in azimuth and distance based on several hundred 3-component event locations made between 1966 and 1968 at station MOX,

Germany. The errors in epicentral distance estimates, based on reading the S-P times in Kirnos BB records, were in the range  $15^\circ < D < 80^\circ$  in 90% of the cases less than  $\pm 1^\circ$  and in the range  $80^\circ < D < 120^\circ$  also mostly less than about  $\pm 1^\circ$  and rarely greater than  $\pm 2^\circ$ . For  $100^\circ < D < 165^\circ$ , when using in the core shadow of P time differences between other seismic phase arrivals, the errors are somewhat larger, but even then the mean errors seldom differed significantly from zero, and where they did this was usually related to specific source regions, respectively distance/azimuth ranges. This was better than for the automatic Tong and Kennett (2000) procedure for which the authors report that “...in most cases the estimated distances are within the error of 7%.”



**Fig. 11.27 Left:** Errors in backazimuth Az (or BAZ) at station MOX estimated using 3-component records of type A (SP) and of type C (Kirnos SKD BB-displacement). **Right:** Errors in estimating the epicentral distance D at station MOX from records of type C using travel-time difference S-P in the distance range  $10^\circ < D < 100^\circ$  or travel-time differences between other seismic phases for  $D > 100^\circ$ . The solid lines give the 90% confidence interval for the mean error with number of the observations; the dash-dot lines are the 90% confidence interval for a single observation within different distance intervals. Figure according to Bormann (1971b)

When measuring BAZ on short-period records systematic errors may be rather large, reaching up to several tens of degrees (Fig. 11.27 upper left). In contrast, when BAZ is determined on LP or BB records (provided that the magnification of the horizontal components is identical or known with high accuracy) than the BAZ errors are much less, with about 90% probability within  $\pm 10^\circ$  (Fig. 11.27 lower left). Here the Tong and Kennett (1996) results are better, at least for events at distances up to  $100^\circ$ , for which the estimated azimuths were within the error of  $5^\circ$  because of the superior polarization analysis possible with high resolution digital data. The reason for the much larger azimuth errors when using short-period data is obvious from Fig. 2.7 in Chapter 2. The particle motion is complicated in SP records and random due to wave scattering and diffraction by small-scale heterogeneities in the crust and by rough surface topography at or near the station site (see also Buchbinder and Haddon, 1990). In contrast, LP or BB records, which are dominated by longer wavelength signals, usually show simpler P waveforms with clearer first-motion polarity than do SP records.

Taking systematic distance and/or azimuth errors into account, the location accuracy can be improved. Nowadays many seismic arrays and networks use routinely multi-phase epicentral distance determinations for improving their slowness-based source locations. Some advanced software for seismogram analysis like SHM (section 11.4.2) and SEISAN (section 11.4.3 and section 6 in IS 11.6) include this complementary interactive analysis feature. EX 11.2 discusses in detail two examples based on earthquake records made at  $D \approx 20^\circ$  and  $\approx 90^\circ$ . Fig. 11.28 below gives an example for a Pakistan earthquake at  $D = 46.2^\circ$ .

Rough estimates of hypocenter, respectively epicenter distance for shallow seismic sources can be made, however, even in the absence of travel-time tables or curves or related computer programs, when using the following simple “rules-of-thumb” (with  $1^\circ = 111.12 \text{ km}$ ):

$$\text{hypocenter distance } d \text{ [in km]} \approx \Delta t(\text{Sg-Pg}) \text{ [in s]} \times 8 \text{ (near range only)} \quad (11.1)$$

$$\text{epicentral distance } D \text{ [in km]} \approx \Delta t(\text{Sn-Pn}) \text{ [in s]} \times 10 \text{ (in Pn-Sn range } < 15^\circ) \quad (11.2)$$

$$\text{epicentral distance } D \text{ [in } ^\circ] \approx \{\Delta t(\text{S-P}) \text{ [in min]} - 2\} \times 10 \text{ (for } 20^\circ < D < 100^\circ) \quad (11.3)$$

For deep earthquakes one may use another “rule-of-thumb” for a rough estimate of source depth from the travel-time difference  $\Delta t(\text{pP-P})$ :

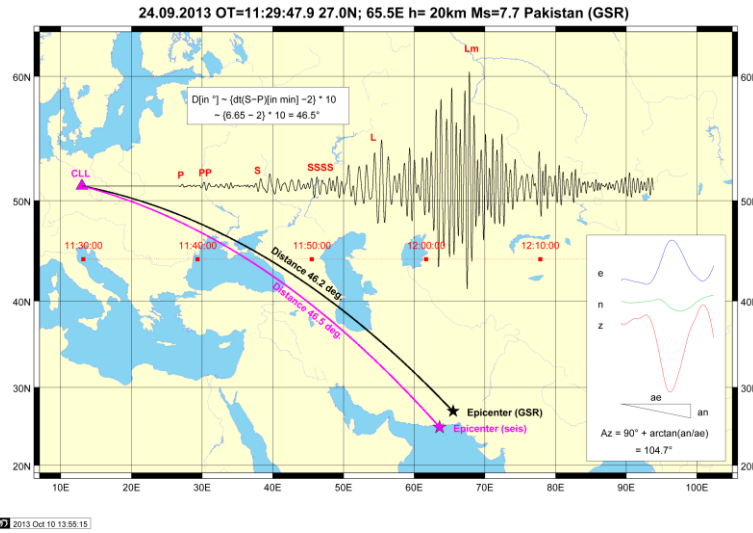
$$\begin{aligned} \text{source depth } h \text{ [in km]} \approx & \Delta t(\text{pP-P})/2 \text{ [in s]} \times 7 \text{ (for } h < 100 \text{ km)} \\ & \text{or } \times 8 \text{ (for } 100 \text{ km} < h < 300 \text{ km)} \\ & \text{or } \times 9 \text{ (for } h > 300 \text{ km)} \end{aligned} \quad (11.4)$$

The Pakistan earthquake was obviously shallow because it produced a well-developed surface-wave train with amplitudes much larger than the preceding body waves. Moreover, from the travel-time difference of 6.65 min between the identified P and S onset follows according to formula (11.3)  $D = 46.5^\circ$ . This differs only about  $0.3^\circ$  or 33 km from  $D(\text{CLL}) = 46.2^\circ$  calculated by GSR, corresponding to a distance deviation of only 0.065%.

When determining the backazimuth BAZ of the event with respect to the recording station from 3-component records in addition, it is possible to locate the event on a globe or on a suitable map projection. The determination of BAZ by using P-wave first-motion polarity readings has been described in detail in section 3.2 of EX 11.2 and in section 2 of IS 11.1. The insert in the upper right corner of Fig. 11.28 shows the magnified first motion direction of the long-period P-wave onset of the 2013 Pakistan earthquake. One clearly recognizes that the motion in the vertical component goes down, i.e., the ground motion shows in the direction towards the source, in the E-W component it moves with large amplitude up, i.e., the source is located in the East, and the N-S component goes weakly down, i.e., the ground moves also a bit southward. Measuring the correct ground amplitudes in the horizontal components one gets the ratio  $A_S/A_E = 0.262$ . This corresponds to an angle of  $14.7^\circ$  and thus a  $\text{BAZ} = 90^\circ + 14.7^\circ = 104.7^\circ$ .

In contrast, the BAZ to station CLL, when calculated for the more precise epicenter determined on the basis of some hundred P-wave onset-time picks on global seismic network records is  $\text{BAZ} = 101.1^\circ$ , i.e., a deviation of only  $3.7^\circ$  from the 3-component  $\text{BAZ}(\text{CLL})$ . Comparing the two epicenter positions in Fig. 11.28 one can conclude that the location of this event in SE Pakistan has been correctly determined from a single station record more than 5000

km away with a total position difference of only  $2.7^\circ$  or 300 km with respect to the global network location. Thus this 3-component single station location agrees best with the conclusion by Tong and Kennett (1996) that for automatic multi-parameter 3-C broadband locations D estimates are mostly within the error of 7%, azimuths within  $5^\circ$  for  $D < 100^\circ$ , and the locations within a  $4^\circ$  cap.



**Fig. 11.28** Epicenter location of the  $M_w = 7.7$  Pakistan earthquake on 24.09.2013 by the Obninsk data center GSR on the basis of global P-wave onset readings (black star) and by single station 3-component analysis at station CLL, Germany (cyan star), using the travel-time difference S-P for determining D and the long-period P-wave first-motion polarity in Z, E, and N component for determining  $BAZ = \arctan(A_S/A_E) + 90^\circ$ . Plotting single station locations on an epicenter map is a good check that the location is reasonable.

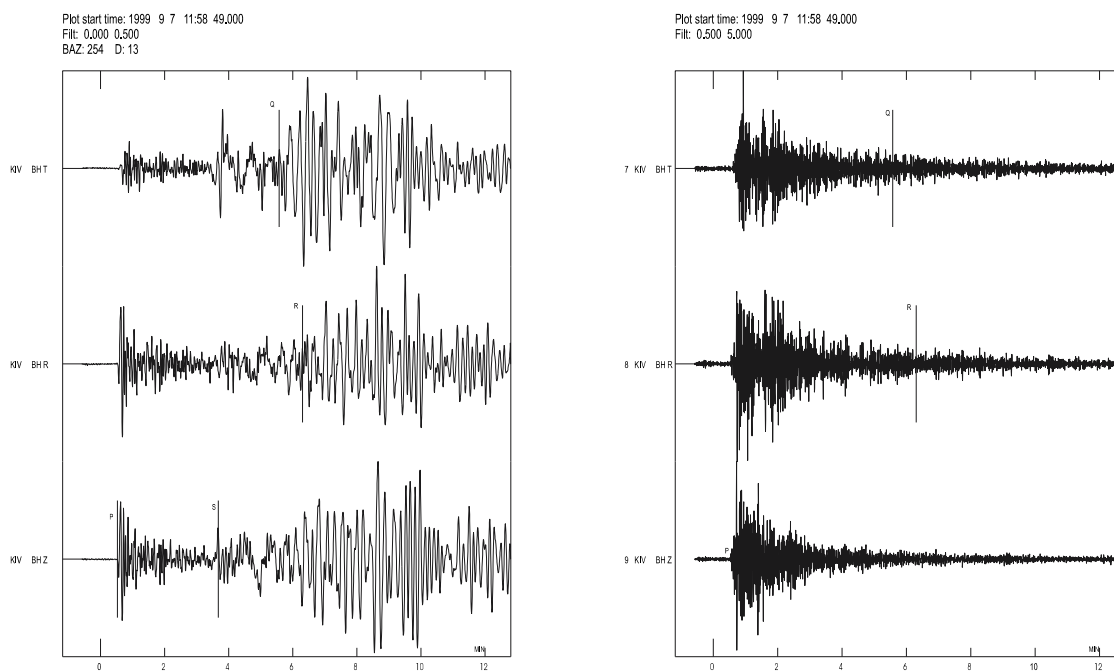
However, also global network epicenter positions may differ up to several 10 km from the true positions due to not accounting for velocity inhomogeneities in the real Earth along the travel paths to the stations (e.g., Engdahl et al., 1998) and also, that the earthquake location itself is not a sharp point but a rupture of finite length which is according to Eq. (3.188) in Chapter 3 for a magnitude  $M = 7.7$  earthquake on average  $L \approx 80$  km.

Bormann (1971a) also showed that in the absence of a sufficiently strong P-wave arrival, the BAZ may also roughly be determined from horizontal components of any later seismic phase which is polarized in the vertical propagation plane, such as PP, PS, PKP or SKS. These phases are often much stronger in BB or LP records than P. However, because of phase shifts on internal caustics (PP, PS, SP, PKPab) the vertical component first motion direction of most of these phases cannot be used reliably for resolving the  $180^\circ$  ambiguity of an azimuth determined from the ratio  $A_E/A_N$  only. However, by considering the inhomogeneous global distribution of earthquake belts, this problem can usually be solved.

In summary, simple 3-component event locations based solely on readings of onset times of identified phases, polarity of P-wave first motions and horizontal component amplitude ratio should proceed as follows:

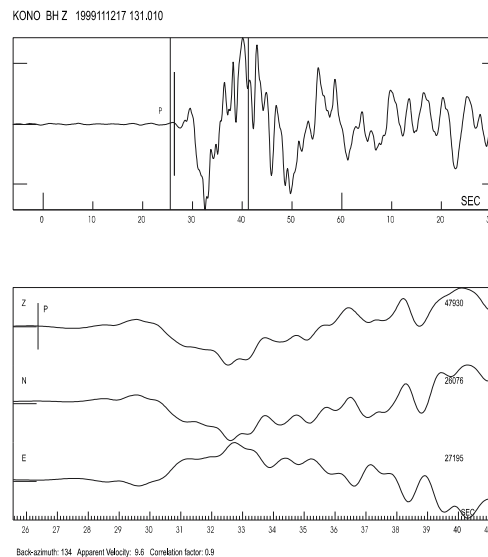
- general event classification (near/far; shallow/deep;  $D < 100^\circ / > 100^\circ$  etc.);

- picking and identifying the most pronounced phases by comparing the 3-component record traces and related polarization characteristics (Fig. 11.29);
- determination of  $D$  by **a)** matching the identified body-wave phases with either overlays of differential travel-time curves of equal time scale (as in Fig. 11.8 or Figures 2 to 4 in EX 11.2), **b)** measuring their onset-time differences and comparing them with respective distance-dependent differential travel-time tables or by using rules of thumb [see formulas (11.1 to 11.4), **c)** computer calculation of  $D$  based on digital time picks for identified phases and local, regional and/or global travel-time models integrated into the analysis program;
- determination of source depth  $h$  on the basis of identified depth phases (see 11.2.5.1 and Figure 7 with related table in EX 11.2). When knowing  $h$  then correct  $D$  by using either travel-time curves, differential  $t$ - $D$  tables or computer assisted time-picks and comparison with travel-time models;
- determination of the backazimuth (against North) from the station to the source from the first-motion directions in the original Z, N and E component records and from the amplitude ratio  $A_E/A_N$ . For details see Figure 1 and related explanations in EX 11.2;
- determination of the epicenter location and coordinates by using appropriate map projections with isolines of equal azimuth and distance from the station (Figures 5 and 6 in EX 11.2) or by means of suitable computer map projections.



**Fig. 11.29 Left:** Low-pass filtered digital broadband record of the Global Seismograph Network (GSN) station KIV from the shallow ( $h = 10$  km) Greece earthquake of 07 Sept. 1999 ( $m_b = 5.8$ ) at a distance of  $D = 13^\circ$ . Note the clearly recognizable polarity of the first P-wave half-cycle! The record components have been rotated into the directions Z, R and T after determination of the backazimuth from first-motion polarities in Z, N and E ( $BAZ = 134^\circ$ ). Accordingly, P and Rayleigh waves are strongest in Z and R while S and Love (Q) wave are strongest in T. **Right:** The recordings after SP bandpass filtering (0.5-5.0 Hz). The SNR for the P-wave first-motion amplitude is much smaller and their polarity less clear. Also later arrivals required for distance determination are no longer recognizable (signal processing done with SEISAN; courtesy of L. Ottemöller).

Modern computer programs for seismogram analysis include subroutines that allow quick determination of both azimuth and incidence angle from particle motion analysis over the whole waveform of P or other appropriate phases. This is done by determining the direction of the principal components of the particle motion, using, as a measure of reliability of the calculated azimuth and incidence angle, the degree of particle motion linearity/ellipticity. Such an algorithm was available in the SEIS89 software (Baumbach†, 1999). Christoffersson et al. (1988) describe a maximum-likelihood estimator for analyzing the covariance matrix for a single three-component seismogram (see also Roberts and Christoffersson, 1990). The procedure allows joint estimation of the azimuth of approach, and for P and SV waves the apparent angle of incidence and, hence, information on apparent surface velocity and thus on epicentral distance. This has been implemented in the SEISAN software (Havskov and Ottemöller, 1999). Fig. 11.30 shows an example of the application of the software to a portion of the BB recording at Kongsberg (KONO) in Norway for the 12 November 1999, Turkey earthquake ( $M_w = 7.1$ ). The program finds a high correlation (0.9) between the particle motions in the three components, gives the estimate of the backazimuth as  $134^\circ$ , an apparent velocity of 9.6 km/s and the corresponding location of this earthquake at  $40.54^\circ\text{N}$  and  $30.86^\circ\text{E}$ . This was only about 50 km off the epicenter located by the global network.



**Fig. 11.30** Example of azimuth determination and epicenter location of the 12 Nov. 1999 Turkey earthquake by correlation analysis of three-component digital BB records at station KONO, Norway. Backazimuth, apparent velocity, and correlation factor are determined from the P-wave record section marked in the upper figure. For more details see text (signal processing done with SEISAN; courtesy of L. Ottemöller).

The TREMORS Algorithm by Talandier et al. (1991) determines automatically the epicenter of local, regional and teleseismic events based on the 3-component recordings of a single broad-band station. The program was originally developed for the fast tsunami warning in the French South Pacific Islands where network locations are not readily available. The Mantle magnitude  $M_m$  (Okal and Talandier, 1989) is used to characterize the size of the earthquake. An open version of the software was published by Dominique Reymond as part of his Seismic Tool Kit STK (Reymond (2002), <http://seismic-toolkit.sourceforge.net/>).

Applying another algorithms to digital 3-component data from short-period P waves recorded at regional distances, Walck and Chael (1991) show that more than 75% of the records yielded backazimuth within 20° of the correct values. They found, however, a strong dependence on the geological structure. Whereas stations located on Precambrian terrains produced accurate backazimuth for SNR > 5 dB, stations on sedimentary rocks with complicated structure had much larger errors. When excluding these stations, the RMS backazimuth error was only about 6° for recordings with SNR > 10 dB.

Ruud et al. (1988) found that 3-component BB locations for epicenters at distances up to about 1000 km seldom deviated at that time more than 50 km from network solutions, such deviations being mainly due to errors in azimuth estimates. For short-period teleseismic P waves, however, location errors occasionally exceeded 800 km, mainly because of poor distance estimates derived from incidence angles (slowness) alone. For stronger sources, using BB records and distance estimates based on travel-time differences, the location errors were reduced to about 1°.

These quantitative values may no longer be representative when comparing nowadays single station 3-component locations with improved modern network locations. However, the general message still holds, namely that already single station 3-component digital BB data allow reasonably good first location estimates when, for what reasons ever, faster and better network locations are not yet available for the interested user. The suitability of the combined use of seismic array and single 3-component locations had been discussed already by Cassidy et al. (1990).

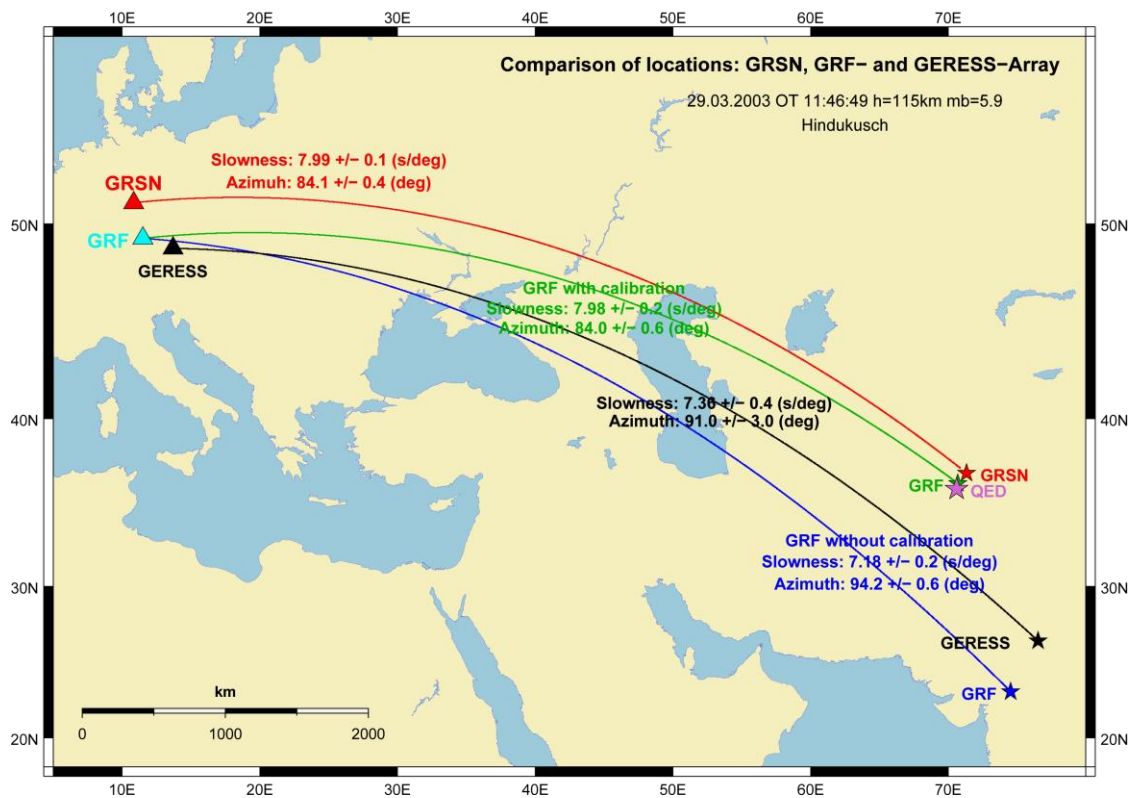
### **11.2.7 Hypocenter location by means of network and array recordings**

Hypocenter location and origin time determination require at least 4 stations if only first arrival times are read or three stations if independent distance estimates are available from travel-time differences between different phases. The more stations with good SNR (high weighting factor!) are available in local, regional or global networks and the more uniformly these stations are distributed around a seismic source in both azimuth and distance (the latter ranging from close-in to long range) and the more correctly identified and carefully measured seismic phase onsets are used for location, the lower is the uncertainty in these parameter estimates. The procedures in both manual and computer assisted multi-station hypocenter location are outlined in much more detail in IS 11.1, which gives the underlying algorithms and error calculations, as well as standard and advanced methods for both absolute and relative location. Also discussed are the influence of deviations from the assumed Earth models on the locations and the improvement of hypocenter relocations achievable with better Earth models.

Other special problems of seismic event location are dealt with in other information sheets. E.g., IS 7.4 looks into the detectability and earthquake location accuracy modeling of seismic networks, IS 8.5 outlines an authoritative operational location scheme for assessing the quality of automatic network locations, IS 8.6 focusses on the identification and collection of seismic events with accurately known locations (ground truth data) for network calibration and IS 8.7 offers and comments on a tutorial about theoretical and empirical approaches aimed at optimizing network configurations for different seismological tasks.



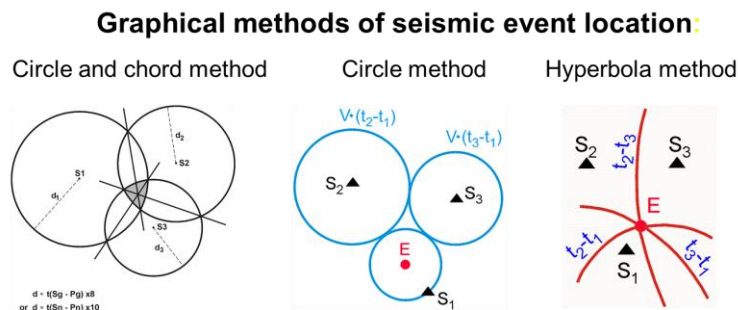
Array methods of event location are described in Chapter 9. They require that the distance of the source from the array/network is much larger than the array/network aperture (flat wave-front approach) and a high degree of waveform coherence/correlation between the station records. Although seismic arrays are very fashionable, mainly because of their potential for SNR improvement and direct slowness measurements (Fig. 11.45 and sections 11.3.3-11.3.5) there locations may not be superior at all as compared to those of well distributed seismic networks surrounding the source areas. The results may however be improved significantly by investigating and accounting for systematic array or network mislocations due to distance and azimuth dependent slowness anomalies (e.g., Bormann et al., 1992 and Fig. 11.31; Krüger and Weber, 1992; Schweitzer, 2002).



**Fig. 11.31** Comparison of the epicenter locations determined by the GRSN and the Gräfenberg (GRF) and GERES seismic arrays in Germany for a Hindukush earthquake at intermediate depth. QED gives the position of the NEIC/USGS Quick Epicenter Determination based on global network records. The QED location and the epicenters determined by the GRSN with the largest aperture (about 590 km × 700 km), and by the calibrated GRF (aperture about 50 km × 120 km) agree very well (differences less than 100 km). This is not the case for uncalibrated locations of GRF and the very small GERES array (aperture only some 4 km). They are strongly affected by local site anomalies (velocity and/or topographic ones; for GERES see, e.g., Harjes et al., 1994; Bokelmann, 1995) and their locations are about 600-700 km off the QED solution.

In the following we outline for general understanding three basic graphical approaches to locate local seismic events. They all require a “flat Earth” or an “Earth flattening” pre-processing (Fig. 11.32):

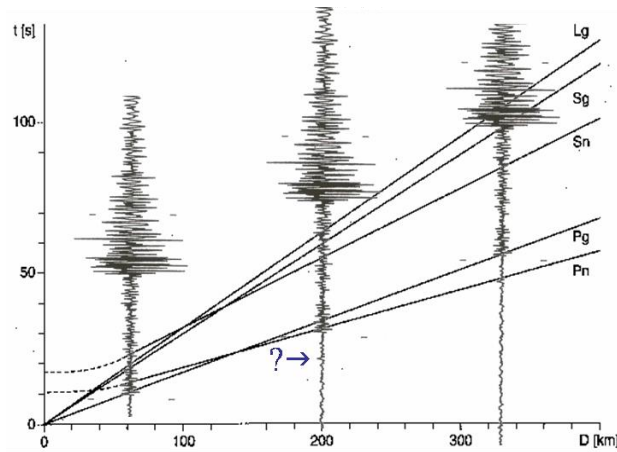
1. The **circle and chord method** calculates for each station the hypocentral distance  $d$ , using a known or assumed travel-time model and the relationship  $\mathbf{d} = (\mathbf{t}_S - \mathbf{t}_P) \{(\mathbf{v}_P \cdot \mathbf{v}_S) / (\mathbf{v}_P - \mathbf{v}_S)\}$ . In a crust with mean P-wave velocity of about 5.9 km/s and  $v_S = v_P/\sqrt{3}$  this can be approximated by  $\mathbf{d}$  [in km]  $\approx 8 \cdot \Delta t$  (Sg-Pg) [in s] [see “rule of thumb” (11.1)]. This method is rather insensitive to absolute variations of  $v_P$  and  $v_S$  as long as the ratio  $v_P/v_S$  remains constant.
2. The **circle method** uses only measured P-wave first arrival times at the stations and an assumed reasonable P-wave velocity  $v$ . Circles are drawn around the stations  $n-1$  with radii equal to  $v(\mathbf{t}_n - \mathbf{t}_1)$ ,  $\mathbf{t}_n$  being the P onset time at station  $n$ . The epicenter is then found in the center of the circle which runs through station 1 and touches also the other circles.
3. The **hyperbola method** uses only P-wave arrival time differences between selected station pairs. It requires no à priori knowledge of the P-wave velocity and yields rather good results. For a detailed description of the method and its benefits see Pujol (2004).



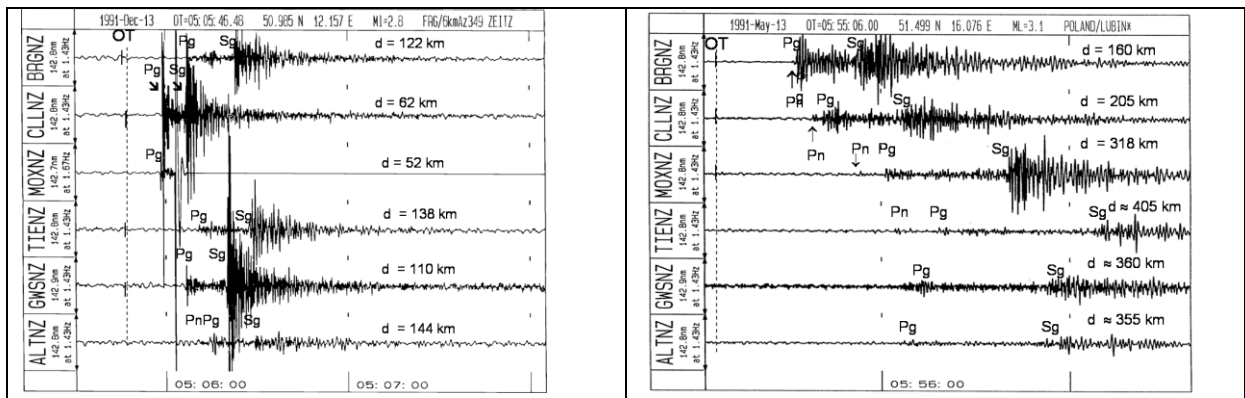
**Fig. 11.32** Illustration of the main principles on which graphical location methods are based.

EX 11.1 in this Manual aims at epicenter location by the simple circle and chord method using records from seismic sources both within and outside of a local station network. It is suitable also as a practical for “Seismology at Schools”. First one tries to identify the dominating seismic phase arrivals in the local records of each station by matching them best with the local travel-time curves of Pg, Sg, Pn and/or Sn for a surface focus (Fig. 11.33). This yields at the same time an estimate of the hypocentral distances from the travel-time differences Sg-Pg for each station and of the origin time OT if the assumption of  $h = 0$  is correct (Fig. 11.34). However, when no local/regional travel-time curve is available one may also use as a first approximation the rules of thumb for  $\Delta t(\text{Sg-Pg})$  or  $\Delta t(\text{Sn-Pn})$  [see formulas (11.1-11.3)]. Then it is shown how one can estimate the likely epicenter position by drawing circles around the stations with the so determined station-hypocenter distances and by connecting the crossings of equal pairs of over-shooting circles with so-called straight-line chords (Fig. 11.35).

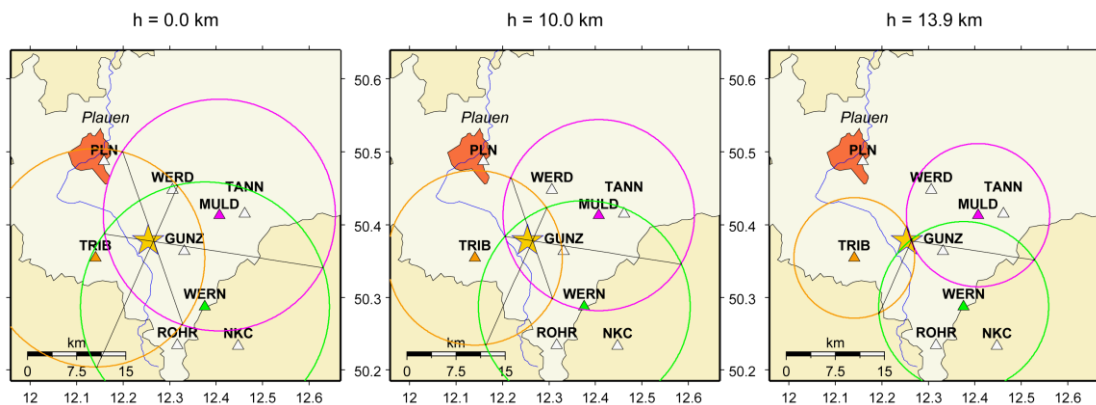
Finally, it is illustrated that the epicenter position, estimated from the “gravity” center of chord crossings, agrees rather well with the computer solution calculated with the common Geiger (1910 and 1912) algorithm and that additionally the source depth  $h$  can be estimated too by step-wise reduction of the circle radii until the circles all meet for the most likely assumed source depth  $h$  in the epicenter (Fig. 11.35). This iterative search for  $h$  with related modification of the depth-dependent local travel-time curve is also shown as a computer animation in IS 1.1.



**Fig. 11.33** Stepwise movement of a seismic record from left to right over a local travel-time curve for a surface focus until a best fit for three recognizable onsets (Pn, Pg, Sg) and the Lg wave group is achieved, here at a distance  $D \approx 327$  km.



**Fig. 11.34** **Left:** Identified local phases, estimated source origin time and station-source hypocenter distances for an earthquake in the distance range where Pg is the first arrival; **Right:** the same for another seismic event outside of the network in the distance range where Pn is the first arrival ( $d > \text{about } 150$  km).



**Fig. 11.35** Sequence of circle & chord plots for an earthquake in the Vogtland swarm earthquake region, Germany. Assumed are different source depths for fixed hypocentral distances  $d_i$  of the  $i$  stations as calculated from their S-P time differences. For  $h = 13.9$  km the circles cross almost ideally in one point at the epicenter. For details see IS 1.1 and EX 11.1.

Although the results of graphical location principally compare reasonably well with computer assisted locations (see EX 11.3), with the general availability of computers it is nowadays only seldom used in the observatory practice of seismic networks. With digital multi-station data and advanced seismogram analysis software, source location calculation seems to become almost a trivial and very fast completed task. One just picks a sufficient number of first arrival times, activates the relevant location program for local, regional and/or teleseismic sources and gets the result of the iteration process in an instant. Nevertheless, in practice, large location errors or even completely wrong locations can be obtained if certain conditions of the seismic network configuration and the quality of the picked phases are not fulfilled for a certain seismic event. Moreover, up to now most or the location algorithms are run by assuming 1D velocity models. So calculated location errors are not a reliable measure of the true location accuracy. Velocity heterogeneities in real Earth may result in location errors up to several 10 km from the true source, respectively hypocenter location (e.g., Engdahl et al., 1998). These real location errors are the larger the larger the azimuthal gaps are in the station distribution (e.g., Bormann, 1972a and IS 8.5). Reliable assessment of the actual location errors of seismic networks and their used procedures requires comparison with ground truth (GT) data of seismic events with precisely known locations and origin time, such as large chemical or nuclear explosions or precision located events by means of dense local networks with location errors less than about 5 to 10 km (see IS 8.6). For details on location with computer programs and related problems see IS 11.1 and for the program package HYPOSAT/HYPOMOD the PD 11.1.

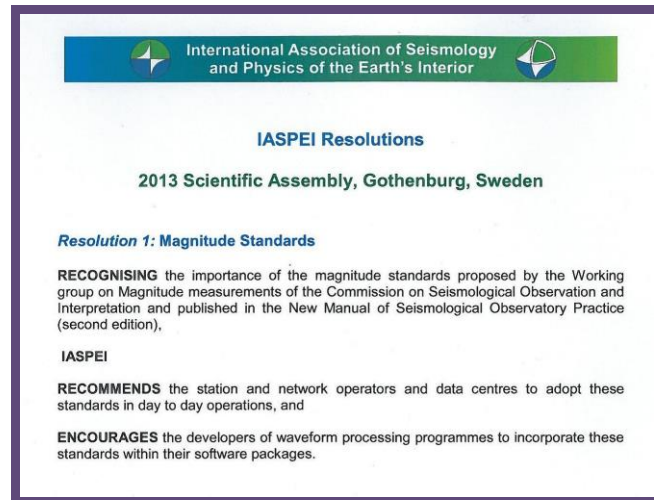
The accuracy of location, particularly of the source depth, can be significantly improved when at least one of the stations is near to the epicenter at  $D < h$  and/or by picking besides P-wave first arrivals also later arrivals which permit a much better distance and depth control than P-wave slowness data alone. Examples for both local and teleseismic event locations based on seismic network and array data are given in the following sections. IS 11.6 demonstrates the use of the SEISAN software for phase picking, event location and magnitude determination in the local, regional and teleseismic distance range based on seismic records from national and global seismic stations, as well as a 3-component earthquake location. Location using array data is described in Chapter 9, together with the underlying theory. Problems related to the optimization of network configuration aimed at best location results are discussed in detail in IS 8.7 and the PPT tutorial linked to this information sheet.

### **11.2.8 Magnitude determination**

When epicentral distance and depth of a seismic source are (at least roughly) known the magnitude of the event can be estimated. The general procedures to be followed in magnitude determination based on the measurement of amplitudes, periods and/or record duration as well as the specifics of different magnitude scales to be used for local, regional or teleseismic recordings are dealt with in detail in section 3.2 of Chapter 3 on more than 100 pages, including also most recent, still experimental and not yet standardized scales. DS 3.1 presents various classical magnitude calibration functions as diagrams or tables together with other auxiliary material. Some of these classical procedures of magnitude determination can be learnt from EX 3.1, which also gives solutions for the different tasks.

In this context reference has to be made to the most recent

IASPEI resolution on magnitude standards (Fig. 11.36). It relates to classical short-period ML and mb, the traditional NEIC 20s surface-wave magnitude Ms and, most importantly, to the new velocity broadband body and surface-wave standards mB\_B and Ms\_BB. For calculating the moment magnitude Mw from seismic moment M<sub>0</sub> a standard formula has now been fixed too.



**Fig. 11.36** IASPEI resolution 2013 on magnitude measurement standards.

The most recent short-version of the new IASPEI (2013) magnitude measurement standards and nomenclature, IASPEI (2013), can be read and downloaded from the IASPEI website via

[http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf).

IS 3.3 elaborates in detail both on the standard procedures and their comparison with still common classical routines of magnitude determination and results. It also presents the partially modified nomenclature for standard magnitudes and related amplitude measurements and discusses admissible deviations in procedural details, especially in the filter responses to be applied for ML, mb and Ms<sub>20</sub> measurements, provided that the results for identical input data do not differ on absolute values average by more than 0.1 magnitude units.

Possible modifications of recommended standards of amplitude and period measurements in fully automatic procedures are discussed in section 3.2.3.2 of Chapter 3, showing their agreement with results of expert interactive magnitude determinations that strictly adhered to the recommended standards and discussing the reasons for partial disagreement and ways for improving currently available algorithms.

IS 3.4 provides a IASPEI standard magnitude reference data set together with guidelines how to use it for comparison and documentation of currently used magnitude procedures at seismological stations or network analysis centers with the new standards. In this context it is of great importance that all contributors of amplitude and period data for magnitude determination to international data centers, the International Seismological Centre (ISC) in particular, document and report to the ISC in detail their current, or, after they have already introduced the new standards, also their past procedures. This information is crucial for reliable assessment of procedural differences, related systematic biases or random error ranges, respectively of the achieved status of data compatibility and homogeneity and thus of a new level of reliability and representativeness of magnitude data for application and research. A related questionnaire, elaborated by the IASPEI Working on Magnitude Measurement, is attached as Annex 2 to IS 3.4.

Note that the widely used data analysis software packages SEISAN, SHM and SeisComp3 have by 2013 already implemented the related IASPEI recommended simulation filters as default or as auxiliary measurement routines.

## 11.3 Routine signal processing of digital seismograms

Standard analysis includes all data pre-processing and processing operations for the interpretation and inversion of broadband seismograms. Important time-domain processes are signal detection, signal filtering, restitution and simulation, phase picking, polarization analysis as well as beamforming and vespagram analysis for arrays. In the frequency domain the main procedures are frequency-wavenumber (f-k) and spectral analysis. Array-techniques as f-k and vespagram analysis, slowness and azimuth determination for plane waves, and beamforming are discussed in detail in Chapter 9 but a few examples are also shown below. Fourier spectral analysis can be applied to determine the frequency content of transient seismic waves, and power density spectral analysis for estimating the frequency content of seismic noise (Chapter 4 and section 7.2 of Chapter 7).

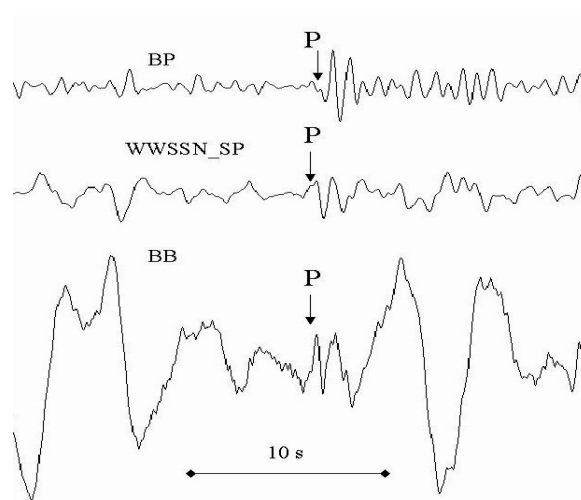
### 11.3.1 Signal detection

The first task of routine data analysis is the detection of a seismic signal. A signal is distinguishable from the seismic background noise either on the basis of its larger amplitudes or its differences in shape and frequency. Various methods are used for signal detection. Threshold detectors and frequency-wavenumber analysis are applied to the continuous stream of data. In practice, the threshold is not constant but varies with the season and the time of the day. For this reason, the threshold detectors determine the average signal power in two moving time windows: one long term (LTA) and one short term (STA). The ratio of the STA to LTA corresponds to the signal-to-noise-ratio (SNR). For details on the STA/LTA trigger and its optimal parameter setting see IS 8.1 and Chapter 9.

BB records are usually filtered before detectors are used although according to Douglas (1997) a detection algorithm is the better the less filtering is required. Useful filters are Butterworth high-pass filters with corner frequencies  $f_c > 0.5$  Hz or standard band-pass types with center frequency  $f = 1$  Hz for teleseismic P waves and high-pass filter with  $f_c > 1$ -3 Hz for local sources. Fig. 11.37 demonstrates detection and onset-time measurement for a weak, short-period P wave. In the lowermost 30 s segment of a BB-velocity seismogram the oceanic microseisms dominate in the period band 4-7 s. The two other traces are short-period seismograms after narrow band-pass (BP) filtering with: (1) a filter to simulate a WWSSN-SP seismogram; and (2) a two-step Butterworth BP filter of 2<sup>nd</sup>-order with cut-off frequencies of 0.7 and 2 Hz, respectively. The latter filter produces, for the noise conditions at the GRF-array, the best SNR for teleseismic signals. Seismic networks designed to detect mainly local seismic events may require other filter parameters that better account for the difference between local event and local noise frequencies. For optimal detection (see IS 8.1).

Generally, a seismic signal is declared when the STA/LTA exceeds a pre-set threshold. Various procedures, some analytical and some based on personal experience, are used to differentiate between natural earthquakes, mining-induced earthquakes and different kinds of

explosions. Usually, the detected signals are analyzed for routine parameter extraction and data exchange.

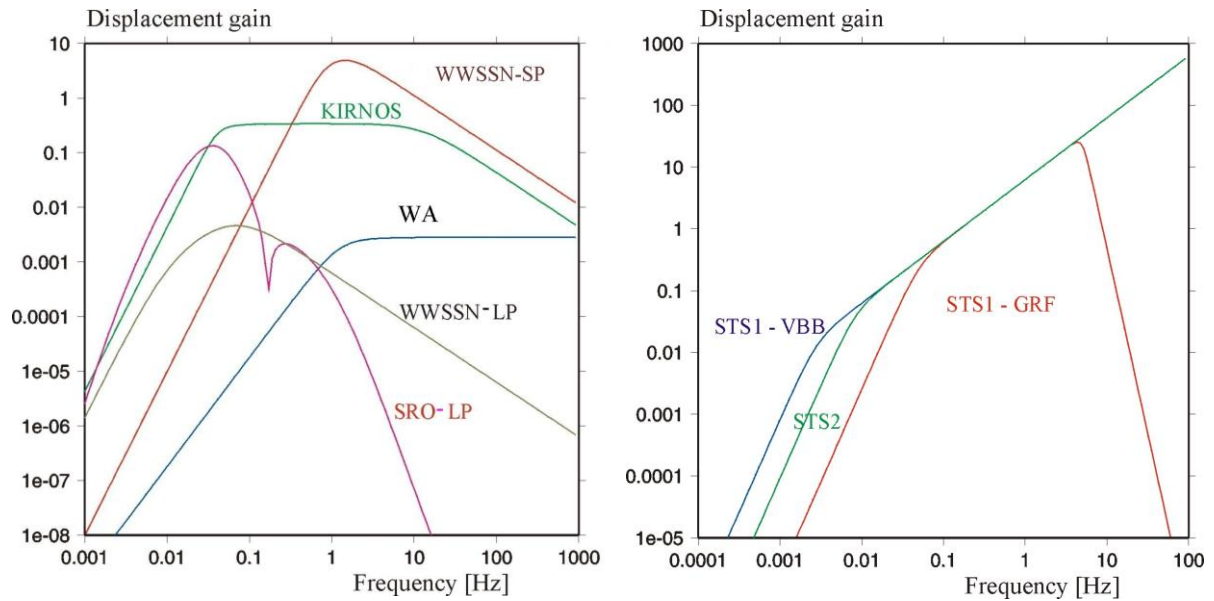


**Fig. 11.37** Bandwidth and SNR: A small short-period P-wave arrival below the noise amplitude level on a BB-velocity record (lower trace) may still be recognized by an experienced record analyst because of its much shorter period but missed by a simple amplitude threshold STA/LTA detector. However, by applying a WWSSN-SP simulation filter (middle trace) or a Butterworth band-pass filter (BP; uppermost trace) the SNR can be significantly improved. While the amplitude-based SNR is 0.2 on the original BB record and about 1 on the WWSSN-SP filter it is about 2 on the BP-filtered trace. (Record of an earthquake in the Kurile Islands on 25 March 2002, 6:18:13 UT, recorded at station GRA1, Germany).

### 11.3.2 Signal filtering, restitution and simulation

Classical broadband seismographs, such as the Russian Kirnos SKD, record ground displacement with constant magnification over a bandwidth of 2.5 decades or about 8 octaves (Fig. 11.38 left). The IDA-system (International Deployment of Accelerometers), deployed in the 1970s, used originally LaCoste-Romberg gravimeters for recording long-period waves from strong earthquakes proportional to ground acceleration over the band from DC to about 0.1 Hz (nowadays replaced by STS1-VBB; Fig. 11.38 right). Modern strong-motion sensors such as the Kinematics Inc. Episensor ES-T have a flat response to ground acceleration in an even broader frequency band from DC to 200 Hz. In contrast, feedback-controlled BB sensors for recording weak-motion usually have a flat response proportional to the ground velocity (Fig. 11.38 right). Such BB recordings, however, are often not suitable for direct visual record analysis and parameter extraction in the time domain. Low-frequency signals and surface waves of weak earthquakes are not or only poorly seen. Therefore, BB data must be transformed by applying digital filters in a way that yields optimal seismograms for specific investigations and analysis.





**Fig. 11.38 Left:** Displacement amplitude response characteristics of classical seismographs; **right:** The same for velocity broadband seismographs STS1(GRF) (old version as used at the Gräfenberg array), STS1 (VBB) (advanced version as used in the IRIS global network) and STS2. For STS1 (VBB) and STS2 the detailed response at higher frequencies is not shown (see comment below Tab. 11.4). The classical responses plotted on the left can be simulated with digital data from these broadband systems. Note the different scale of the vertical axes.

For some research tasks and ordinary routine analysis of BB seismograms the application of high-pass, low-pass and band-pass filters is usually sufficient. However, simultaneous multi-channel data processing or the determination of source parameters according to internationally agreed standards (such as body- and surface-wave magnitudes, which are defined on the basis of former analog band-limited recordings) often require simulation of a specific response, including those of classical analog seismograph systems (Seidl, 1980). Another special problem of simulation is “restitution”. Restitution is the realization of a seismograph system whose transfer function is directly proportional to ground displacement, velocity or acceleration in the broadest possible frequency range. The restitution of the true ground displacement down to (near) zero frequencies is a precondition for seismic moment-tensor determinations both in the spectral and the time domain (e.g., signal moment; Fig. 11.6). It is achieved by extending the lowermost corner frequency of the seismometer computationally far beyond that of the physical sensor system. Both the simulation of arbitrary band-limited seismograph systems and the extreme broadband “restitution” of the true ground motion are necessary steps in pre-processing of digital BB data.

Simulation is the mapping of a given seismogram into the seismogram of another type of seismograph, e.g., those of classical analog recordings such as WWSSN-SP, WWSSN-LP, Kirnos SKD, SRO-LP, and Wood-Anderson (WA). Up to now, amplitudes and periods for the determination of body- and surface-wave magnitudes  $m_b$  and  $M_s_{20}$  are measured on simulated WWSSN-SP and WWSSN-LP or SRO-LP seismograms, respectively, and the maximum amplitude for the original local Richter magnitude is measured on Wood-Anderson simulated seismograms. Fig. 11.38 (left) depicts the displacement response of these classical seismographs with smaller bandwidth.



The possibility of carrying out these simulations with high accuracy and stability defines the characteristics that have to be met by modern digital broadband seismograph:

- large bandwidth;
- large dynamic range;
- high resolution;
- low instrumental seismometer self-noise (Chapter 4, section 4.3.2 and Chapter 5, section 5.6.2);
- low noise induced by variations in air pressure and temperature (Chapter 5, sections 5.3.4, 5.3.5, and 7.4.4: for seismometer shielding see IS 5.4);
- analytically exactly known transfer function (see 5.2).

Fig. 11.38 (right) depicts the displacement responses of a few common BB-velocity sensors such as:

- the original Wielandt-Streckeisen STS1 with a bandwidth of 2 decades between the 3-dB roll-off points at frequencies of 0.05 Hz and 5 Hz (anti-aliasing filter). These seismographs were deployed in the world's first broadband array (GRF) around Gräfenberg/Erlangen in Germany (Fig. 11.3a);
- the advanced STS1 that is generally used at the global IRIS network of very broadband (VBB) stations (velocity bandwidth of about 3.3 decades between 5 Hz and 360 s; see also DS 5.1);
- the STS2 seismographs (see DS 5.1) that are usually operated in the frequency range between 0.00827 Hz and 40 Hz (velocity bandwidth of 3.7 decades or about 12 octaves, respectively). They are used at the stations of the GRSN (Fig. 11.3a), deployed world-wide at stations of the GFZ GEOFON network (see <http://www.gfz-potsdam.de/geofon/>) and at many other networks or single stations and have meanwhile also replaced the more narrowband original STS1 at the stations of the GRSN.

All these seismographs can be considered to be linear systems within the range of their usual operation. The transfer function  $H(s)$  of a linear system can be calculated from its poles and zeros by using the following general equation:

$$H(s) = N * \prod (s - z_i) / \prod (s - p_k) \quad (11.5)$$

where  $N$  is the gain factor,  $s = j\omega$  with  $\omega = 2\pi f$  and  $j$  the complex number  $\sqrt{-1}$ ,  $z_i$  are the zeros numbering from  $i = 1$  to  $m$  and  $p_k$  the poles with  $k = 1$  to  $n$ . Zeros are those values for which the numerator in Eq. (11.5) becomes zero while the poles are the values for which the denominator becomes zero.

Tab. 11.3 summarizes the poles and zeros of the classical standard responses WWSSN-SP, WWSSN-LP, WA (Wood-Anderson), Kirnos SKD and SRO-LP which control the shape of the response curves and Tab. 11.4 gives the poles and zeros for the three broadband responses shown in Fig. 11.38 on the right. Not given are the gain factors because they depend on the specific data acquisition system and its sensitivity.

**Note:** The poles and zeros in this tables are for WWSSN-SP, WWSSN-LP, WA (Wood-Anderson) close but not identical with those of the respective standard responses of these seismographs as now recommended by the IASPEI-CoSOI Working Group on Magnitude Measurements (Figure 1 and accompanying text in IS 3.3 ).

**Tab. 11.3** Zeros and poles corresponding to the displacement transfer functions depicted in Fig. 11.38 left for the classical analog standard seismographs WWSSN-SP, WWSSN-LP, WA, Kirnos SKD and SRO-LP (according to K. Stammer, personal communication 2013).

| Seismograph | Zeros                                                                                                                         | Poles                                                                                                                                                                                                                                                                                  |
|-------------|-------------------------------------------------------------------------------------------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| WWSSN-SP    | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0)                                                                                        | (-3.3678, -3.7315) (=p <sub>1</sub> )<br>(-3.3678, 3.7315) (=p <sub>2</sub> )<br>(-7.0372, -4.5456) (=p <sub>3</sub> )<br>(-7.0372, 4.5456) (=p <sub>4</sub> )                                                                                                                         |
| WWSSN-LP    | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0)                                                                                        | (-0.4189, 0.0)<br>(-0.4189, 0.0)<br>(-6.2832E-02, 0.0)<br>(-6.2832E-02, 0.0)                                                                                                                                                                                                           |
| WA          | (0.0, 0.0)<br>(0.0, 0.0)                                                                                                      | (-6.2832, -4.7124)<br>(-6.2832, 4.7124)                                                                                                                                                                                                                                                |
| Kirnos SKD  | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0)                                                                                        | (-0.1257, -0.2177)<br>(-0.1257, 0.2177)<br>(-80.1093, 0.0)<br>(-0.31540, 0.0)                                                                                                                                                                                                          |
| SRO-LP      | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0)<br>(-5.0100E+01, 0.0)<br>(-0.0, 1.0500)<br>(-0.0, -1.0500)<br>(0.0, 0.0)<br>(0.0, 0.0) | (-1.3000E-01, 0.0)<br>(-6.0200, 0.0)<br>(-8.6588, 0.0)<br>(-3.5200E+01, 0.0)<br>(-2.8200E-01, 0.0)<br>(-3.9300, 0.0)<br>(-2.0101E-01, 2.3999E-01)<br>(-2.0101E-01, -2.3999E-01)<br>(-1.3400E-01, 1.0022E-01)<br>(-1.3400E-01, -1.0022E-01)<br>(-2.5100E-02, 0.0)<br>(-9.4200E-03, 0.0) |

**Tab. 11.4** Zeros and poles corresponding to the displacement transfer functions of the velocity-proportional broadband seismographs STS1(GRF), STS1-VBB(IRIS) and STS2 of generation 1 as depicted in Fig. 11.38 right. From their output data seismograms according to the classical analog standard seismographs WWSSN-SP, WWSSN-LP, WA, Kirnos SKD and SRO-LP are routinely simulated at the SZO in Hannover, Germany (Table and comment below according to K. Stammer, personal communication 2013).

| Seismograph          | Zeros                                                                       | Poles                                                                                                                                                               |
|----------------------|-----------------------------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| STS2<br>generation 1 | (0.0,0.0)<br>(0.0,0.0)<br>(-15.15,0.0)<br>(-318.6,401.2)<br>(-318.6,-401.2) | (-3.674E-2, -3.674E-2)<br>(-3.674E-2, 3.674E-2)<br>(-15.99, 0.0)<br>(-417.1, 0.0)<br>(-100.9, 401.9)<br>(-100.9, -401.9)<br>(-7454.0, -7142.0)<br>(-7454.0, 7142.0) |
| STS2 - approx        | (0.0,0.0)<br>(0.0,0.0)<br>(0.0,0.0)                                         | (-3.674E-2, -3.674E-2)<br>(-3.674E-2, 3.674E-2)                                                                                                                     |

|            |                                        |                                                                                                                                                                                         |
|------------|----------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| STS1(GRF)  | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0) | (-0.2221, -0.2222)<br>(-0.2221, 0.2222)<br>(-31.416, 0.0)<br>(-19.572, 4.574)<br>(-19.572, -24.574)<br>(-7.006, 30.625)<br>(-7.006, -30.625)<br>(-28.306, 13.629)<br>(-28.306, -13.629) |
| STS1(VBB)) | (0.0, 0.0)<br>(0.0, 0.0)<br>(0.0, 0.0) | (-1.2341E-02, 1.2341E-02)<br>(-1.2341E-02, -1.2341E-02)                                                                                                                                 |

**Comment:** The detailed characteristics of the STS2 at high-frequencies depends on the production date. There are three different generations available, shown here is the first generation of STS2. In most cases high frequencies of 40 Hz or more are not relevant. Then a simple approximation of the STS2 transfer function, as STS2-approx in Tab. 11.4, is sufficient. When high frequencies are important and are close to the Nyquist frequency (e.g. for 100 Hz sampled data the Nyquist frequency would be 50 Hz) then also the characteristics of the digitizer has to be taken into account. In order to prevent aliasing digitizers one should apply steep low-pass filters just below the Nyquist frequency, usually represented as cascades of FIR filters with dozens of FIR coefficients per stage. Details on FIR filters can be found in Scherbaum and Bouin (2007).

Using the data given in these tables, the exact responses of the respective seismographs can be easily found. As an example, we calculate the response curve of the WWSSN-SP. According to Tab. 11.3 it has three zeros and four poles. Thus we can write Eq. (11.5) as

$$H(s) = N * s^3 / (s-p_1)(s-p_2)(s-p_3)(s-p_4) \quad (11.6)$$

with

$$\begin{aligned} p_1 &= -3.3678 - 3.7315j \\ p_2 &= -3.3678 + 3.7315j \\ p_3 &= -7.0372 - 4.5456j \\ p_4 &= -7.0372 + 4.5456j. \end{aligned}$$

The squared lower angular corner frequency of the response (that is in the given case the eigenfrequency of the WWSSN-SP seismometer) is  $\omega_l^2 = p_1 \cdot p_2$  whereas the squared upper angular eigenfrequency (which used to be in the classical SP records that of the galvanometer) is  $\omega_u^2 = p_3 \cdot p_4$ . Since the product of conjugate complex numbers  $(a + bj)(a - bj) = a^2 + b^2$  it follows for the poles:

$$\begin{aligned} \omega_l^2 &= 25.27 & \text{with } f_l &= 0.80 \text{ Hz} & \text{and} \\ \omega_u^2 &= 70.18 & \text{with } f_u &= 1.33 \text{ Hz}. \end{aligned}$$

When comparing these values for the corner frequencies of the displacement response of WWSSN-SP in Fig. 11.38 (left) one recognizes that the maximum displacement magnification (slope approximately zero) lies indeed between these two values. Further, as outlined in EX 5.6, a conjugate pair of poles such as  $p_1$  and  $p_2$  or  $p_3$  and  $p_4$  correspond to a second order corner of the amplitude response, i.e., to a change in the slope of the asymptote to the response curve by 2 orders. Moreover, the number of zeros controls the slope of the response curve at the low-frequency end, which is three in the case of the WWSSN-SP (see Eq. (11.6) and Tab. 11.3). Thus, at its low-frequency end, the WWSSN-SP response has according to its three zeros a slope of 3. This changes at the first pair of poles, i.e., at  $f_l = 0.8$

Hz, by 2 orders from 3 to 1 (i.e., to velocity proportional), and again at  $f_u = 1.33$  Hz by two orders from 1 to  $-1$ . This is clearly seen in Fig. 11.38.

In the same manner, the general shape of all the responses given in that figure can be assessed or precisely calculated according to Eq. (11.5) by using the values for the poles and zeros given in Tabs. 11.3 and 11.4. Doing the same with the values given in Tab. 11.3 for WWSSN-LP one gets for  $f_l = 0.06667$  Hz, corresponding to the 15 s seismometer and  $f_u = 0.009998$  Hz corresponding to the 100 s galvanometer, used in original WWSSN-LP seismographs. The aim of the exercise in EX 5.5 is to calculate and construct with the method shown above the responses of seismographs operating at several seismic stations of the global network from the data given in their SEED header information and EX 5.6 outlines this in detail for the calculation of the response of WWSSN-LP.

When comparing poles and zeros in Tab. 11.3 for WWSSN-SP with those given in Table 1 of IS 3.3 for the respective IASPEI standard response than one realizes that the former has besides the 3 zeros two conjugate poles whereas the latter has two conjugate and three single poles. Single poles mean, however, that at their respective frequency the response slope changes not by two but only by one order. But since the standard WWSSN-SP has three single poles instead of the second conjugate pole, the standard response (Figure 1 in IS 3.3) decays steeper towards higher frequencies than the WWSSN-SP response in Fig. 11.38 left. But the response slope towards longer periods, which is of third order and controlled by the number of zeros, is identical in both cases. The slight difference in response shape, however, results in a slight difference of amplitude-period measurements, which make mb(IASPEI) on average 0.033 m.u. smaller than mb(BGR) (Figure 11 in section 5.1.2 of IS 3.3). This is still acceptable within the IASPEI magnitude standard requirements as long as the absolute average differences for magnitudes of the same type but measured with different procedures is  $< 0.1$  m.u.).

Note that the poles and zeros given in Tabs. 11.3 and 11.4 are valid only if the input signal to the considered seismographs is ground displacement (amplitude  $A_d$ ). Consequently, the values in Tab. 11.3 are not suitable for simulating the responses of the classical seismographs if the input signal to the filter is not displacement. From the output of the STS2, any simulation filter gets as an input a signal, which is approximately velocity-proportional within the frequency range between 0.00827 Hz and 40 Hz. In this range its amplitude is  $A_v = \omega A_d$ . Accordingly, the transfer function of the simulation filter  $H_{fs}(s)$  has to be the convolution product of the inverse of the transfer function  $H_r(s)$  of the recording instrument and the transfer function  $H_s(s)$  of the seismograph that is to be simulated:

$$H_{fs}(s) = H_r^{-1}(s) * H_s(s). \quad (11.7)$$

Thus, even for the same  $H_s(s)$  to be simulated, the poles and zeros of the simulation filter differ depending on those of the recording seismograph. Tab. 11.5 gives, as an example from the SZ0, the poles and zeros of the displacement filters for simulating the responses shown in Fig. 11.38 (left), and the poles and zeros given in Tab. 11.4, from output data of the STS2-approx.

**Tab. 11.5** Poles and zeros of the simulation filters required for simulating standard seismograms of WWSSN-SP, WWSSN-LP, WA, Kirnos SKD and SRO-LP, respectively, from the SZO STS2-approx BB-velocity records (according to K. Stammli, personal communication 2013).

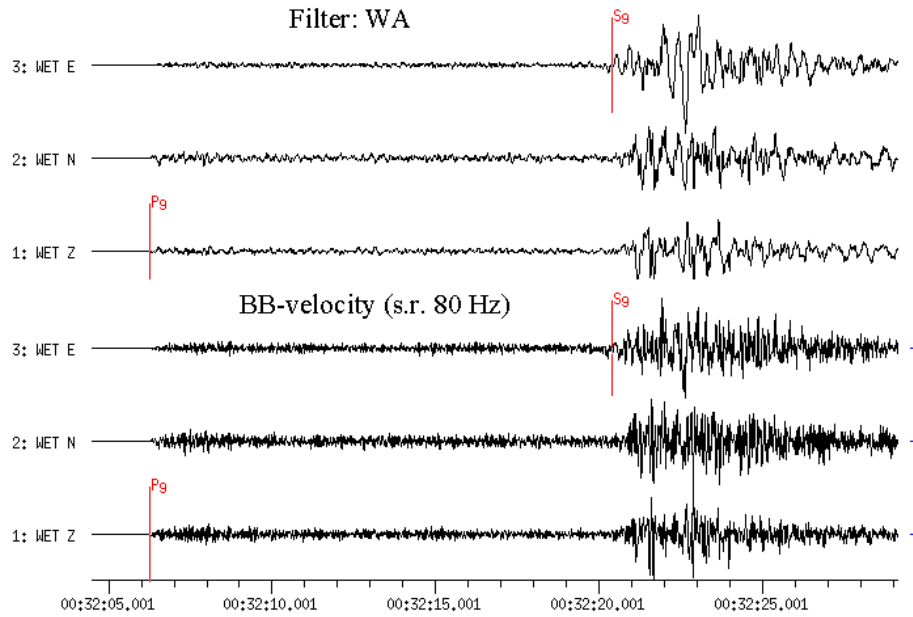
| Simulation filter | Zeros                                                                                                                                    | Poles                                                                                                                                                                                                                                                                                      |
|-------------------|------------------------------------------------------------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| WWSSN-SP          | (-3.6743E-02, -3.6754E-02)<br>(-3.6743E-02, 3.6754E-02)                                                                                  | (-3.3678, -3.7316)<br>(-3.3678, 3.7315)<br>(-7.0372, -4.5456)<br>(-7.0372, 4.5456)                                                                                                                                                                                                         |
| WWSSN-LP          | (-3.6743E-02, -3.6754E-02)<br>(-3.6743E-02, 3.6754E-02)                                                                                  | (-0.4189, 0.0)<br>(-0.4189, 0.0)<br>(-6.2832E-02, 0.0)<br>(-6.2832E-02, 0.0)                                                                                                                                                                                                               |
| WA                | (-3.6743E-02, -3.6754E-02)<br>(-3.6743E-02, 3.6754E-02)                                                                                  | (-6.2832, -4.7124)<br>(-6.2832, 4.7124)<br>(0.0, 0.0)                                                                                                                                                                                                                                      |
| Kirnos SKD        | (-3.6743E-02, -3.6754E-02)<br>(-3.6743E-02, 3.6754E-02)                                                                                  | (-0.12566, -0.2177)<br>(-0.1257, 0.2177)<br>(-80.1094, 0.0)<br>(-0.3154, 0.0)                                                                                                                                                                                                              |
| SRO-LP            | (-3.6744E-02, -3.6754E-02)<br>(-3.6743E-02, 3.6754E-02)<br>(-5.0100E+01, 0)<br>(-0, 1.0500)<br>(-0, -1.0500)<br>(0.0, 0.0)<br>(0.0, 0.0) | (-1.3000E-01, 0.0)<br>(-6.0200, 0.0)<br>(-8.6588, 0.0)<br>(-3.5200E+01, 0.0)<br>(-2.8200E-01, 0.0)<br>(-3.9301E+00, 0.0)<br>(-2.0101E-01, 2.3999E-01)<br>(-2.0101E-01, -2.3999E-01)<br>(-1.3400E-01, 1.0022E-01)<br>(-1.3400E-01, -1.0022E-01)<br>(-2.5100E-02, 0.0)<br>(-9.4200E-03, 0.0) |

If one wishes, however, to simulate accurately the WWSSN-SP response in Fig. 11.38 left only in the frequency range of interest for this type of seismograph, namely about 0.1 to 10 Hz in which the STS seismographs yield a perfect velocity-proportional output signal, then the simulation filter would just require one zero less but the same two conjugate poles as in Tab. 11.3.

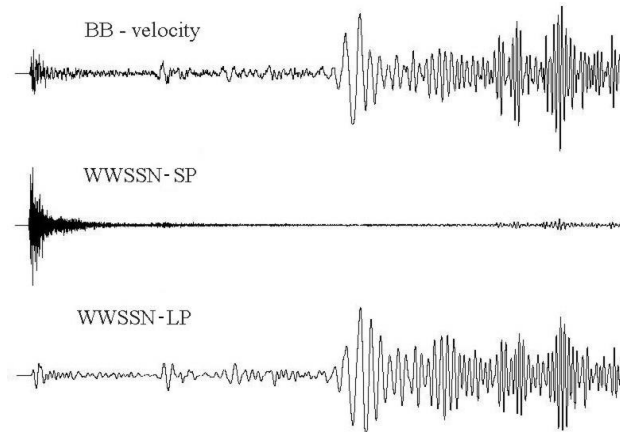
Fig. 11.39 shows a comparison of the original three-component BB-velocity record of an STS2 at station WET from a local earthquake in Germany with the respective seismograms of a simulated Wood-Anderson (WA) seismograph. For a teleseismic earthquake Fig. 11.40 gives the STS2 BB-velocity record together with the respective simulated records for WWSSN-SP and LP. Figs. 11.41 and 11.42 give two more examples of both record simulation and the restitution of very broadband (VBB) true ground displacement. VBB restitution of ground displacement is achieved by convolving the given displacement response of the recording seismometer with its own inverse, i.e.,:

$$H_{\text{rest}}(s) = H_s^{-1}(s) * H_s(s). \quad (11.8)$$

However, Eq. (11.8) works well only for frequencies smaller than the upper corner frequency (anti-alias filter!) and for signal amplitudes that are well above the level of ambient, internal (instrumental), and digitization noise.



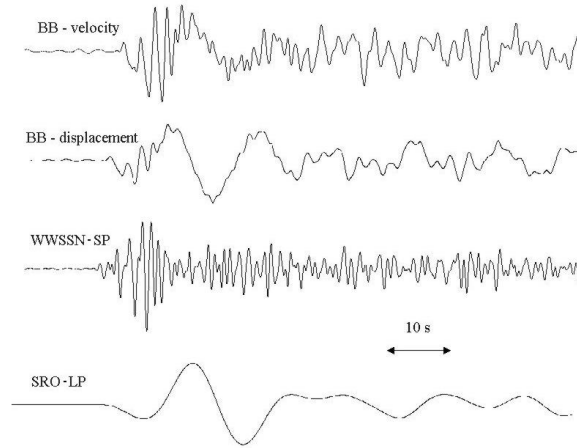
**Fig. 11.39** 3-component recordings at station WET (Wettzell) of a local earthquake at an epicentral distance of  $D = 116$  km. Lower traces: original STS2 records with sampling rate of 80 Hz; upper traces: simulated Wood-Anderson (WA) recordings. Note that the displacement-proportional WA record contains less high frequency oscillations than the velocity-proportional STS2 record (compare responses shown in Fig. 11.38).



**Fig. 11.40** BB-velocity seismogram (top) and simulated WWSSN-SP (middle) and WWSSN-LP seismograms (bottom). Note the strong dependence of waveforms and seismogram shape on the bandwidth of the simulated seismographs.

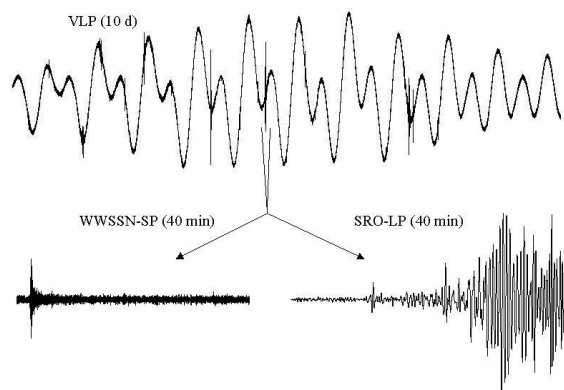
Both Fig. 11.39 and 11.40 clearly show the strong influence of differences in bandwidth and center frequencies of the seismometer responses on both the individual waveforms and the general shape of the seismogram. This is particularly obvious in the simulated teleseismic records of Fig. 11.41 showing the teleseismic P-wave group of an earthquake in California on 16 Sept. 1999. In the BB-velocity seismogram one recognizes clearly the superposition of a

low-frequency signal and a high-frequency wave group. The latter is clearly seen in the WWSSN-SP record but is completely absent in the SRO-LP simulation. From this comparison it is obvious that both the BB-velocity and the SP seismograms enhance short-period signal amplitudes. Therefore, only the former recordings are well suited for studying the fine structure of the Earth and determining the onset time and amplitude of short-period P waves. In contrast, BB-displacement seismograms and LP- filtered seismograms suppress the high-frequencies in the signals. Generally they are more suited to routine practice for surface-wave magnitude estimation and for the identification of most (but not all!) later phases (Figs. 11.16, 11.17, 11.48 and Fig. 2.52).



**Fig. 11.41** From top to bottom: The original BB-velocity seismogram recorded at station GRFO; the BB-displacement record derived by restitution; the simulated WWSSN-SP; and the simulated SRO-LP seismograms of the P-wave group from an earthquake in California (16 Sept. 1999;  $D = 84.1^\circ$ ;  $M_s = 7.4$ ).

Fig. 11.42 shows 10-days of a VBB record from an STS1 vertical-component seismograph (corner period  $T_c = 360s$ ) at station MOX and simulated WWSSN-SP and SRO-LP seismograms for a short (40 min) time segment of this VBB record.

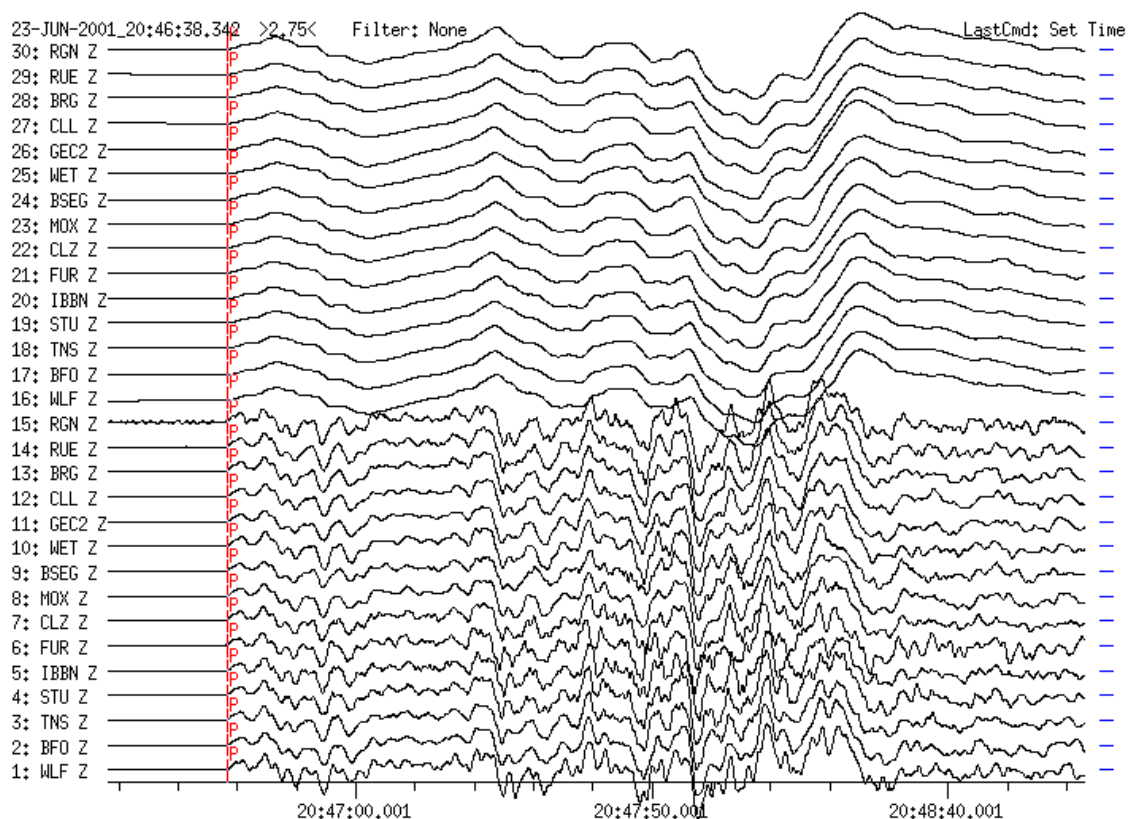


**Fig. 11.42** STS1 ( $T_c = 360s$ ) vertical-component seismogram with a length of 10 days (upper trace) as recorded at MOX station, Germany. In the seismogram we recognize Earth's tides and different earthquakes as spikes. For one of these earthquakes a WWSSN-SP and SRO-LP simulation filter was applied (lower traces). The length of the filtered records is 40 minutes.

Figs. 11.43a-d demonstrate, with examples from the GRSN, the restitution of (“true”) displacement signals from BB-velocity records as well as the simulation of WWSSN-SP, Kirnos BB-displacement and SRO-LP records. All traces are time-shifted for the P-wave group and summed (they are aligned on trace 16). The summation trace forms a reference seismogram for the determination of signal form variations. Generally, this trace is used for the beam (see 11.3.5 below). The different records clearly demonstrate the frequency dependence of the spatial coherence of the signal. Whereas high-frequency signals are incoherent over the dimension of this regional network (aperture about 500 to 800 km) this is not so for the long-period records which are nearly identical at all recording sites.

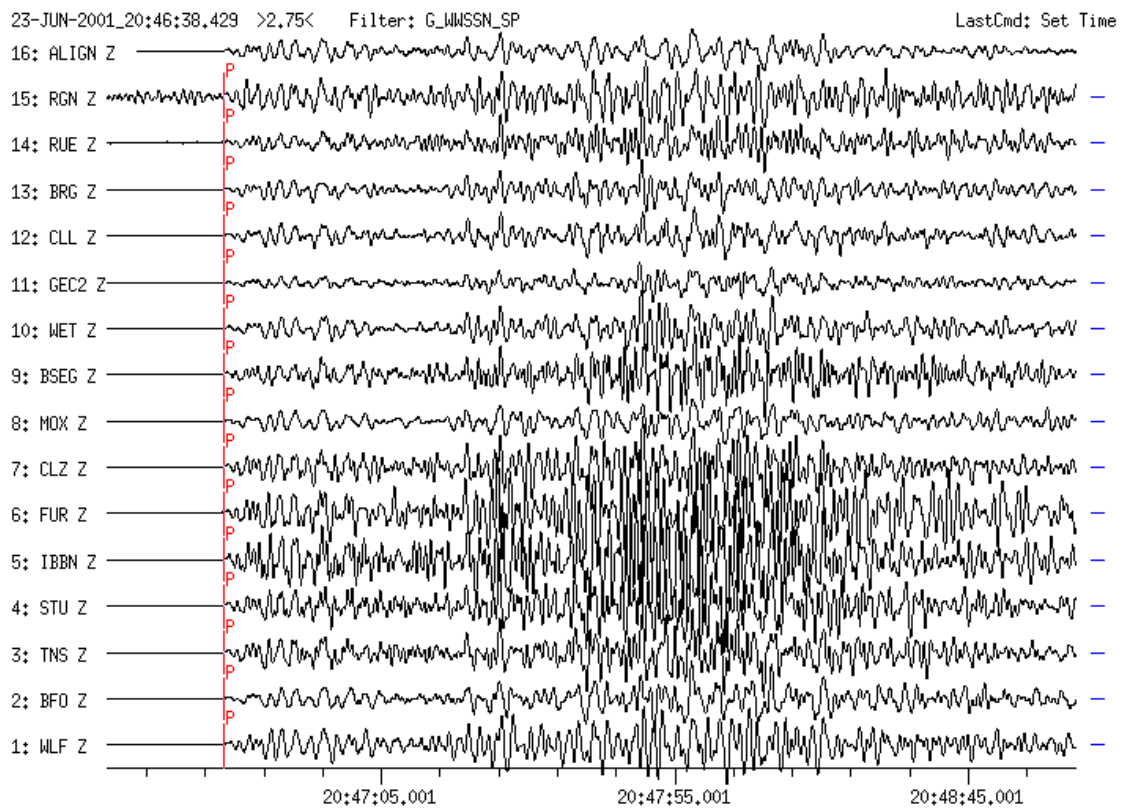
The following features are shown in Figs. 11.43a-d:

- a) Time shifted BB-displacement (traces 16-30) and BB-velocity seismograms (traces 1-15) with a duration of 145 s of the P-wave group from an earthquake in Peru on 23 June 2001 ( $M_s = 8.1$ ) as recorded at 15 stations of the GRSN. The BB-displacement seismogram suppresses the high-frequencies, which are clearly shown on the BB-velocity record.
- b) WWSSN-SP simulations for the same stations as in Figure 11.43a. The high-frequency signals are relatively enhanced by the short-period bandpass but the shape and amplitudes of the waveforms are shown to vary considerably within the network, i.e., the coherence is low.
- c) Kirnos SKD BB-displacement and d) SRO-LP simulations for the same stations as in Fig.11.43a. The high-frequency signals are masked. All traces show coherent waveforms.

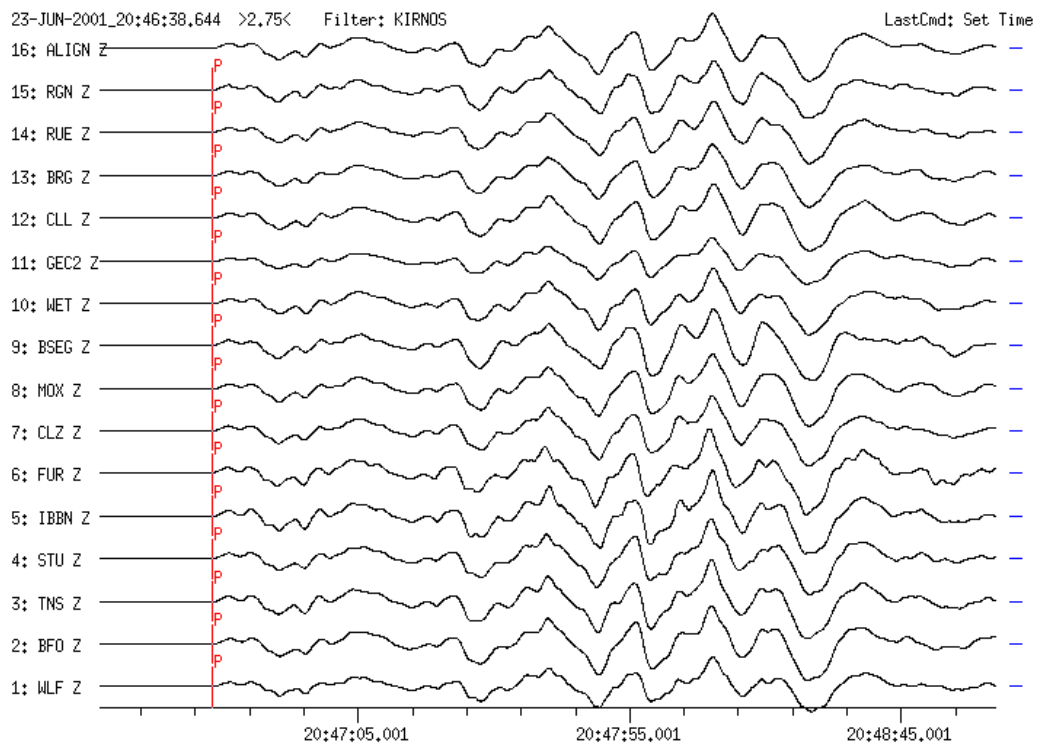


**Fig. 11.43a** (for explanation see preceeding text)

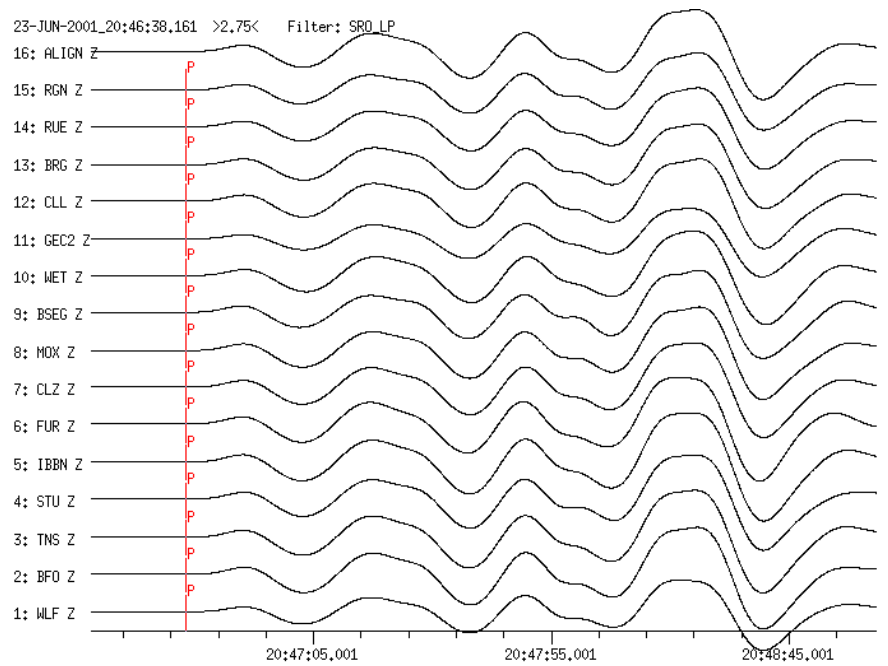




**Fig. 11.43b** As Fig. 11.43a however with short-period WWSSN-SP simulation.



**Fig 11.43c** As Fig. 11.43a but for displacement-proportional Kirnos SKD simulation.



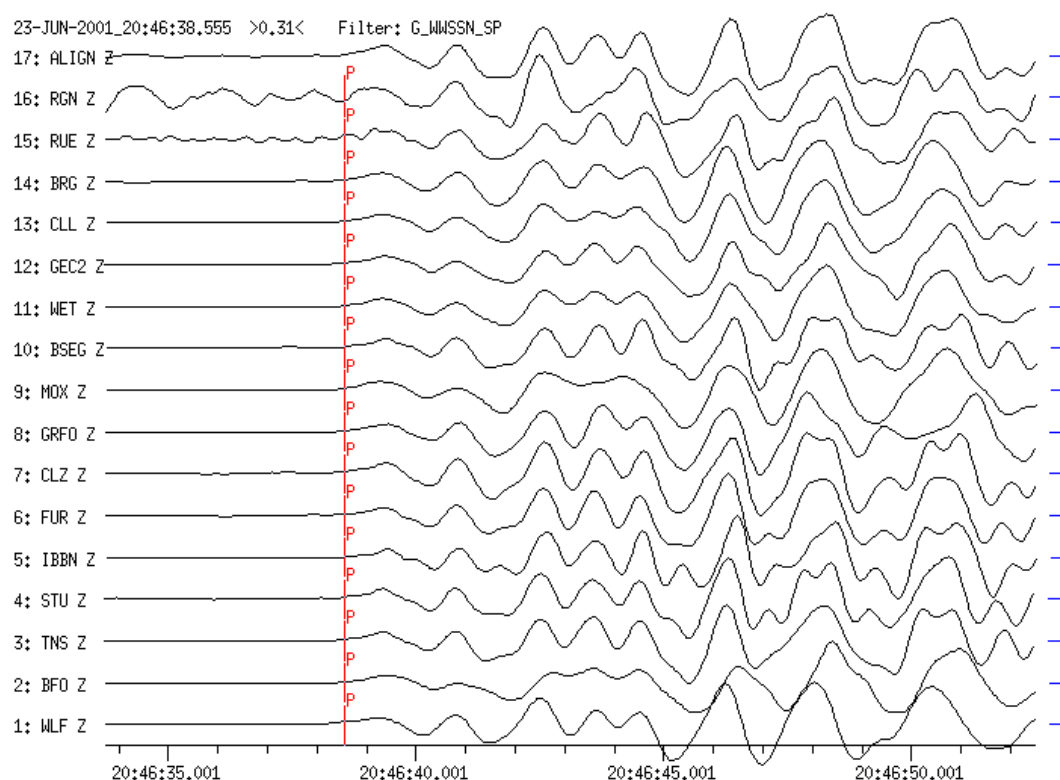
**Fig. 11.43d** As Fig. 11.43a but for long-period SRO-LP simulation.

**Fig. 11.43a-d** Restitution, simulation and coherency of seismograms demonstrated with records of the GRSN from an earthquake in Peru (23 June 2001,  $M_s = 8.1$ ) in the epicentral distance range from  $96^\circ$  to  $100^\circ$ ; for explanation see text).

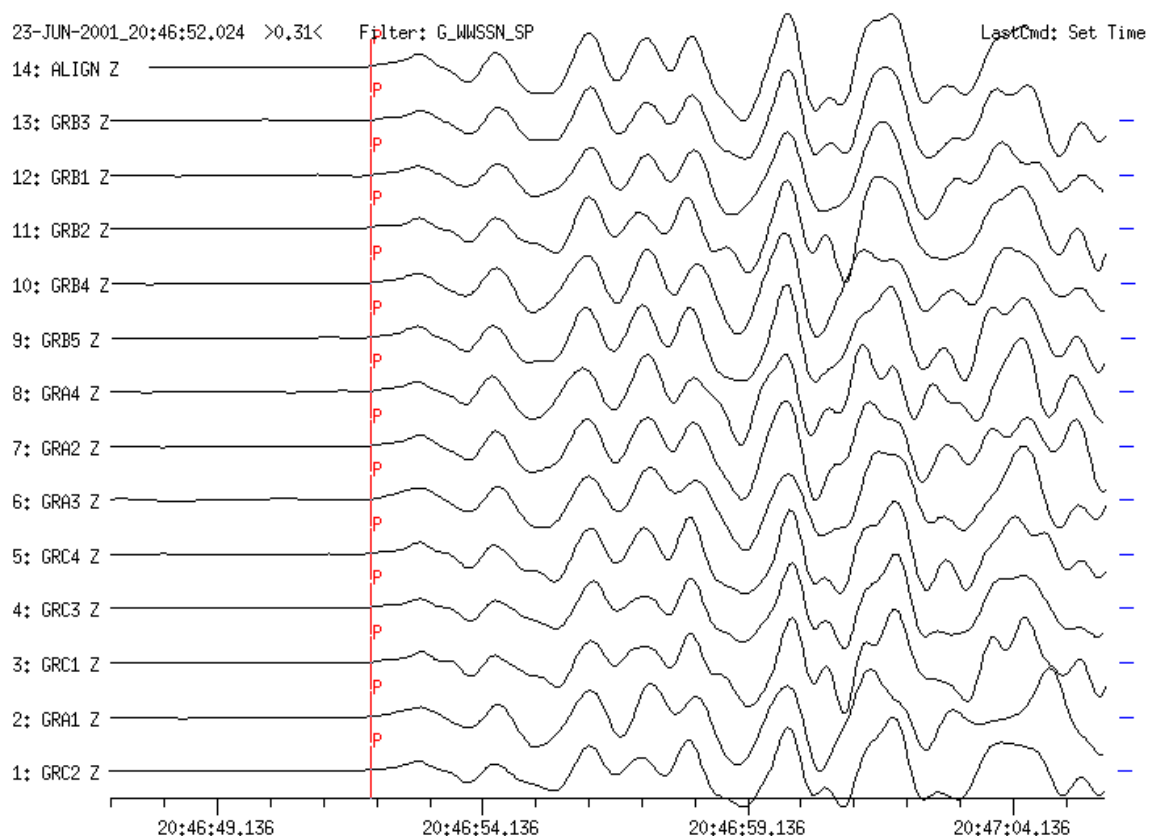
### 11.3.3 Signal coherency at networks and arrays

Heterogeneous crustal structure and the array aperture limit the period band of spatially coherent signals. The larger the aperture of an array the more rapidly the signal coherence falls off with frequency. At short periods the array behaves like a network of single stations whereas at long periods the array behaves like a sensitive single station. For the GRF-array (aperture about 50 to 120 km) for instance, the signals are coherent for periods between about 1 and 50 s. For the GRSN the band of coherent signals is at longer periods than for the small aperture detection arrays like GERES in Germany or NORES in Norway (aperture 4 and 3 km, respectively) where signals are coherent at periods shorter than 1 s. In the coherency band itself, waveforms vary depending on their dominant frequency, apparent horizontal velocity and azimuth of approach. For instance, coherent waveforms are observed from the GRSN for BB-displacement records, Kirnos SKD simulation and all long-period simulated seismograms (Figs. 11.43a-d) whereas for simulated WWSSN-SP seismograms the waveforms have low coherence or are incoherent.

Figs. 11.44a and b show a comparison of the first 14 s of a P-wave group recorded by the GRSN and the GRF-array. The coherence is clearly higher in the short-period range for the recordings at the smaller GRF-array than for the GRSN. The GRSN works as an array for periods longer than about 10 s but it is a network for shorter periods where the GRF-array works as an array down to periods of about 1 s. This discussion is valid for teleseismic signals only, where the epicentral distance is larger than the aperture of the station network or array.



**Fig. 11.44a** (Figure caption below)

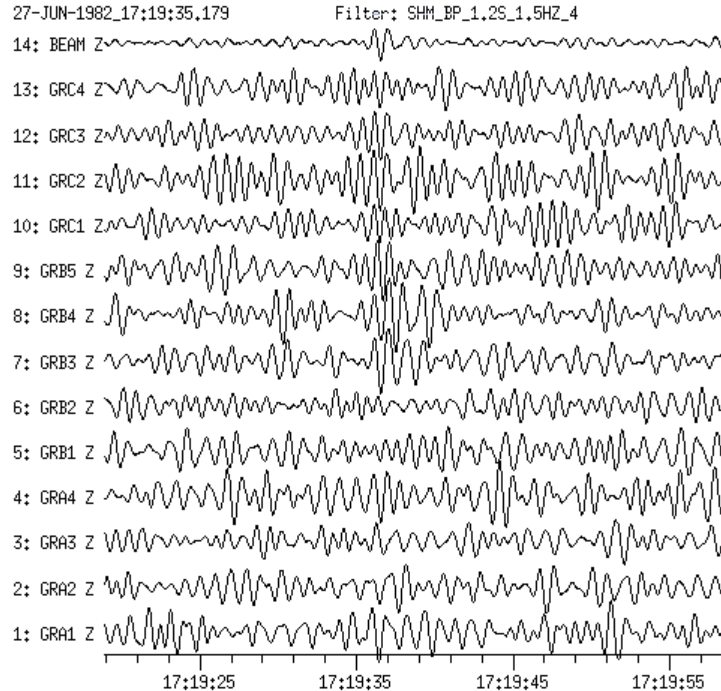


**Fig. 11.44b** (Figure caption below)

**Fig. 11.44** WWSSN-SP simulations of the first 14 s after the P-wave onset from the same Peru earthquake as in Fig. 11.43. **a)** Recordings at the GRSN. **b)** Recordings at the GRF array. Note the lower coherence of the waveforms recorded at the stations of the regional network, which has an aperture much larger than the GRF-array (Fig. 11.3a). The summation traces 17 and 14, respectively, are reference seismograms for the determination of signal waveform variations.

### 11.3.4 Beamforming

Beamforming improves the SNR of a seismic signal by summing the coherent signals from array stations (see 9.4.5). Signals at each station are time shifted by the delay time relative to some reference point or station. The delay time depends on ray slowness and azimuth and can be determined by trial and error or by f-k analysis (see 11.3.5). The delayed signals are summed “in phase” to produce the beam. In delay-and-sum beamforming with  $N$  stations the SNR improves by a factor  $\sqrt{N}$  if the noise is uncorrelated between the seismometers. In the summation the increase in amplitude of the coherent signal is proportional to  $N$ . For incoherent waves (random seismic noise in particular), it is only proportional to  $\sqrt{N}$ . Thus, beamforming and f-k analysis are helpful for routine analysis if very weak signals have to be detected and analyzed. Fig. 11.45 shows an example. The signal on the beam trace is the PKP wave of an underground nuclear explosion at Mururoa Atoll. The onset time and signal amplitude of a weak seismic signal can only be read on the beam. The peak-to-peak amplitude is only about 2 nm with a period around 1 s.

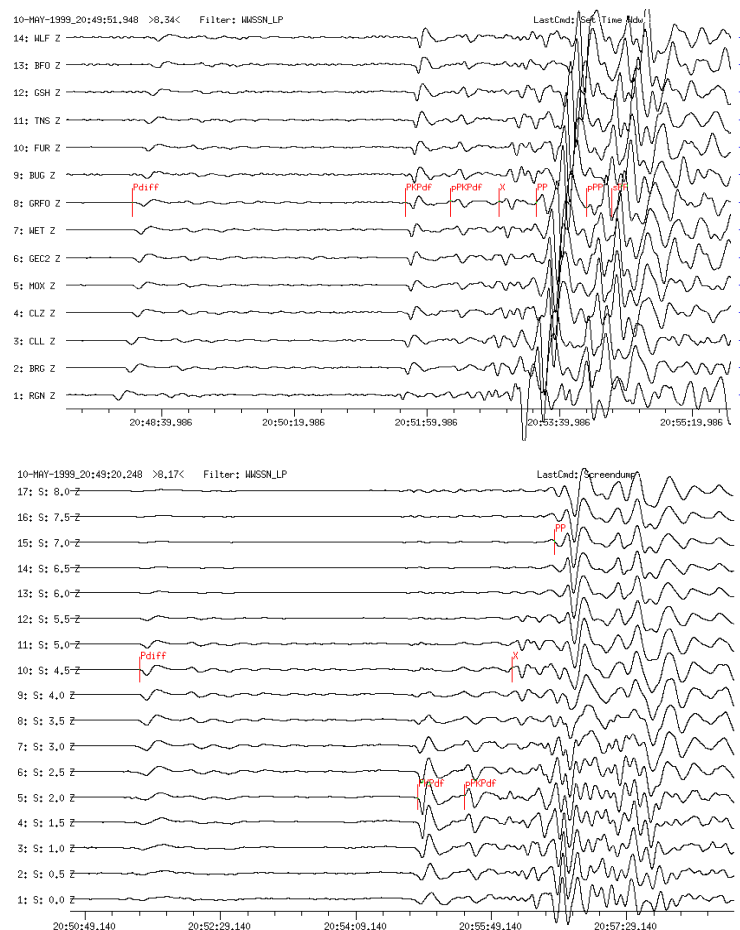


**Fig. 11.45** Detection of the PKP wave of a nuclear explosion at Mururoa Atoll on 27 June 1982 at the Gräfenberg array using the delay-and-sum-method and a very narrowband Butterworth bandpass filter (BP) centered around 1 Hz. The event occurred at an epicentral distance of  $146^\circ$  and the explosion yield was approximately 1 kt TNT.

### 11.3.5 f-k and vespagram analysis

Array-techniques such as f-k and vespagram analysis (Chapter 9) should be applied only to records with coherent waveforms. Vespagram analysis or the velocity spectrum analysis is a method for separating signals propagating with different apparent horizontal velocities. The seismic energy reaching an array from a defined backazimuth with different slownesses is plotted along the time axis. This allows identification of later phases based on their specific slowness values. The best fitting slowness is that for which a considered phase has the largest amplitude in the vespagram. Fig. 11.46 shows the original records from the GRSN (top) and the related vespagram (bottom). More vespagrams are given in Figures 12e–g of DS 11.2.

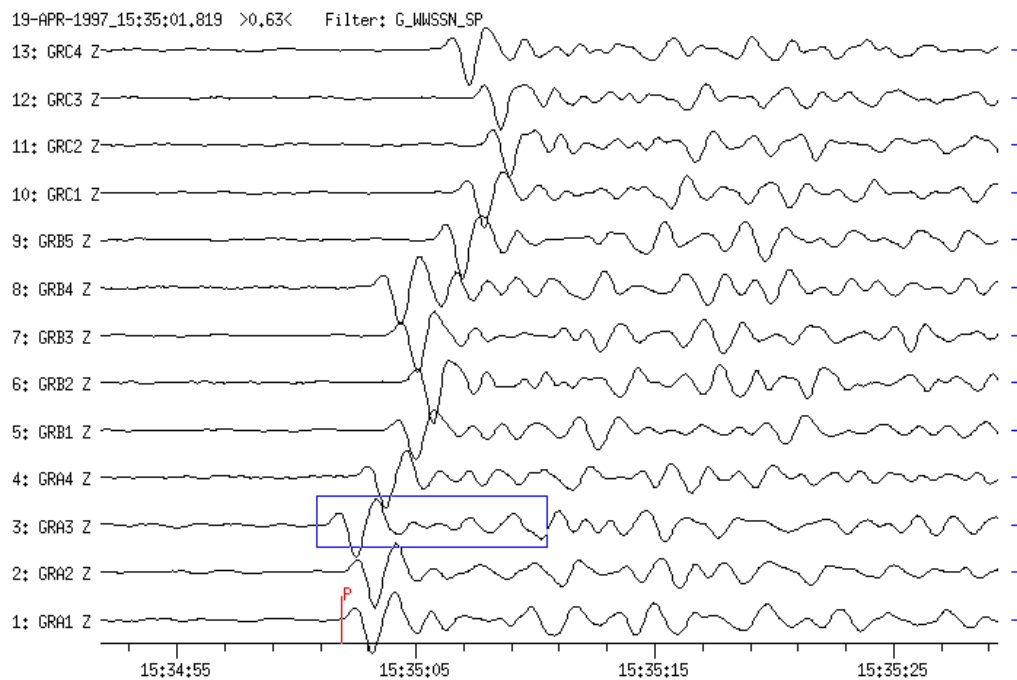
**Note:** The original vespagram or VESPA process was defined for plotting the observed energy from a specific azimuth for different apparent velocities as isolines versus time (Davies et al., 1971) and not beam traces with maximum amplitudes for maximum energy from a specific azimuth for different apparent velocities (or slowness) as here and in DS 11.2. For “true” vespagrams see Chapter 9, e.g., Fig. 9.30). But we feel that the trace beam presentation with different maximum amplitudes instead of the energy isolines eases the phase identification.



**Fig. 11.46 Top:** Simulated vertical-component WWSSN-LP seismograms from an earthquake in the region of Papua New Guinea. Source data NEIC-QED: 10 May 1999; depth 137 km; mb = 6.5; D = 124° to GRF, BAZ = 51°. The phases Pdif (old Pdif), PKPdf, pPKPdf, PP, pPP, sPP and an unidentified phase X have been marked. **Bottom:** Vespagram of the upper record section. The analysis yields slowness values of 4.5s/° for Pdif, 2.0s/° for PKPdf and pPKPdf, 7.0s/° for PP, and a value that corresponds to the slowness for Pdif for the unknown phase X.

The f-k analysis is used to determine slowness and backazimuth of coherent teleseismic wave groups recorded at an array. Fig. 11.47 shows an example. The epicentral distance must be much larger than the aperture of the recording array. The f-k analysis transforms the combined traces within a current time window (Fig. 11.47a) into the frequency-wavenumber domain. The result in the f-k domain is displayed in a separate window (Fig. 11.47b) with amplitudes (corresponding to wave energy) coded in color. A good result is achieved when there is a single, prominent color in the maximum (yellow in Fig. 11.47b). This maximum denotes slowness and backazimuth of the investigated phase and is helpful for source parameter determination and phase identification. The example was recorded at the GRF-array from an earthquake in East of Severnaya Zemlya. Slowness and backazimuth (BAZ) values are 7.39 s/° and 12.3°, respectively. These values have been used for producing the beam.

**Figs. 11. 47a-b (below)** Illustration of the procedure of frequency-wavenumber (f-k) analysis: a) coherent P-wave signals recorded at the GRF-array stations from an earthquake on 19 April 1997 (the box marks the time window selected for the f-k-analysis); b) energy (coded in colors) in the frequency range 0.39-2.97 Hz as a function of wavenumber k. A good result is achieved because the single, prominent maximum (in yellow) shows the presence of a coherent signal. The estimated slowness and BAZ values are 7.39 s/° and 12.3°, respectively.



**Fig. 11.47a**

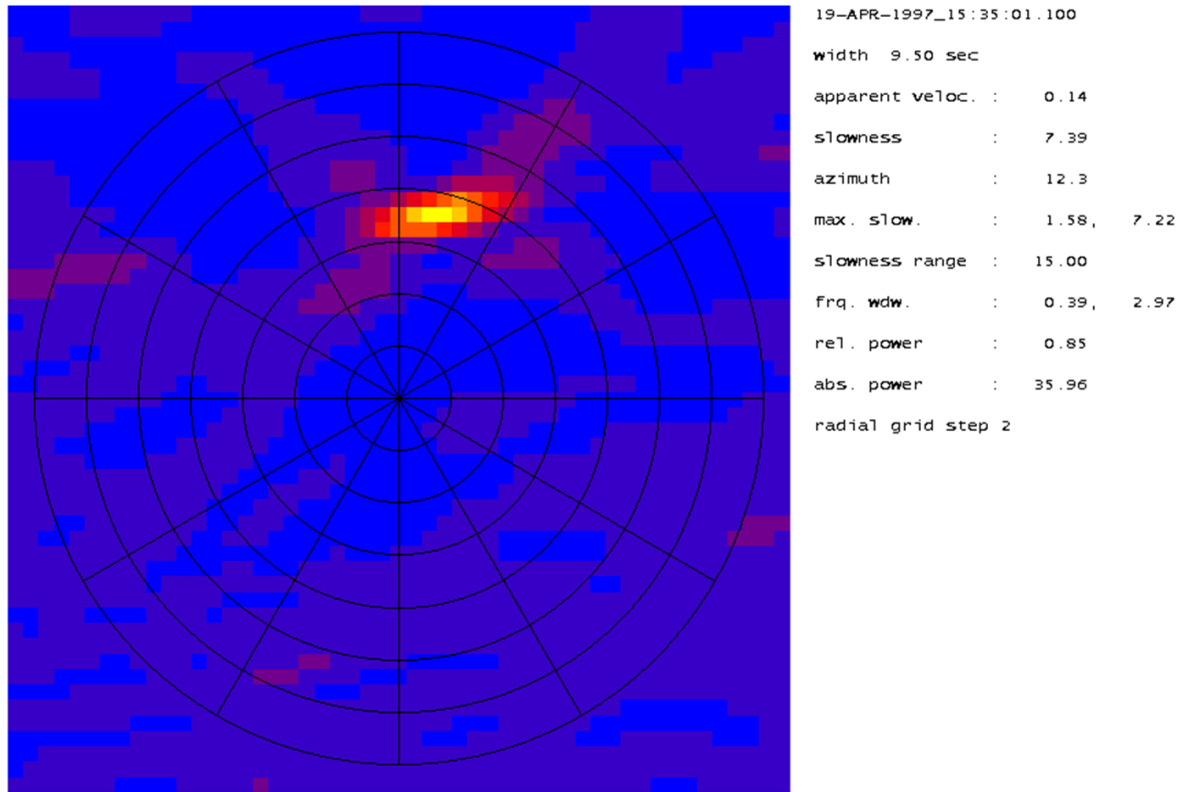


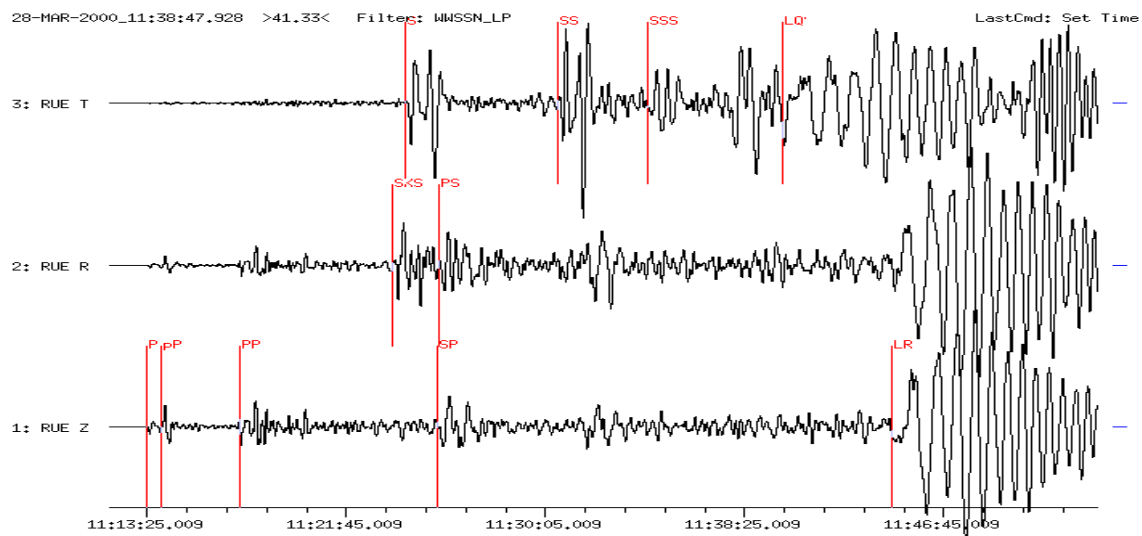
Fig. 11.47b

### 11.3.6 Polarization analysis

The task of polarization analysis is the transformation of recorded three component seismograms into the ray-oriented co-ordinate systems. For linearly polarized and single pulse P waves in a lateral homogeneous Earth, this task is simple, at least for signals with a high SNR; the direction of the polarization vector of the P wave clearly determines the orientation of the wave co-ordinate system. However, when propagating through heterogeneous and anisotropic media, the seismic waves have three-dimensional and frequency-dependent particle motions and the measured ray-directions may scatter by ten degrees and more about the great circle path from the epicenter to the station (Figs . 2.7 and 2.8 in Chapter 2). Accordingly, one may also describe the wave polarization by its attributes *rectilinearity*, *planarity* and/or the *largest eigenvalue* (e.g., Wagner and Owens, 1996). If the three main axes of particle motion in space are  $\lambda_1$ ,  $\lambda_2$  and  $\lambda_3$  than the rectilinearity is  $rect = 1 - (\lambda_2 + \lambda_3)/2\lambda_1$  and the planarity  $plane = 1 - 2\lambda_3/(\lambda_1 + \lambda_2)$  (Flinn, 1965). Their values may vary between 1 (ideal linearity or planarity) and 0.

Determination of particle motion is included in most of the analysis software. For identification of wave polarization and investigation of shear-wave splitting, the rotation of the traditional components N, E, and Z into either a ray-oriented co-ordinate systems (L, Q, T) or in the directions R (radial, i.e., towards the epicenter) and T (transverse, i.e., perpendicular to the epicenter direction) is particularly suitable for the identification of secondary later phases. An example for the comprehensive interpretation of such phases in a teleseismic record is given in Fig. 11.48.





**Fig. 11.48** Simulation of three-component WWSSN-LP seismograms of the Volcano Islands earthquake of 28 March 2000, recorded at station RUE in Germany ( $D = 94^\circ$ ,  $h = 119$  km). The horizontal components N and E are rotated into R and T components. The phases P, pP, SP and the beginning of the dispersed surface Rayleigh wave train LR are marked on the vertical-component seismogram, SKS, PS on the radial component (R) and S, SS, SSS and the beginning of the Love waves LQ on the transverse component (T), respectively. Not marked (but clearly recognizable) are the depth phases sS behind S, sSS behind SS, and SSSS+sSSSS in the T component before LQ. The plotted record length is 41min.

## 11.4 Software for routine analysis

### 11.4.1 Brief overview and software library links

There are many different software packages available for automatic and manual processing of earthquake recordings. Some are mainly suited for scientific analysis of the data, others permit the fast routine processing, which includes the generation of reports and maps as needed for the information to the authorities and the public about relevant seismic activity. The companies which sell seismic networks provide commercial software for these tasks. A well-known software of this type is *Antelope* ([www.brtt.com](http://www.brtt.com)). For a briefing Chapter 8, section 8.4.2.2. Other programs were created by university groups or seismological institutions to improve their work such as SHM, SEISAN, SeisComp3 and Earthworm. They were published under free or partially free licencing schemes and applied by other seismologists with so good results that they were later adopted by many observatories and network centers. For more details sections 11.4.2 to 11.4.5. Most of the analysis examples shown in this Chapter have been processed interactively by using the SHM and SEISAN software which belong to the latter type of software. Several software packages combine the fully automatic on-line and seismic data acquisition, archiving, integration, processing and interactive reprocessing such as the commercial SeisComP. Olivieri and Clinton (2012) published a detailed comparison of the features and performance of Earthworm and SeisComp3.



The ORFEUS Software Library (<http://orfeus.knmi.nl/>) presents a comprehensive list and links to available software in seismology, both for routine seismogram analysis and different research tasks. The focus is on shareware, but some relevant commercial sites are also included. Links to the USGS (<http://earthquake.usgs.gov/research/software/>) and the IRIS software libraries (<http://www.iris.edu/manuals/>) are included as well. The latter provides also detailed information about the Seismic Analysis Code (SAC; Goldstein et al, 2003; Goldstein and Snoke, 2005). SAC2000 is used by more than 400 institutions worldwide on a diversity of platforms. Its signal processing capabilities include data inspection, signal correction, and quality control; travel-time analysis and magnitude estimation; spectral analysis and spectrograms, array and three-component analysis. These features are widely used not only in earthquake, explosion and volcanic source studies and the investigation of Earth structure, wave path and site effects but also in the analysis of other geophysical data related to electromagnetic and hydroacoustic phenomena.

Available at the [Orfeus ftp site](#) is also the program *PREPROC* for preprocessing (filtering, restitution, simulation etc.) of seismic traces, seismograph calibration, and fast ray synthetic seismograms for body waves (Authors: Axel Plesinger, Miroslav Zmeskal and [Jan Zednik](#)).

An older, widely used interactive analysis program, *PCEQ*, for IBM compatible PCs, had been written by C. M. Valdés. It was published as Volume 1 of the IASPEI Software Library (Lee, 1995) and works with the location program HYPO71 for local events. Its principal features are: picking P- and S-wave arrivals; filtering the seismogram for better P- and S-wave picks, and computing the spectra of selected seismogram sections. Quite many analysis examples presented in this Manual, e.g., on event location, spectral and polarization analysis, had been processed with the rather comfortable and appealing software *SEIS89* (Baumbach †, 1999) which is, regrettably, no longer maintained because of the death of the author.

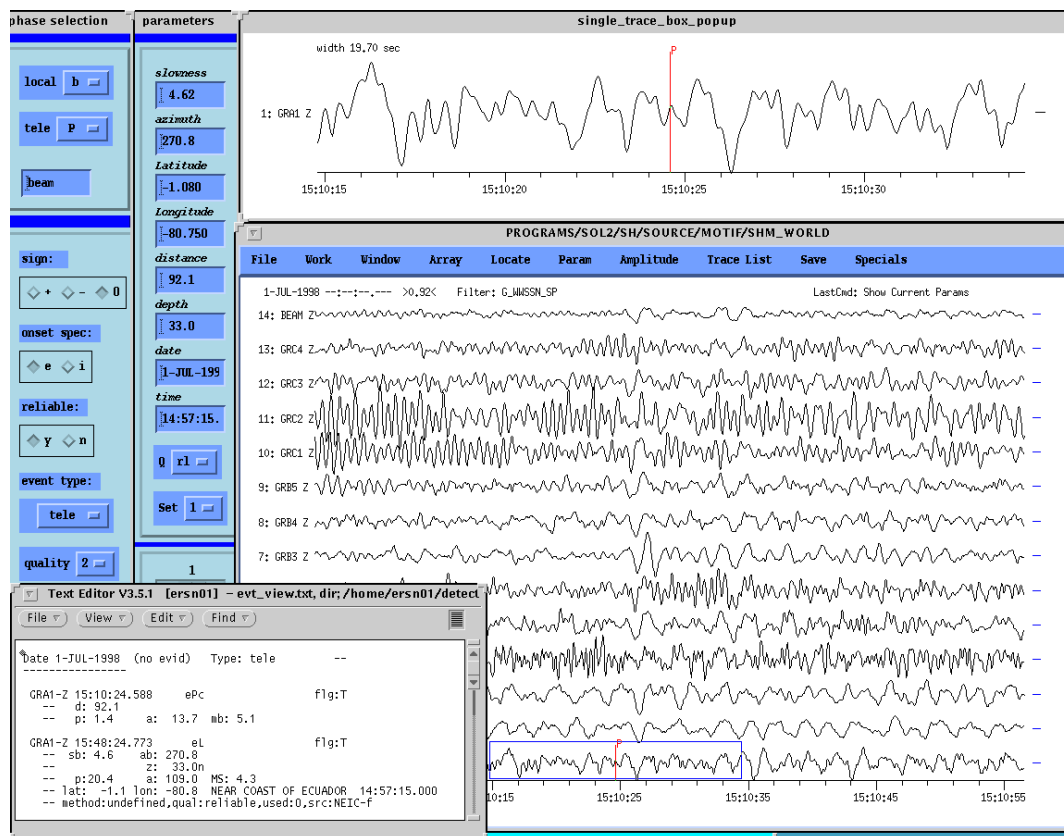
Andrey Petrovich Akimov wrote the program WSG (in English AWP: Automated workplace of seismologists). It works in an environment of Windows 95/98/NT and is used at single stations and seismic networks for estimating parameters from local, regional and teleseismic sources. The program converts different seismic data formats such as XDATA, PCC-1, CSS 2.8 and 3.0, DASS, CM6 GSE2 and can import via the TCP/IP protocol data from NRTS and LISS systems (miniSEED). Program and documentation in Russian for version 4.5 are available via [http://www.ceme.gsras.ru/engl/stations/wsg\\_arm.htm](http://www.ceme.gsras.ru/engl/stations/wsg_arm.htm).

Another relevant complementary software, named *Geotool*, has been developed at the CTBTO (Chapter 15) for interactively displaying and processing seismoacoustic data from the International Monitoring /system (IMS). A Software User Tutorial provides a set of exercises that cover the full range of functionality of the software system. It is available via [http://www.ctbto.org/fileadmin/user\\_upload/procurement/2013/Ap\\_D\\_Part1.pdf](http://www.ctbto.org/fileadmin/user_upload/procurement/2013/Ap_D_Part1.pdf), The tutorial covers, amongst others, filter procedures, display and aligning of waveforms, measurement and review of signal arrivals and amplitudes, particle motion analysis and component rotation, waveform analysis (Fourier Transform and spektrograms), f-k and cepstrum analysis, signal correlation, event location, production of bulletins, maps and map overlays.

The wide variety of computer platforms, file formats, and methods to access seismological data often requires considerable effort in preprocessing such data. Although pre-processing work-flows are usually very similar, few software standards exist to accomplish this task. *ObsPy* offers solutions to this problem (section 11.4.6).

## 11.4.2 SHM

The Seismic Handler Motiv, SHM, is a powerful program for analyzing local, regional and teleseismic recordings with station networks or arrays. It has been developed at the BGR (Federal Institute for Geosciences and Natural Resources) originally for the analysis of data from the Gräfenberg (GRF) array and the German Regional Seismic Network (GRSN) (author: Klaus Stammler; recent developments by Marcus Walther). Program and descriptions are available via <http://www.seismic-handler.org>. A screen display of SHM is shown in Fig. 11.49.



**Fig. 11.49** Screen display of the seismic analysis program SHM. Different windows display a number of station recordings (large window), a zoomed single-station window, two seismogram and source parameter windows (left side) and an output window for the results of

Main features of SHM are:

- application of array procedures to a set of stations (slowness- and backazimuth determination by means of beamforming and f-k analysis);
- location algorithms (teleseismic locations using travel-time tables and empirical correction vectors, local and regional locations via external programs, e.g., LocSAT).

The basic program has some (more or less) standardized options, e.g.:

- manual and automatic phase picking (Figs. 11.5 and 11.6);
- trace filtering with simulation and bandpass filters (Figs. 11.37 to 11.40);
- determination of amplitudes, periods and magnitudes (Fig. 11.4);
- display of theoretical travel times on the traces (Fig. 11.48 and 11.52).

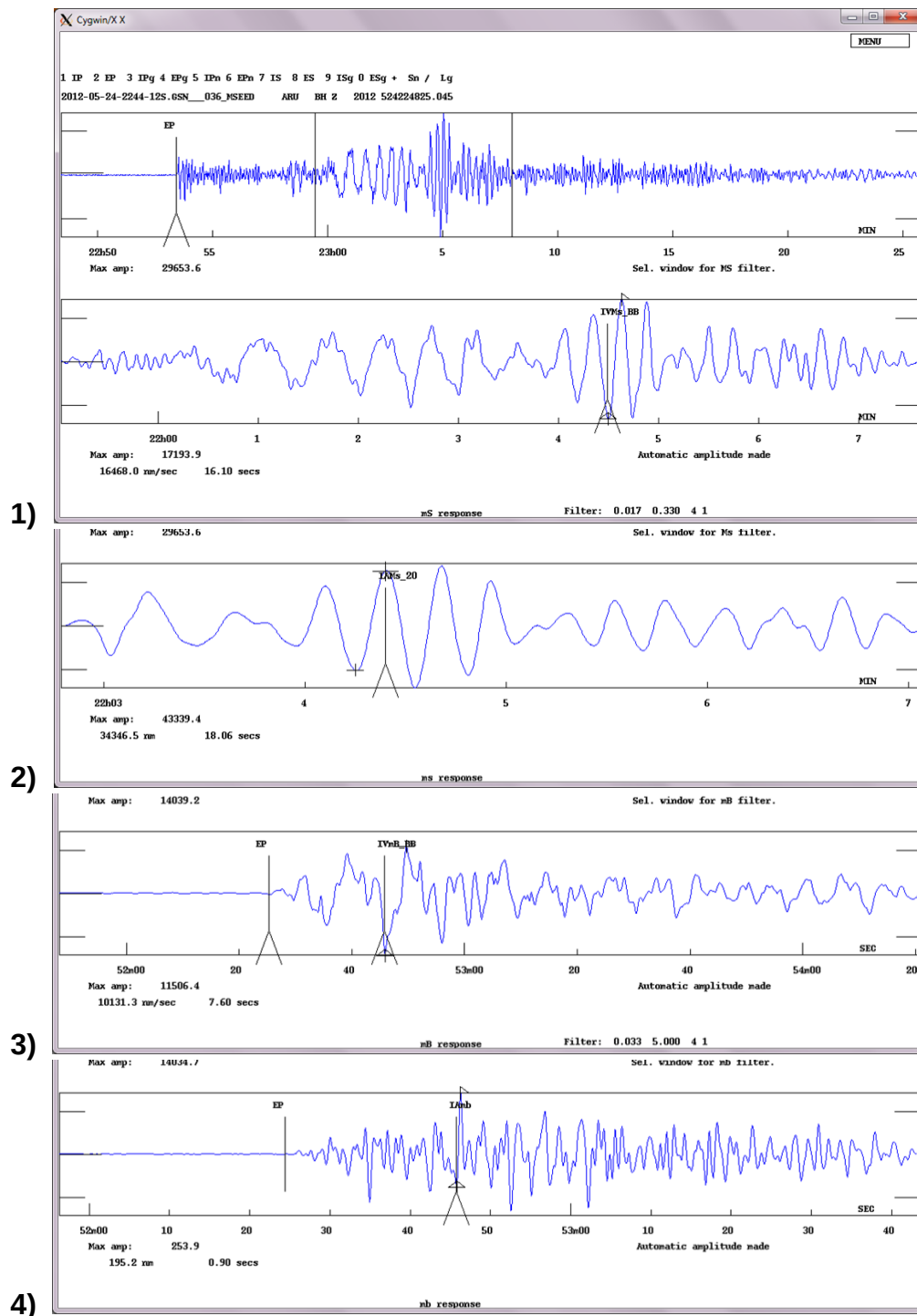
Furthermore, the following tasks are implemented:

- rotation of horizontal components (Figs. 11.17, 11.18 and 11.48);
- particle motion diagrams (Fig. 2.7);
- trace amplitude spectrum (Fig. 11.60);
- vespagram trace display (Fig. 11.46);
- determination of signal/noise ratio (Fig. 11.60); and
- trace editing functions.

In 2012 an optional plugin for the magnitude determinations following the IASPEI standards has been implemented, as an alternative to the slightly differing traditional magnitude routines and filter parameters of the Gräfenberg observatory (mb, MS, ml). All resulting parameters are stored in an easily readable and extendable text file format. Different waveform data formats are supported, however the preferred data format is continuous MiniSEED. SHM is supported on a number of Linux distributions (see web page).

### 11.4.3 SEISAN

Another widely used seismic analysis system is SEISAN developed at University of Bergen, Norway (Ottemöller, Voss, and Havskov, J. (Eds.) (2013). It contains a complete set of programs and a simple database for analyzing digital recordings. SEISAN can be used, amongst other things, for phase picking, spectral analysis, azimuth determination, and plotting seismograms. SEISAN is supported by DOS, Windows95, SunOS, Solaris and Linux and contains conversion programs for the most common data formats. The program, together with a detailed Manual, is available via <http://www.ifjf.uib.no/seismo/software/seisan/seisan.html> and can also be downloaded from the Manual front page via the link *Download programs and files*. IS 11.6 gives examples of record and parameter presentation, the picking of phase onset times, amplitudes and periods, event location and magnitude determination in the local, regional and teleseismic distance range and the determination of source mechanisms by using either first motion polarities or S/P amplitude ratios. Since its version 8.3 SEISAN is so far the only widely used interactive analysis software (besides the NEIC Hydra software) that has implemented as default the IASPEI recommended filters for determining the standard magnitudes. On the records, where the respective amplitudes and periods are measured, the cursors are annotated with the related unique standard amplitude phase names (Table 4 in IS 3.3 and Fig. 11.50). However, as of now (October 2013), neither SEISAN nor any other currently operating analysis software, with the exception of the NEIC Hydra system, presents the standard magnitudes, with the exception of mb and ML, also in their standard nomenclature (e.g., mB\_BB, Ms\_20, Ms\_BB). Therefore, with its resolution No. 1 of 2013 (Fig. 11.36) IASPEI encourages developers of waveform processing programs to incorporate these standards into their software and the station and network operators as well as data centers to adopt these magnitude and amplitude standards in their day to day operations.



**Fig. 11.50** Compiled from SEISAN analysis example 3 in IS 11.6. **1) Uppermost trace:** Velocity broadband record at station ARU ( $D = 26.2^\circ$ ) of the 24 May 2012 Mw6.2 Mohns Ridge earthquake, offshore northern Norway. **Lower trace:** broadband velocity surface wave window with the marked position for  $M_s$ \_BB amplitude measurement, annotated with the standard amplitude phase name  $IVM_s\_BB$ . **2) to 4)** Filtered traces for  $M_s$ \_20,  $mB\_BB$  and  $mb$  amplitude measurements according to IASPEI standards with marked positions annotated with the respective standard amplitude phase names  $IAM_s\_20$ ,  $IVmB\_BB$  and  $IAmb$ . I stands for IASPEI standard, A for displacement amplitude and V for velocity amplitude.

Since the beginning of the 1990s, SEISAN was installed in many developing countries in Central and South America, the Caribbean, Asia and Africa by the seismological institutions University of Bergen and NORSAR within Norwegian technical assistance projects on seismic hazard and risk mitigation. The package served very well the needs of the seismological observatories of these countries, was improved and developed taking into account the variable environments and user inputs. The package unites routine seismic processing, reporting, and programs for scientific analysis, for crustal structure, seismic hazard estimation and other topics. The program had always a companion program named SEISLOG for data acquisition and event detection. A new module RTQUAKE is in development which includes also automatic event location. A combination of RTQUAKE and SEISAN could be a good alternative for seismologists at small networks who fear the difficulties to install Earthworm or SeisComP which have a rather steep learning curve.

#### 11.4.4 SeisComP3

SeisComP (Seismological Communication Processor; Hanka et al., 2000 and 2010; <http://www.seiscomp3.org>) is the most commonly used software in Europe within the scope of real-time earthquake monitoring and increasingly also on a global scale with meanwhile about 300 installations/licenses granted world-wide. It had originally been developed in conjunction with the GEOFON network (Hanka and Kind, 1994; see also Chapter 8, Fig. 8.5) and the GEOFON earthquake information system operated at GFZ (German Research Centre for Geosciences in Potsdam) (see <http://geofon.gfz-potsdam.de/>). All stations in the global GEOFON network send their data in real time to the GFZ.

A major breakthrough has been the development of the SeedLink protocol for data acquisition and transfer (Hanka et al., 2000) as part of the GEOFON's SeisComP concept. It permits the connection of digitizers and data loggers from all major manufacturers and all major acquisition systems, the conversion of the native data formats to MiniSEED (for formats Chapter 10) and the uniform transmission, acquisition and archival of heterogeneous network data (Hanka et al., 2010). Automatic real-time data processing became part of SeisComP in 2003. The early GFZ prototype earthquake monitoring system was upgraded with SeisComP3 (for short SC3) which forms a powerful system able to issue within a few minutes automatic but nevertheless rather reliable global and regional earthquake parameter information. Moreover, SC3 allows later interactive refinement and distribution of revised solutions. This system provides now the backbone for the German Indonesian Tsunami Early Warning System (GITEWS; <http://www.gitews.org/index.php?id=6>), for other tsunami warning centers in the Indian Ocean and the NEAMTWS (Northeastern Atlantic and Mediterranean Tsunami Warning System) area as well as for the GEOFON Extended Virtual Network (GEVN) which presently comprises more than 900 stations world-wide (Hanka et al., 2010).

Besides reliable near real-time automatic data processing SC3 allows very rapid visualization of the results, quick graphical review, and interaction of seismic experts in the warning centers at any time in order to improve the automatic results. This was a precondition under the GITEWS time constraints to issue tsunami warnings already within 5 min after origin time. According to Hanka et al. (2010) the basic automatic system of SC3 consists of modules for:

- quality control
- picking
- event location

- amplitude and magnitude calculation
- waveform quality assessment and
- event and station parameter management.

The interactive part of the system provides according to Hanka et al. (2010):

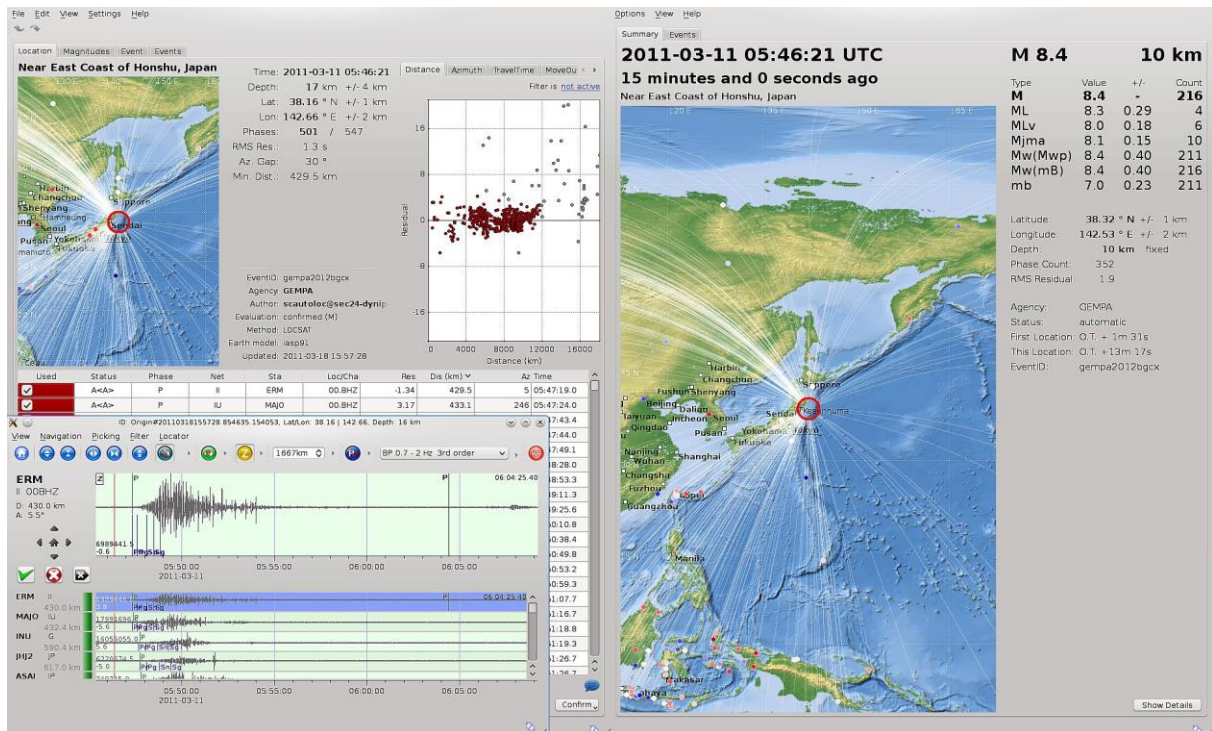
- graphical user interfaces for visualizing in a map the overall situation regarding earthquake location and station status (Fig. 11.51)
- real-time trace and event summary display
- toolkits for analyzing the earthquake epicenter, source depth and magnitude
- manual picker optimized for rapid verification of pre-calculated picks from strong earthquakes
- automatic loading of newly acquired real-time data with picks associated to the ongoing event
- automatic amplitude scaling and trace alignment
- conventional off-line event analysis, also of small and moderate earthquakes that have not been processed by the automatic system.

The latter point allows for more detailed and accurate seismic phase and magnitude measurements for the final station and/or network parameter bulletins and thus also for making valuable contributions to final regional and global seismological parameter bulletins, e.g., of the EMSC and the ISC, in agreement with IASPEI recommended measurement standards. Making full use of the off-line data processing and analysis potential offered by SC3 reduces procedure dependent data scatter and systematic biases. This assures best representativeness, long-term compatibility and accuracy of its final data products also on a global scale, thus improving the usefulness of international parameter data bases for earthquake statistics, hazard assessment and other research tasks.

Although SC3 measures in its automatic mode, besides several other magnitudes, in principle already two of the four IASPEI recommended teleseismic standard magnitudes, namely mb and mB\_BB, the stringent time limit set for early warning applications requires to measure mb already within the first 30 s and mB within the first 60 s of the event record traces. For the greatest or very slow strong earthquakes this means, however, that the largest amplitude in the P-wave train may still be missed by such a pre-set measurement time window, more frequently for mb but occasionally also for mB (see, e.g., Figs. 3.6, 3.13, 3.46, 3.62 and 3.65 in Chapter 3). Then these two body-wave magnitudes but also the Mw proxy estimate Mw(mB) (see Bormann and Saul, 2008) may be underestimated, up to about 0.3 magnitude units (m.u.) This can be corrected easily by later off-line analysis, when the whole wave train is available. Then also the very important complementary broadband surface-wave magnitude standard Ms\_BB (Chapter 3, section 3.2.5), can be measured without any restrictions set by the rapid automatic processing mode when applied to large global sets of waveform data for early warning purposes. For smaller networks, however, it is said to be practicable already even in the automatic mode.

Originally developed for tsunami early warning, SC3 evolved also to be a very powerful system for local earthquake monitoring including off-line analysis, providing typically needed functionality as customizing and applying filters to the records, project theoretical phase onset times on the records, allow axes rotations to ease phase detection and identification and provide a first motion analysis to determine source mechanisms.





**Fig. 11.51** SeisComP3 screenshots of an authentic playback from the Tohoku earthquake on March 11, 2011. **Left** side: Screenshot from the SC3 module to review/revise hypocenter solutions; **right side** shows the event summary view (Figure by courtesy of B. Weber).

## 11.4.5 Earthworm

Earthworm is a non-commercial seismological software package (Johnson et al., 1995; <http://folkworm.ceri.memphis.edu/ew-doc/> and <http://www.earthwormcentral.org/>). On its web page it is described as follows:

“Earthworm is an open architecture, open source public software for data acquisition, processing, archival and distribution. It was originally developed by the United States Geological Survey. Earthworm binaries and source files are freely available to everyone. Development and maintenance are now carried out by the Earthworm User Community, [ISTI](http://www.isti.com/) (Instrumental Software Technologies, Inc., New York, USA) and [CERI](http://www.ceri.edu/) (Center for Earthquake Research and Information, University of Memphis, USA).”

The software is written in C and can be compiled for Windows, Linux, Solaris and Mac OS X operating systems. Precompiled binaries can be downloaded from the Earthworm websites. Earthworm seismic data acquisition and processing software can be used for single stations as well as local, regional and global seismic networks. The development of the package started in 1993. Nowadays, over 150 institutions have registered installations of Earthworm. Many more “Earthworms” run on the office computers or Laptops of seismologists, geophysicists, and hobbyists without formal registration as the software is completely free. In addition to seismology, Earthworm's data acquisition and management capabilities are used for other waveform data, including Infra-sound, Geomagnetic, and Atmospheric measurements.

Earthworm is very structured and consists of around 130 modules which the user can configure according to his needs and combine them for executing a specified task. This task can be so simple as to acquire the data of a single component and write its data to disk. The script which controls the execution would contain only a few lines and an experienced user would edit it in some minutes. The task can also be rather complicated as to acquire, in real time, the data of a network of several hundred stations, perform band pass filtering and phase picking for all the traces, event definition, automatic localization and magnitude determination. To control such a demanding task necessitates, however, a rather large script which might require many weeks or months of development and testing to get it working properly. Simpler Earthworm configurations can run on very cheap and small computers such as the Raspberry Pi. For a small or medium seismic network a low cost PC with two cores could be sufficient. Larger networks of several hundred stations require more powerful computers. Different tasks could even been distributed to several servers in the seismic data center.

There are around 30 data acquisition modules suited for many of the A-D-cards and digital seismographs on the market, even for some cheap A-D-cards used mainly by hobbyists. Earthworm systems can communicate with each other as with other data acquisition and processing systems (e.g. SeisComP) by using special modules. For instance, an Earthworm installed on a simple computer with an A-D card could acquire data at a seismic station and transmit them in real time to another Earthworm or SeisComP which is running in a server of the seismic data center and performs the automatic earthquake detection and processing. Earthworm has also modules for sending out email messages triggered by seismic event detection but Earthworm lacks sophisticated visualization tools for seismogram.

“Glowworm” is a special suite of Earthworm modules that has been developed at the USGS in the late 1990s. It contains data archiving and processing methods and visualization tools typical for volcano monitoring. Nowadays, Glowworm seems to be outdated as several of its modules were not adapted to structural changes of the main stream of Earthworm. Nevertheless, it is still in use in many volcano observatories. The most important processing modules of original Glowworm, e.g. for the continuous calculation of RSAM (*Real-time Seismic-Amplitude Measurement*, which represents the overall signal size over periods of 10 minutes) and SSAM (*Seismic Spectral-Amplitude Measurement*, which shows the relative signal size in different frequency bands) were ported to the standard Earthworm.

Standard Earthworm is dedicated only to automatic data acquisition and routine processing. Therefore it has no own interactive processing modules. But it contains modules which permit the generation of event files in data formats needed by other seismological data processing programs, for instance SEISAN or SAC. These packages can then be used for interactive routine processing or for scientific research. A widely used combination at seismological observatories with smaller networks and lack of personnel experienced in computing and programming is to use Earthworm for data acquisition, event detection and generation of event files; and SEISAN for the interactive routine processing of the events. This avoids the automatic location and magnitude determination which is not easy to configure. Based on Earthworm the Alaska Tsunami Warning Center (ATWC) developed the *Early Bird* package (Luckett et al., 2008) for the detection, automatic location and magnitude determination of local, regional and teleseismic events, especially for tsunami warning. It contains also modules for seismogram visualization and interactive processing, phase picking and generation of epicenter maps.



## 11.4.6 ObsPy

*ObsPy* is an open-source project (Beyreuther et al. 2010; Megies et al., 2011) aimed at facilitating rapid application developments for seismology. It provides a toolbox using *Python* language (<http://www.python.org/> and <http://en.wikipedia.org/wiki/Python>) and simplifies the usage of Python programming for seismologists. It provides parsers for common file formats, clients to access data centers and seismological signal processing routines that allow the manipulation of seismological time series. Good starters are [ObsPy Gallery](#) and Tutorial (<http://docs.obspy.org/index.html>). *ObsPy* is currently [running and tested](#) on Linux, Windows XP/Vista/7/8 and Mac OS X.

In *ObsPy* the following essential seismological processing routines are implemented and ready to use:

- reading and writing data only SEED/MiniSEED and Dataless SEED ([http://www.iris.edu/manuals/SEEDManual\\_V2.4.pdf](http://www.iris.edu/manuals/SEEDManual_V2.4.pdf));
- XML-SEED (Tsuboi et al. 2004);
- GSE2 ([http://www.seismo.ethz.ch/autodrm/downloads/provisional\\_GSE2.1.pdf](http://www.seismo.ethz.ch/autodrm/downloads/provisional_GSE2.1.pdf)) and
- SAC (<http://www.iris.edu/manuals/sac/manual.html>), as well as filtering, instrument simulation, triggering, and plotting.
- Modules to read SEISAN data files (Havskov and Ottemöller 1999) and to retrieve data with the IRIS/FISSURES data handling interface (DHI) protocol (Malone 1997);
- Support to retrieve data from ArcLink (a distributed data request protocol for accessing archived waveform data, see Hanka and Kind 1994) or a SeisHub database (Barsch 2009).

While Python gives the user all the features of a full-fledged programming language including a large collection of scientific open-source modules, *ObsPy* extends Python by providing direct access to the actual time series, thus allowing the use of powerful numerical array-programming modules like NumPy (<http://numpy.scipy.org>) or SciPy (<http://scipy.org>) and the visualization of the results. This is an advantage over SAC, SEISAN, and SeismicHandler which do not provide methods for general numerical array manipulation.

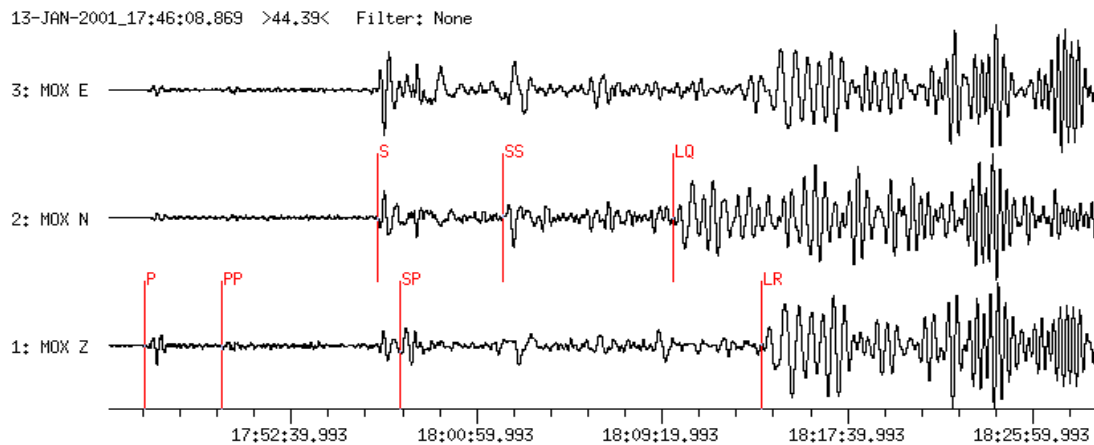
## 11.5 Examples of seismogram analysis

### 11.5.1 Introductory remarks

The character of a seismogram depends on the source mechanism and size, the source depth, the signal-to-noise ratio (SNR) and whether the epicenter of the source is at local, regional or teleseismic distances. Many seismic phases of different origin and character may occur in these different ranges. For explanations of their phase names and origin see IS 2.1. In the following we deliberately show mostly records with good SNR because we wish to illustrate what types of seismic phases, also relatively weak ones with respect to the strongest phases, may occur at which distances and what their basic features such as relative travel-time, period, amplitude and polarity are with respect to other phases. However, most seismic events are relatively weak and therefore the SNR is low and most signal phases are buried in the noise, especially in broadband records. Then filter procedures have to be applied in order to improve the SNR. But what types of filters are suitable depends on the frequency, wave number and polarization properties of the signal and of the noise. These problems, including

those of bandwidth-dependent waveform distortion due to filtering, are outlined in detail in Chapter 4. There also many examples are presented on SNR-improvement by applying different types of filters and their pros and cons are discussed. For array-specific methods of SNR improvement see also Chapter 9 and in the Data Sheets DS 11.1-11.3 more examples are given that show the different record appearance depending on the applied filter and time resolution. There are no straightforward standardized receipts for optimal filtering. Which filter procedure and parameters are best suited also depends on which parameters (such as onset time, amplitude, period, slowness or polarization) have to be determined best in which frequency range and of what type of seismic phase. Only rich own interactive and task-dependent experience can help to develop the proper “feeling” for it. Who only uses and relies on some basic default parameter settings of automated procedures will never get a deeper understanding of it and learn to discriminate between real signal improvements, distortions or even filter-induced artifacts.

Seismograms of small local earthquakes are characterized by short duration of the record from a few seconds to 1-2 minutes, higher frequencies, and a characteristic shape of the wave envelope, usually an exponential decay of amplitudes after the amplitude maximum, termed “coda” (Figure 1b in DS 11.1 and Figure 2 in EX 11.1). However, records of large earthquakes with longer rupture duration and thus radiating dominantly longer periods may look differently in the local distance range. At teleseismic distances records show generally lower frequencies (because high-frequency energy has already been reduced by anelastic attenuation and scattering), and much longer duration from say fifteen minutes to several hours (e.g., Figs. 1.6, 2.20, 2.21 and 11.52).



**Fig. 11.52** Example of 3-component broadband velocity seismograms of a teleseismic  $M_w = 7.7$  earthquake in El Salvador recorded at station MOX, Germany ( $D = 86.5^\circ$ ).

Records of regional events at distances of several hundred kilometers have intermediate features, and are - as local events - characterized by still high frequency content and only a few distinct phase onsets with much signal-generated noise in between due to scattering at crustal heterogeneities. This hardly allows to recognize and pick for sure the onsets of weaker but theoretically expected phases such as  $P_b$ ,  $S_b$ , and  $S_n$  (Figs. 11.53 and 11.54 in section 11.5.1).

The various wave groups above the noise level, arriving at a station over different paths, are called seismic phases. They have to be recognized, identified and their parameters determined (onset time, amplitude, period, polarization, etc.). Phase symbols should be assigned according to the IASPEI recommended standard nomenclature of seismic phases. For phase names, their definition and ray paths see IS 2.1. Types and characteristics of seismic phases vary with distance, depending also on the source type and depth. These characteristics will be discussed in more detail in the following sections.

There is no unique standard definition yet for the distance ranges termed “near” or “local” and “regional” (near to far regional), or “distant” (“teleseismic”; sometimes subdivided into “distant” and “very distant”). Regional variations of crustal and upper-mantle structure make it impossible to define a single distance at which propagation of local or regional phases stops and only teleseismic phases will be observed. In the following we consider a source as local if the direct crustal phases Pg and Sg arrive as first P- and S-wave onsets, respectively. In contrast, the phases Pn and Sn, which have their “bottoming” (or starting) point in the uppermost mantle, are the first arriving P and S waves in the regional distance range. However, as discussed in Chapter 2, section 2.6.1 and shown in Fig. 2.47, the distance at which Pn takes over as first arrival depends on the crustal thickness, average wave speed and the dip of the crustal base. The “cross-over” distance  $x_{co}$  between Pn and Pg is - according to Eq. (6) in IS 11.1 for shallow (near surface) sources - roughly  $x_{co} \approx 5 \times z_m$  where  $z_m$  is the Moho depth. Note, however, that as focal depth increases within the crust,  $x_{co}$  decreases, down to about  $3 \times z_m$ . Accordingly, the local distance range may vary from region to regions and range between about 100 km and 250 km. The CTBTO Technical Instructions (see IDC Documentation, 1998) considers epicentral distances between  $0^\circ$  to  $2^\circ$ , where Pg appears as the primary phase, as local distance range.

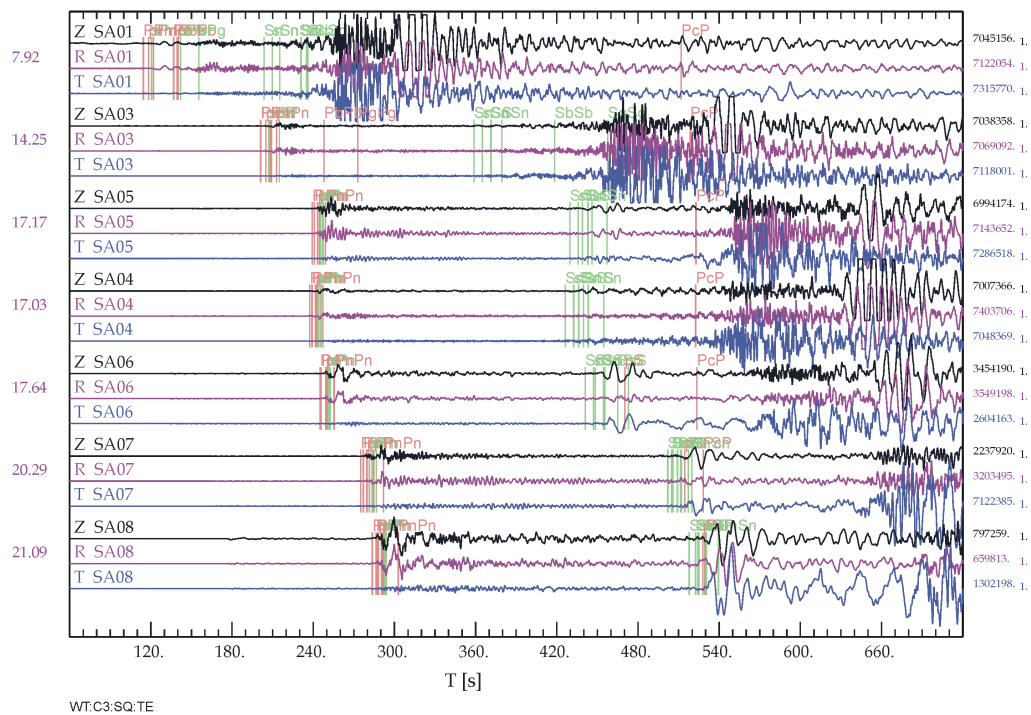
The Willmore (1979) Manual defines as near earthquakes those which are observed up to about 1000 km (or  $10^\circ$ ) of the epicenter, and P and S phases observed beyond  $10^\circ$  as usually being teleseismic phases. However, regional phases such as Pn, Sn and Lg, will generally propagate further in stable continental regions than in tectonic or oceanic regions. According to the Earth model IASP91, Pn may be the first arrival up to  $18^\circ$ . The rules published in the IDC Documentation (1998) allow a transitional region between  $17^\circ$  and  $20^\circ$  in which phases may be identified as either regional or teleseismic, depending on the frequency content and other waveform characteristics. Accordingly, one might roughly define seismic sources as local, regional and teleseismic if their epicenters are less than  $2^\circ$ , between  $2^\circ$  and  $20^\circ$ , or more than  $20^\circ$  away from the station. Sometimes, the regional range is further subdivided into  $2^\circ$ - $6^\circ$ , where also the phase Rg may be well developed (Fig. 11.23), and  $6^\circ$ - $20^\circ$  where only Sn and Lg are strong secondary phases.

On the other hand, the calibration curve for the local magnitude  $M_L$  according to Richter (1935) has originally been given for distances up to 600 km and has been extended now for some regions far beyond (up to 1500 km, Table 2 in DS 3.1) making  $M_L$  an even far regional magnitude. On the other hand, the new IASPEI standards for teleseismic magnitudes  $m_b$  and  $M_s(20)$  are measured between  $20^\circ$  and  $100^\circ$ , respectively  $20^\circ$  and  $160^\circ$ , whereas the broadband surface-wave magnitude  $M_s(BB)$ , measured between  $2^\circ$  and  $160^\circ$ , is in fact both a regional and a teleseismic magnitude.

At distances beyond  $15^\circ$  Pn and Sn amplitudes become too small (except in some shield regions). The first arriving P and S phases have then already taken travel paths through deeper parts of the upper mantle, thus being affected by the pronounced velocity increases at the

bottom of the upper mantle (410 km discontinuity) and at the bottom of the transition zone to the lower mantle (660 km discontinuity) (Chapter 2, Figs. 2.77 and 2.79). This gives rise to the development of two triplications of the P- and S-wave travel-time curves with prograde and retrograde branches, the second one reaching up to about  $28^\circ$  epicentral distance. In some distance ranges the related P and S onsets follow closely to each other (Figs. 2.32 and 2.76), thus forming rather long and complex waveforms (Fig. 11.53). The largest amplitudes occur in the range of the left-side cusp of the 660 km discontinuity triplication (P660P) between about  $18^\circ$  and  $21^\circ$  (termed “ $20^\circ$ -discontinuity”; see also the amplitude-distance curves in Fig. 11.10) but with weaker P-wave first arrivals some 5 to 10 s earlier. Accordingly, small differences in epicentral distance can lead to large differences in the amplitude of the body-wave groups in seismic records (see also Figure 2b in DS 11.2).

SK1: 1993.232.05.r7

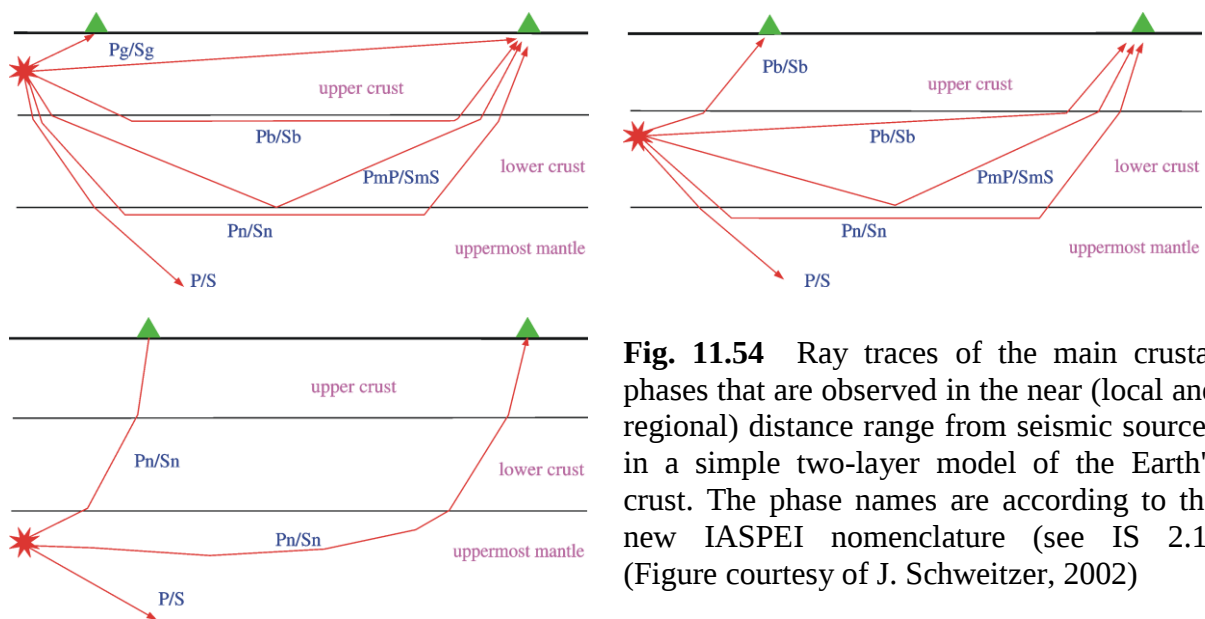


**Fig. 11.53** 3-component records in the distance range between  $7.9^\circ$  and  $21.1^\circ$  by a regional network of portable broadband instruments deployed in Queensland, Australia (seismometers CMG3ESP; unfiltered velocity response). The event occurred in Papua New Guinea at 15 km depth. The predicted phase arrival times for the AK135 model are depicted. Primary, depth and other secondary arrivals (such as PnPn in the P-wave group and SbSb as well as SgSg in the S-wave group) superpose to complex wavelets. Also note that several of the theoretically expected phases have such weak energy that they cannot be recognized on the records at the marked predicted arrival times above the noise level or the signal level of other phases (e.g., PcP) (courtesy of B. Kennett).

These complications and different pros and cons for various definitions notwithstanding we present, slightly different from NMSOP-1 and DS 11.2, in the following the examples from near to regional sources up to about  $20^\circ$ , in the teleseismic range between about  $20^\circ$  and  $100^\circ$ , and in the far teleseismic “core shadow” range beyond  $100^\circ$ ,

## 11.5.2 Seismograms from near and regional sources ( $0^\circ < D < \approx 20^\circ$ )

Seismograms recorded in this distance range are dominated by P and S waves that have traveled along different paths through the crust and the upper mantle of the Earth. They are identified by special symbols for “crustal phases” (see IS 2.1). Pg and Sg, for example, travel directly from a source in the upper or middle crust to the station whereas the phases PmP and SmS have been reflected from, and the phases Pn and Sn critically refracted along or “dived beneath” the Moho discontinuity (Fig. 11.53). Empirical travel-time curves are given in Exercise EX 11.1 (Figure 4) and a synthetic record section of these phases in Fig. 2.82. In some continental regions, phases are observed which have been critically refracted from a mid-crustal discontinuity or have their turning point in the lower crust. They are termed Pb (or P\*) for P waves and Sb (or S\*), for S waves, respectively. For shallow sources, crustal “channel-waves” Lg (for definition see IS 2.1) and surface waves Rg are observed after Sg-waves. Rg is a short-period Rayleigh wave ( $T \approx 2$  s) which travels in the upper crust and is usually well developed in records out to about 300 km of near-surface sources such as quarry blasts or mining induced events and thus suitable for discriminating such events from local tectonic earthquakes (Figs. 11.23a-d).



**Fig. 11.54** Ray traces of the main crustal phases that are observed in the near (local and regional) distance range from seismic sources in a simple two-layer model of the Earth's crust. The phase names are according to the new IASPEI nomenclature (see IS 2.1) (Figure courtesy of J. Schweitzer, 2002)

Usually, Sg and SmS (the supercritical reflection, which often follows Sg closely at distances beyond the critical point; Fig. 2.43) are the strongest body wave onsets in records of near seismic events whereas Pg and PmP (beyond the critical point) have the largest amplitudes in the early part of the seismograms, at least up to 200 – 400 km. Note that for sub-crustal earthquakes no reflected or critically refracted crustal phases exist. However, according to the new IASPEI nomenclature, P and S waves from sub-crustal earthquakes with rays traveling from there either directly or via a turning point in the uppermost mantle back to the surface are still termed Pn and Sn (Fig. 11.54, lower left). At larger distances such rays arrive at the surface with apparent “sub-Moho” P and S velocities (see below).

Typical propagation velocities of Pg and Sg in continental areas are 5.5-6.2 km/s and 3.2-3.7 km/s, respectively. Note, that Pg and Sg are direct waves only to about 2° to 3°. At larger distances the Pg-wave group may be formed by superposition of multiple P-wave reverberations inside the whole crust (with an average group velocity around 5.8 km/s) and the Sg-wave group by superposition of S-wave reverberations and SV to P and/or P to SV conversions inside the whole crust. According to the new IASPEI phase nomenclature the definitions given for Lg waves and Sg at larger distances are identical, with the addition that the maximum energy of an Lg crustal “channel” wave travels with a group velocity around 3.5 km/s. According to the definition of the IASPEI mb\_Lg standard magnitude (see IS 3.3) the amplitude of Lg is measured as the “...*third largest amplitude in the time window corresponding to group velocities of 3.6 to 3.2 km/s, in the period (T) range 0.7 s to 1.3 s*”.

In routine analysis, usually only the first onsets of these wave groups are picked without noting the change in character at larger distances. According to the Technical Instructions of the IDC Documentation (1998), stations of the CTBTO International Monitoring System (IMS) generally tend to name the strongest transverse arrival Lg and not Sg. A reliable discrimination is still a subject of research and not yet one of routine analysis and data reporting. Therefore, no simple and unique criteria for discrimination, which also depend on source type and propagation path, can be given here. They may be added to this Manual at a later time.

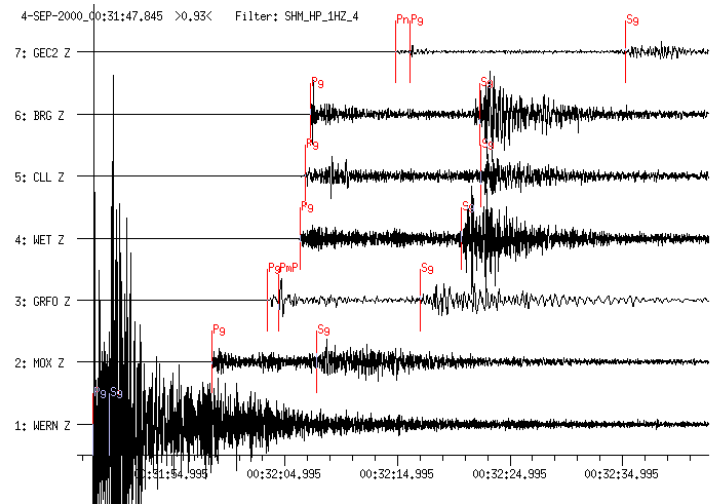
Lg waves may travel in continental shield regions over large distances (Fig. 2.15), even beyond 20° whereas Rg waves, which show clear dispersion and longer periods than Lg (Fig. 11.23a and b), are more strongly attenuated and generally not observed beyond 6°. The apparent velocities of Pn and Sn are controlled by the P- and S-wave velocities in the upper mantle immediately below the Moho and typically range between 7.5 - 8.3 km/s and 4.4 - 4.9 km/s, respectively.

**Note:** Seismograms from local and regional seismic sources are strongly influenced by the local crustal structure which differs from region to region and even between local stations. This may give rise to the appearance of other onsets (which may be strong) between the mentioned main crustal phases. These “unexpected” phases cannot simply be explained as direct or refracted/reflected waves radiated from a near-surface source in a single or two-layer crustal model. Some of these phases may relate to converted waves and/or depth phases such as sPmP (e.g., Bock et al., 1996). Also, at larger distances, up to about 30°, multiples such as PgPg, PbPb, PnPn, PmPPmP etc. and their related S phases, are expected to follow each other closely within the P and S-wave group according to the AK135 travel-time model (Fig. 11.53).

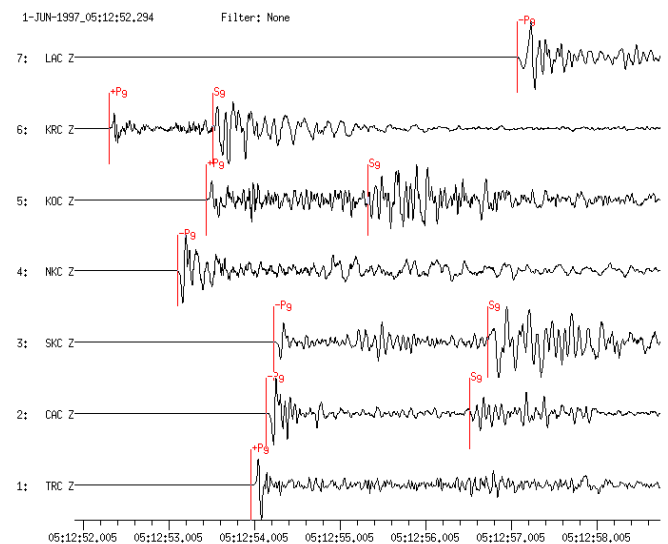
However, usually these details cannot be handled in routine data analysis and epicenter location and require specialized study. For routine purposes, as a first approximation, the IASP91 or AK135 global models (see DS 2.1) can be used for the analysis and location of near events based on the main crustal phases. However, one should be aware that crustal structure and velocities may differ significantly from region to region. Therefore, the event location can be significantly improved when local travel-time curves or crustal models are available (e.g., IS 11.1, Figures 11 and 12).

Fig. 11.55 shows seismograms of a shallow local earthquake from the Vogtland/NW Bohemia region in Central Europe, recorded at seven GRSN stations in the epicentral distance

range 10 km (WERN) to 180 km (GEC2). Stations up to  $D = 110$  km (BRG) show only the direct crustal phases Pg, Sg, except GRFO, which in addition shows PmP. At GEC2 Pn arrives ahead of Pg with significantly smaller amplitude. The onset times of phases Pg, Sg and Pn were used to locate the epicenter of this event with a precision of about 2 km. If more stations close to the epicenter are included (e.g., Fig. 11.56;  $D = 6 - 30$  km), the precision of the hypocenter location may be in the order of a few hundred meters.



**Fig. 11.55** Filtered short-period vertical component seismograms (4<sup>th</sup> order Butterworth high-pass filter,  $f = 1$  Hz) from a local earthquake in the Vogtland region, 04 Sept. 2000 ( $50.27^\circ\text{N}$ ,  $12.42^\circ\text{E}$ ;  $h = 8$  km,  $M_l = 3.3$ ). Sampling rates at the stations differ: 80 Hz for MOX, WET, CLL, and BRG, 100 Hz for WERN and 20 Hz for GRFO and GEC2. Traces are sorted according to epicentral distance (from 10 to 180 km). The local phases have been marked (Pg and Sg at all stations, PmP at GRFO and Pn at the most distant station GEC2 ( $D = 180$  km)).



**Fig. 11.56** Short-period recordings from stations of the local network of the Czech Academy of Sciences in Prague from a small ( $M_l = 3.3$ ) local earthquake in the German/Czech border region on June 1, 1997. The epicentral distance range is 6 km and 30 km. Such local networks allow hypocenters to be located to better than a few hundred meters.

From these figures and the animation file 1 in IS 11.3 the following conclusions can be drawn:

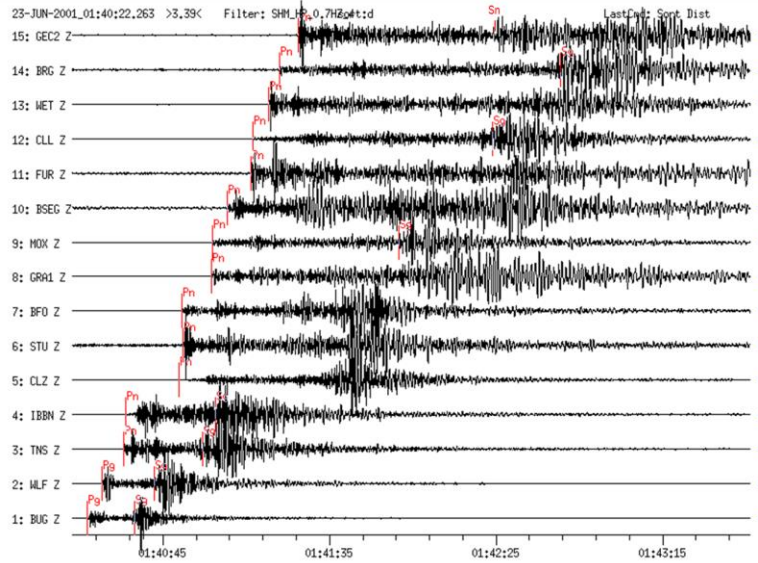
- at some stations the arrival times are in good agreement with the times predicted from an average crustal model for the considered region, at other stations they are not. This implies lateral and/or distance-dependent variations in crustal velocities;
- the amplitude ratio  $P_g/S_g$  varies strongly with azimuth because of the different radiation patterns for P and S waves. This can be used to derive the fault-plane solution of the earthquake (Figs. 3.100 - 3.103 in section 3.4.1 of Chapter 3).

Fig. 11.57 shows recordings from an earthquake in the Netherlands ( $M_l = 4.0$ ) in the distance range 112 km to 600 km, and Fig. 11.58 those from a mining-induced earthquake in France ( $M_l = 3.7$ ) in the range 80 km to 500 km. These records again show obvious variations in the relative amplitudes of  $P_n$ ,  $P_g$  and  $S_g$ . The relative amplitudes depend on both distance and azimuth of the stations relative to the radiation pattern of the source, and particularly with respect to the differences in take-off angles of the rays for the direct and the critically refracted waves (Fig. 11.54). The source depth with respect to the major crustal discontinuities may also influence the relative amplitude ratio between these various phases.

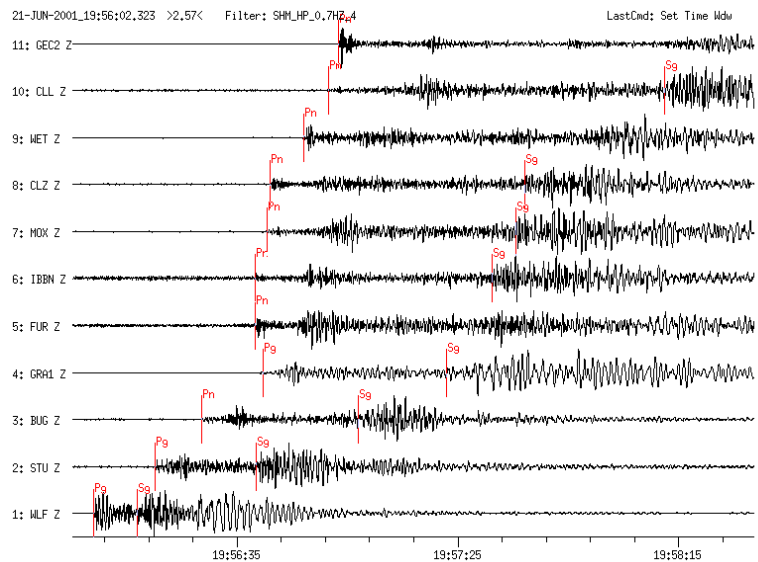
As in Figs. 11.23a and b, also the record of station WLF in Fig. 11.58 of this shallow mining induced event shows well developed longer-period  $R_g$  waves that follow  $S_g$ .  $R_g$  is usually not present in records of tectonic earthquakes (Figs. 11.55-11.57).

For near-surface sources and distances smaller than about 400 km,  $P_n$  is usually much smaller than  $P_g$  (see also Fig. 11.34 and Figures 3a and 3c in Datasheet 11.1). But for larger distances the relative amplitudes of  $P_n$  and  $S_n$  may grow so that these phases may even become the dominating first arriving the P and S waves (Fig. 2.17 and the uppermost traces in Figs. 11.57 and 11.58). This is not only because of the stronger attenuation of the direct waves that travel mostly through the uppermost heterogeneous crust, but also because P and S near the critical angle of refraction at the Moho form so-called “diving” phases which are not critically refracted into the Moho but rather travel within the uppermost mantle with sub-Moho velocity. Such diving phases have much larger amplitudes than the ones to be expected by theory for critically refracted inhomogeneous waves. The sub“diving”  $P_n$  and  $S_n$  at distance beyond 600-800 km becoming the dominating first arriving P and S onsets, usually have longer periods than  $P_n$  and  $S_n$  at shorter distances.



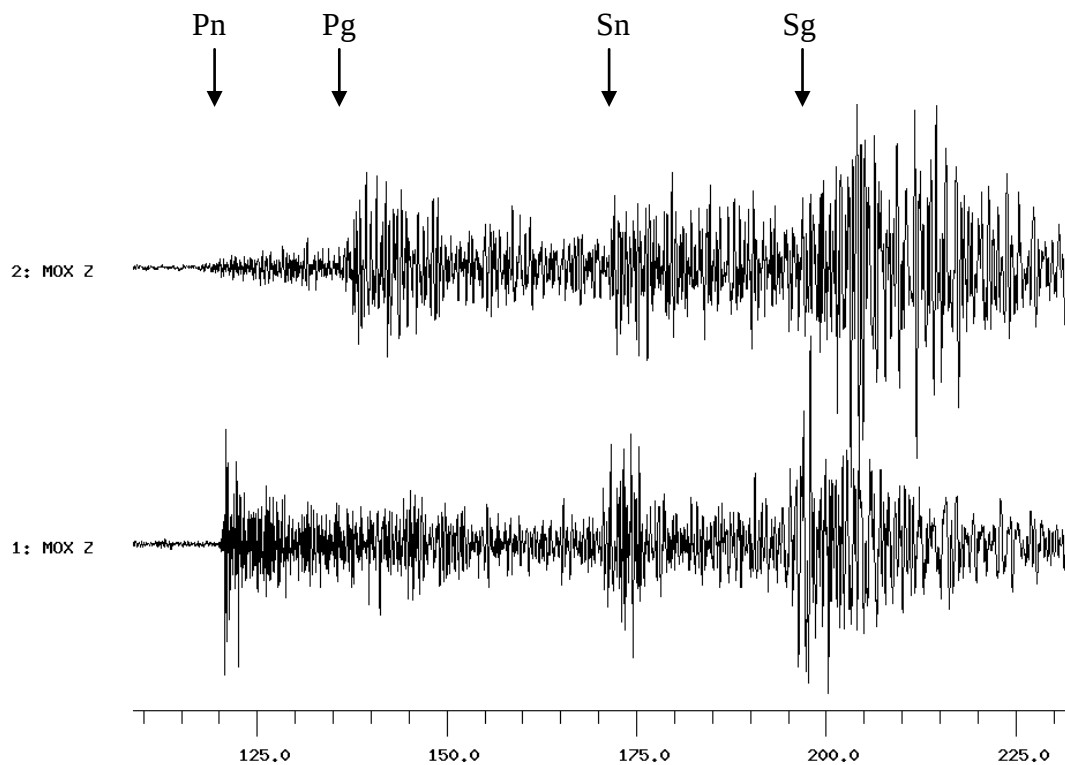


**Fig. 11.57** Vertical-component short-period filtered broadband seismograms (4<sup>th</sup> order Butterworth high-pass filter,  $f = 0.7$  Hz; normalized amplitudes) from a local earthquake at Kerkrade, Netherlands, recorded at 15 GRSN, GRF, GERES and GEOFON stations.  $M_l = 4.0$ ; epicentral distances between 112 km (BUG) and 600 km (GEC2). The time difference  $S_n$ - $P_n$  of 60 s in the GEC2 records yields also with the “rule of thumb” in Eq. (1.2) the calculated epicentral distance for this station. Note the variability of waveforms and relative phase amplitudes of local/regional earthquakes in network recordings in different azimuths and epicentral distances. The suitability of filters for determination of local phase onsets has to be tested. Local magnitudes are measured on a Wood-Anderson record simulation.



**Fig. 11.58** Vertical-component short-period filtered BB seismograms (4<sup>th</sup> order Butterworth high-pass filter,  $f = 0.7$  Hz, normalized amplitudes) from a local mining-induced earthquake at the French-German border recorded at 11 GRSN, GERES and GEOFON stations ( $M_l = 3.7$ ; D between 80 km (WLF) and 501 km (GEC2)). Note the well developed more long-period and dispersive  $R_g$  group following  $S_g$  in the record of WLF. Usually one does not find this phase in records of deeper tectonic earthquakes (e.g., Figs. 11.55-11.57).

But Fig. 11.59 shows the great variability in the appearance of waveforms and relative amplitudes in near-earthquake recordings, which often deviate from such average rules. Even seismic records at the same station from two different sources at nearly the same distance and with similar azimuth may look very different. This may be because the waves from the two earthquakes travel along slightly different paths through the highly heterogeneous Alpine mountain range. However, the fault-plane orientation and related energy radiation with respect to the different take-off angle of Pn and Pg may also have been different for these two earthquakes.



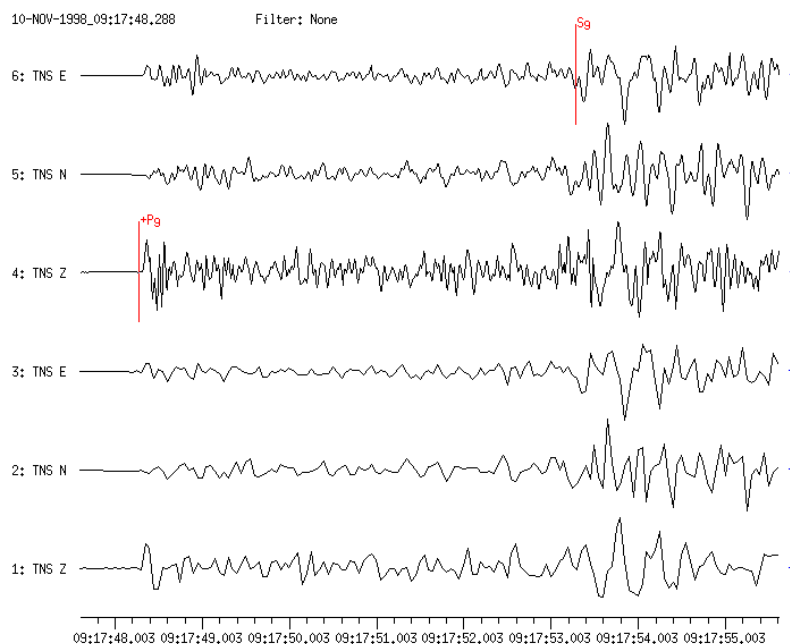
**Fig. 11.59** Comparison of Z-component short-period filtered records at station MOX, Germany, of two earthquakes in Northern Italy (trace 1: 28 May 1998; trace 2: 24 Oct. 1994) at about the same epicentral distance ( $D = 505$  km and  $506$  km, respectively) and with only slightly different backazimuth ( $BAZ = 171^\circ$  and  $189^\circ$ , respectively). Note the very different relative amplitudes between Pn and Pg, due to either crustal heterogeneities along the ray paths or differences in rupture orientation with respect to the different take-off angles of Pn and Pg rays.

The recognition and differentiation of the various crustal phases and the precision of onset-time picking can be significantly improved by stretching the time scale in digital records (compare, e.g., Figures 3a and 3c in DS 11.1). But local and regional earthquakes do not only appear in short-period recordings. Stronger near events with magnitudes above 4 usually generate already strong long-period waves (e.g., Figs. 11.11 and 11.13). In BB records of shallow earthquakes Pn and Sn will then be followed by well-developed surface-wave trains of even much larger amplitudes. Figure 6a in DS 11.1 shows a typical 3-component BB-velocity record at station GRA1, Germany ( $D = 10.3^\circ$ ) of such an earthquake in Italy. Figure 6c shows the respective BB records at 10 stations of the GRSN ( $D = 8^\circ - 12^\circ$ ). Pg and Sg are no longer recognizable.

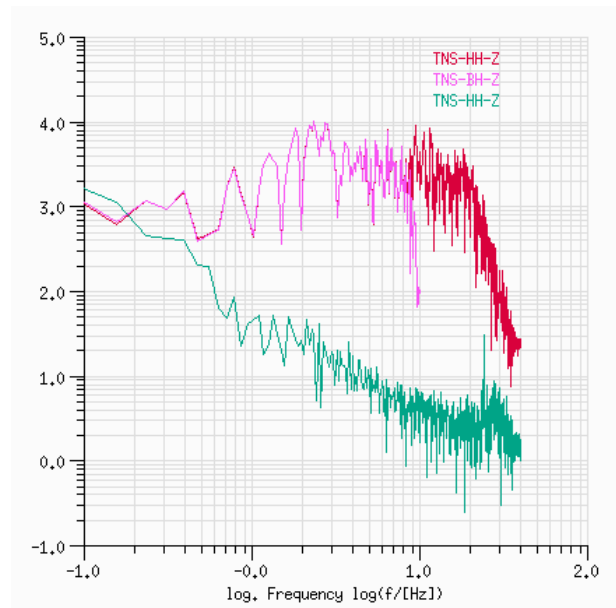
With the exception of the P- and S-wave triplications and thus complicated and elongated P- and S-wave groups at distances beyond  $15^\circ$  the records in the far regional range look rather simple. Usually, only P, S, and surface waves are clearly developed (Fig. 11.53 and Figures 2b and 4c in DS 11.2). The surface waves of shallow earthquakes are strong, clearly separated from the S waves and very useful for surface wave magnitude estimation, with  $M_s(BB)$  even down into the near regional range. In the case of deep earthquakes and thus negligible surface waves the the core reflection PcP may sometimes also be observed in the far regional distance range ( $6^\circ$ - $20^\circ$ +) in short-period seismograms of stronger events ( $M_l > 4$ ). PcP, which gives a good control of source depth, can be identified in array recordings because of its very small slowness (e.g., Fig. 9.14 in Chapter 9).

In general, regional stations and local networks complement each other in the analysis of smaller sources at local distances. Additionally, source processes and source parameters can be estimated using local station data. For this purpose, first motion polarities (compression c or +, dilatation d or -) for phases Pg, Pn, Sg and amplitude ratios (P/SV) should be measured for fault plane solution and moment tensor inversion (see 3.4 and 3.5). In regions with poor azimuthal and distance coverage by stations, the mean precision of location may be several kilometers and source depths may then only be estimated via depth phase modeling and fitting (Fig. 2.85).

An important aspect to consider in digital recordings and data analysis of local and regional seismograms is the sampling rate. Sampling with more than 80 s.p.s. is generally suitable for near seismic events. With lower sampling rate some of the most essential information about the seismic source process such as the corner frequency of the spectrum and its high-frequency decay, may be lost. Fig. 11.60 gives an example, which shows that only the 80 Hz data stream with a Nyquist frequency of 40 Hz allows the corner frequency near 20 Hz to be determined. In some regions, however, or when studying small local earthquakes, still higher frequencies have to be analyzed. This may require sampling rates between 100 and 250 Hz.



a)



b)

**Fig. 11.60 a):** BB-velocity seismogram from a local earthquake near Bad Ems, Germany (11 Oct. 1998;  $M_l = 3.2$ ) recorded at the GRSN station TNS ( $D = 40$  km). Different sampling rates were used for data acquisition. Traces 1 – 3 were sampled at 20 Hz and traces 4 – 6 at 80 Hz. In the records with the higher sampling rate, the waveforms are much more complex and contain higher frequencies. The high frequency content is suppressed with the lower sampling rate. **b):** Fourier spectrum of traces 1 (sampling rate 20 Hz – pink) and 4 (sampling rate 80 Hz – red). The lower sampling rate cuts off the high-frequency components of the seismic signal. Thus the corner frequency of the signal at about 20 Hz and the high-frequency decay could not be determined from the pink spectrum. The green spectrum represents the seismic noise.

Note that besides the crustal phases so far mentioned a very different type of seismic wave, the so-called T-phase, may appear in records of oceanic earthquakes at near coastal stations. The T-phase forms a rather long envelope-shaped wave train with a faint onset and ending and follows well behind the crustal phase wave train. For records and more explanations about the origin of these phases Chapter 2, section 2.6.5, Figs. 2.71 and 2.72.

In Box 1 a summary is given of essential features that can be observed in records of local and regional seismograms. For more records in this distance range see DS 11.1.

### BOX 1: General rules for local and regional events

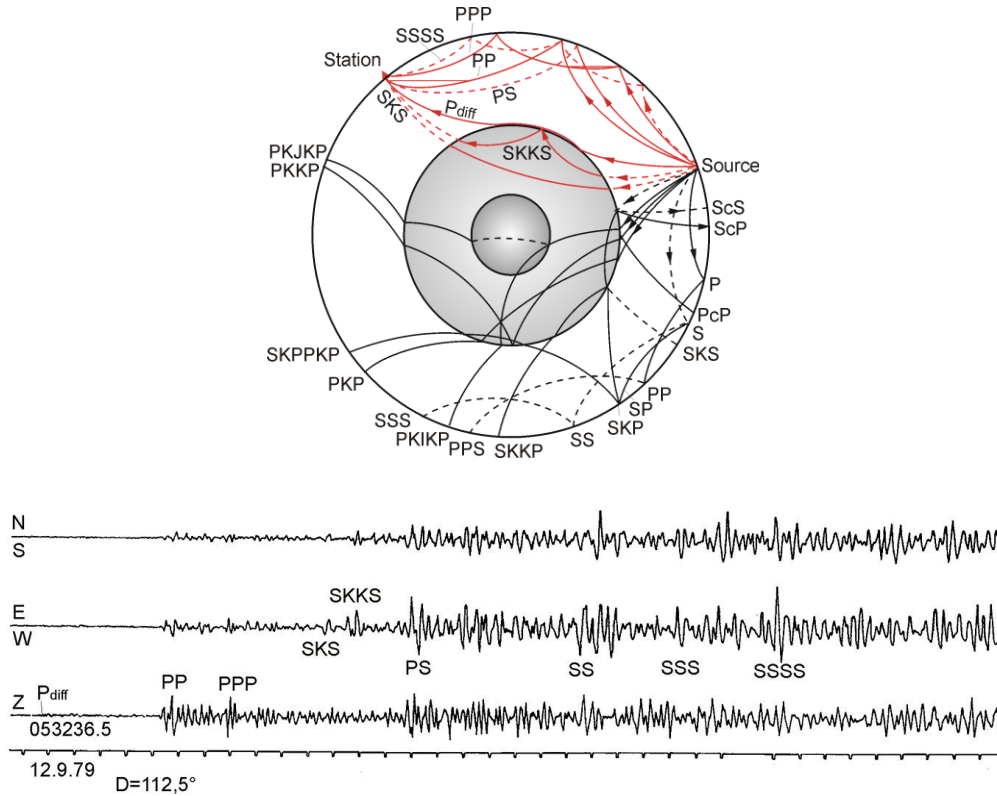
- The frequency content of local events ( $D < 2^\circ$ ) is usually high ( $f \approx 0.2 - 100$  Hz). Therefore they are best recorded on SP or SP-filtered BB instruments with sampling rates  $f \geq 80$  Hz. The overall duration of short-period local and regional ( $D < 20^\circ$ ) seismograms ranges between a few seconds and several minutes.
- Stronger local/regional sources radiate long-period energy too and are well recorded by BB and LP seismographs. In the far regional range the record duration may exceed half an hour (Chapter 1, Fig. 1.2).

- Important seismic phases in seismograms of local sources are Pg, Sg, Lg and Rg and in seismograms of regional sources additionally Pn and Sn, which arrive beyond  $1.3^{\circ}$ - $2^{\circ}$  as the first P- and S-wave onsets. The P waves are usually best recorded on vertical and the S waves on horizontal components.
- Note that Pg is not generally seen in records from sources in the oceanic crust. Also, deep (sub-crustal) earthquakes lack local and regional crustal phases.
- For rough estimates of the epicentral distance D [km] of local sources, multiply the time difference Sg-Pg [s] by 8, and in the case of regional sources the time difference Sn-Pn [s] by 10. For more accurate estimates of D use local and regional travel-time curves or tables or calculations based on more appropriate local/regional crustal models.
- The largest amplitudes in records of local and regional events are usually the crustal channel waves Lg (sometimes even beyond  $15^{\circ}$  in continental platform areas), and for near-surface sources the short-period fundamental Rayleigh mode Rg. For near-surface explosions or mining-induced earthquakes, Rg, with longer periods than Sg, may dominate the record, however usually not beyond  $4^{\circ}$ .
- In records of oceanic earthquakes at near coastal stations occasionally T phases may be observed.
- For routine analysis the following station/network readings should be made: (1) the onset time and polarity of observed first motion phases; (2) onset times of secondary local and regional phases; (3) local magnitude based either on maximum amplitude or duration. If local/regional calibration functions, properly scaled to the original magnitude definition by Richter (1935) or the IASPEI standard formula, are not available it is recommended to use the original Richter equation and calibration function, together with local station corrections.

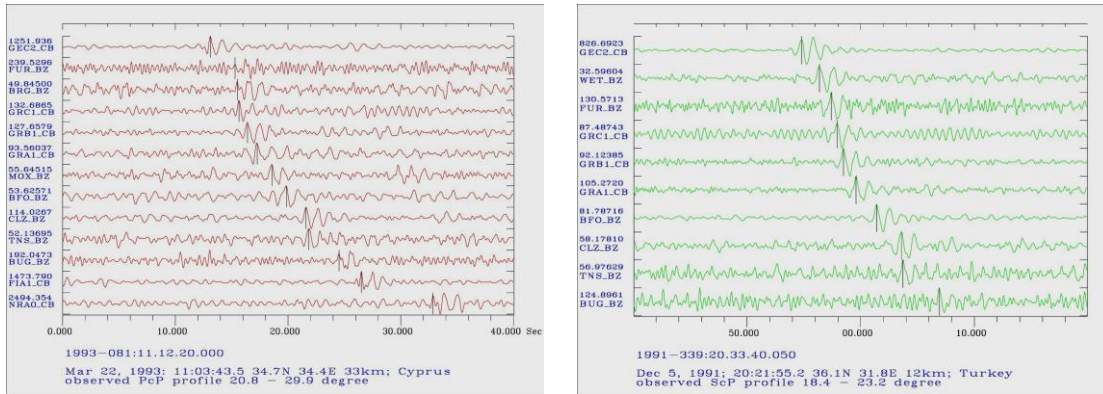
### 11.5.3 Teleseismic earthquakes ( $20^{\circ} < D < 180^{\circ}$ )

#### 11.5.3.1 Distance range $20^{\circ} < D \leq 100^{\circ}$

The main arrivals at this distance range up to about  $80^{\circ}$  have traveled through the lower mantle and may include reflections from the core-mantle boundary (CMB) (Figs 11.61, 11.62 and 11.64). The lower mantle is more homogeneous than the upper mantle (Fig. 2.79). Accordingly, P and S waves and their multiples form rather simple long-period seismograms (Figs. 11.12 and 11.63; see also files 5 and 6 in IS 11.3). Between  $30^{\circ}$  and  $55^{\circ}$ , the waves reflected from the core (e.g., PcP, ScP etc.) are also often recorded as sharp pulses on short-period records, particularly on records of deep earthquakes where depth phases appear well after the core reflections (Fig. 11.21). At around  $40^{\circ}$ , the travel-time curve of PcP intersects those of PP and PPP (Figs. 11.8 and 11.64) and in horizontal components PcS intersects S, and ScS intersects SS and SSS (Fig. 11.8). This complicates proper phase separation, at least for the later phases on long-period records, where SS and SSS may be strong. ScP, however, may also be rather strong on short-period vertical components (Fig. 11.64). The amplitudes of the core reflections decrease for larger distances but they may be observed up to epicentral distance of about  $80^{\circ}$  (ScP and ScS) or  $90^{\circ}$  (PcP), respectively, beyond which ScS merges with the travel-time curves of SKS and S and PcP with that of P (see Figs. 11.8, 11.20, 11.21 and 11.66). Note that PP, PS, SP and SS are Hilbert-transformed (see 2.5.4.3). Their onset and amplitude picks can be improved by inverse Hilbert transformation, which is part of modern analysis software such as Seismic Handler (SH and SHM).

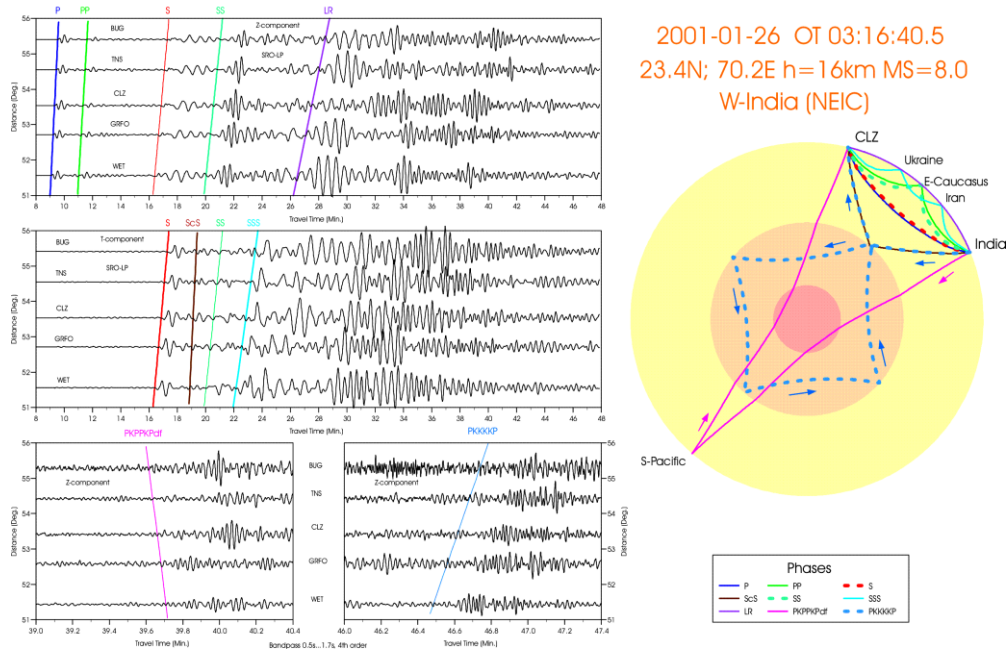


**Fig. 11.61** Seismic ray paths through the mantle and core of the Earth with the respective phase names according to the international nomenclature (Chapter 2, Figs. 2.58 and 2.60, and IS 2.1 for phase names and their definition). The red rays relate to body wave onsets in the 3-component analog Kirnos SKD BB-displacement record below from an earthquake at  $D = 112.5^\circ$  at station MOX, Germany.

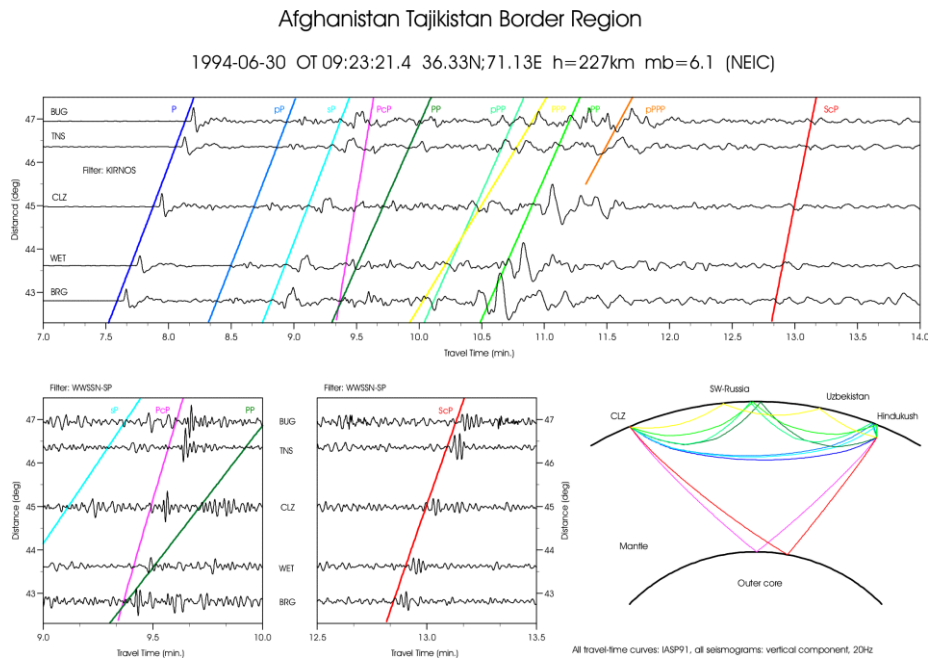


**Fig. 11.62 Left:** Array and single station observations of PcP for an  $m_b = 5.3$  and  $h = 29.7$  km deep Cyprus earthquake on 1993 March 22. The beam traces of the arrays are named by their reference stations GEC2 (GERES), GRA1, GRB1, GRC1 (Gräfenberg), FIA1 (FINES), and NRA0 (NORES), respectively. All data are bandpass filtered between 0.8 and 3 Hz. The un-restituted, velocity proportional traces are scaled to their maximum amplitude, which is given in digital counts. **Right:** The same for ScP observations of an  $m_b = 5.0$  and  $h = 106.6$  km deep earthquake in Turkey on 1991 December 5. The beam traces of the arrays are named by their reference stations GEC2 (GERES), GRA1, GRB1, and GRC1 (Gräfenberg), respectively. Color versions of Figures 2 and 3 of Schweitzer (2002); © Geophys. J. Int.





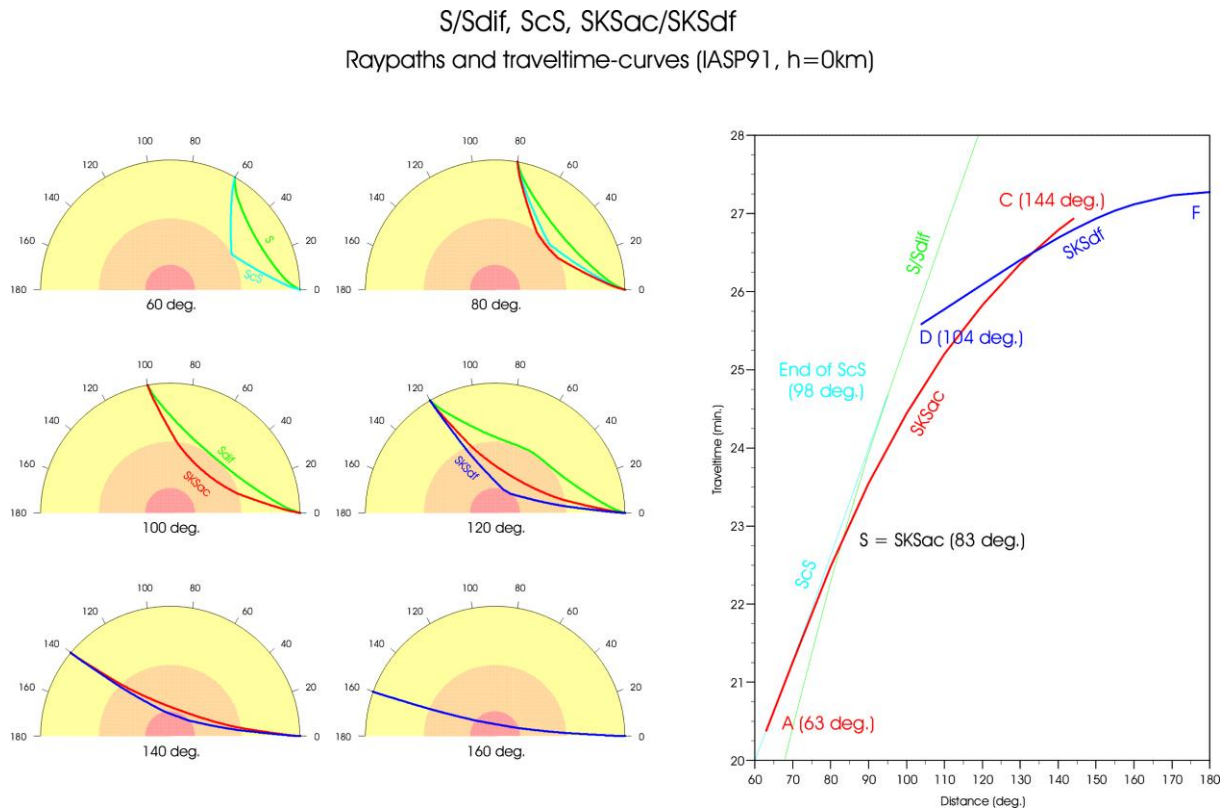
**Fig. 11.63** Long-period Z- (top record section) and T-component seismograms (middle record section) of a shallow earthquake in western India recorded in the distance range 51° to 56° at stations of the GRSN. The two cut-out vertical component short-period record sections at the bottom left and middle show the in the later part of the surface waves arriving multiple core phases PKPPKpdf (= P'P') and P4KP. For all phases identified in these records the related ray paths are depicted on the right side diagram (for animation see file 5 in IS 11.3).



**Fig. 11.64** Vertical-component Kirnos SKD BB-displacement (top record section) and WWSSN-SP seismograms (lower left two record sections) from an intermediate depth ( $h = 227$  km) earthquake in the Afghanistan-Tajikistan border region recorded at stations of the GRSN. Besides P the depth phases pP, sP, pPP, sPP and pPPP the core reflections PcP and ScP are clearly visible, particularly on the short-period records. The ray traces of these phases are shown in the lower right diagram (see also file 4 in IS 11.3 and related animation).

**Remark:** For many of the above and following record sections with accompanying ray diagrams one finds in DS 11.2 and 11.3 more clear records with better time resolution.

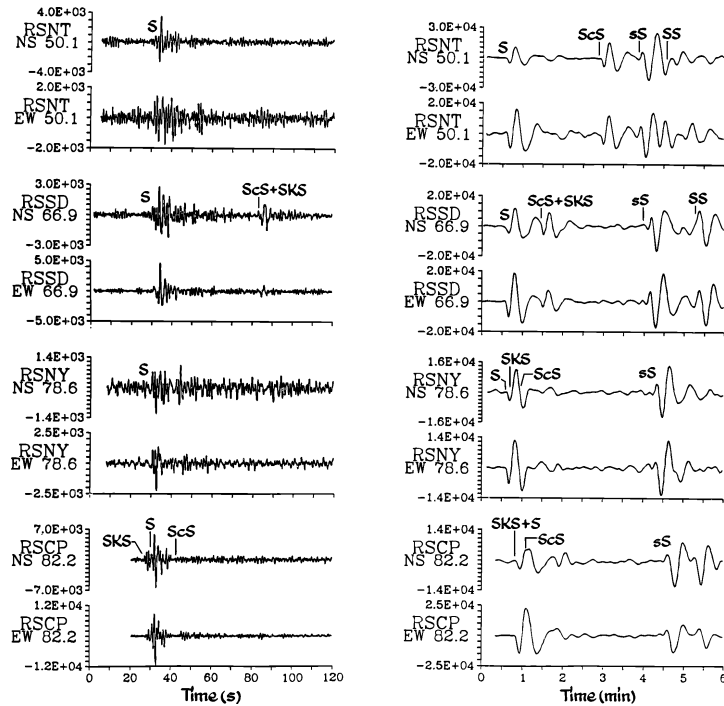
Fig. 11.65 shows the ray paths for S, ScS and SKS and their related travel-time curves according to the IASP91 model for the distance range from 60° to 180° and Fig. 11.66 both short- and long-period records of these waves between 50° and 80°.



**Fig. 11.65** Ray paths for S, ScS and SKS and their related travel-time curves according to the IASP91 model for the distance range from 60° to 180°.

Arrays and network records, which also allow f-k and vespagram analysis, are very useful for identifying the core reflections PcP, ScP and ScS because their slownesses differ significantly from those of P, S and their multiple reflections (Fig. 11.62 and Figures 6a and b and 7b in DS 11.2). Surface reflections PP, PPP, SS and SSS are well developed in this distance range in long-period filtered records and converted waves PS/SP at distances above 40°. Sometimes the surface reflections are the strongest body-wave onsets at large distances (Figures 10c and 11b in DS 11.2). Their identification can be made easier when network records are available so vespagram analysis can be used (e.g., Figure 11c in DS 11.2). In short-period filtered network records it is sometimes possible to correlate in this distance range multiple reflected core phases such as PKPPKP (or P'P'), SKPPKP and SKPPKPPKP (Chapter 1, Fig. 1.10)





**Fig. 11.66** SP (left) and LP (right) horizontal-component seismograms from a deep-focus earthquake in the Sea of Okhotsk (20 April 1984,  $m_b = 5.9$ ,  $h = 588$  km) recorded at stations in the distance range  $50.1^\circ$  to  $82.2^\circ$ . Note the different amplitude scaling. Accordingly, the amplitudes of various transverse phases are 2 to 10 times larger in long-period records when compared with short-period records. Four distinct phases are identified: S, ScS, sS and SS. SKS, however, which emerges at distances larger than  $60^\circ$ , overlaps with ScS between about  $65^\circ$  and  $75^\circ$ . S, ScS and SKS start to coalesce as distance increases toward  $82^\circ$ . Beyond this distance SKS arrives before S, SKKS and ScS (reprinted from *Anatomy of Seismograms*, Kulhánek, Plate 40, p.137-138; © 1990; with permission from Elsevier Science).

Beyond  $83^\circ$  SKS arrives for near surface sources - but earlier already for deeper earthquakes - ahead of S and its amplitude relative to S increases with distance. Network and array analysis yields different slowness values for S and SKS because of their diverging travel-time curves (Fig. 11.65). This helps to identify these phases correctly. Note that the differential travel time SKS-P increases only slowly with distance (Fig. 11.8). Misinterpretation of SKS as S may therefore result in an underestimation of D by up to  $20^\circ$ ! Since SKS is polarized in the vertical plane it can be observed and separated well from S in radial and vertical components of rotated seismograms (Fig. 11.67 and Figures 10c, 13e, 14e, and 15b in DS 11.2). This also applies to PcS and ScP that are polarized in the vertical propagation plane as well (Fig. 11.64).

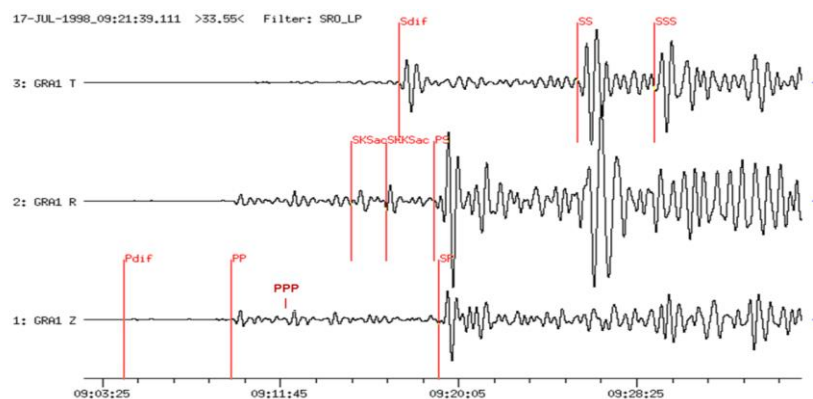
In the distance range between about  $30^\circ$  and  $105^\circ$  multiple reflected core phases P'N or between about  $10^\circ < D^\circ < 130^\circ$  the phases PNKP, with N-1 reflections either at the free surface (P'N) or from the inner side of the core-mantle boundary (PNKP), respectively, may appear in short-period records some 13 min to 80 min after P. An example for PKPPKP (P'P') and PKKKKP (P4KP) is given in Fig. 11.63. These phases are particularly strong near caustics, e.g., P'P' (Fig. 11.77) and P'P'P' (P'3) near  $70^\circ$  and PKKP near  $100^\circ$  (Fig. 11.78) but they are not necessarily observable at all theoretically allowed distances. Figures 9 and 10 in EX 11.3, however, document the rather wide distance range of real observations of these

phases at station CLL (for P'P' between 40° and 105° and for PKKP between 80° and 126°). Note the different, sometimes negative slowness of these phases. More record examples, together with the ray paths of these waves, are presented in a special section on late core phases (11.5.4). For differential travel-time curves of PKKP-P and PKPPKP-P Figures 9 and 10 in EX 11.3. Also PKiKP, a weak core phase reflected from the surface of the inner core (ICB), may be found in short-period array recordings throughout the whole distance range, about 4.5 to 12 min after P. Its slowness is less than 2s/°. Beyond 95°, P waves show regionally variable, fluctuating amplitudes. Their short-period amplitudes decay rapidly (Fig. 11.10) because of the influence of the core (core-shadow) while long-period P waves may be diffracted around the curved core-mantle boundary (Pdif, Figs. 11.70 and 11.74 as well as Figures 1, 2, 4b and 6c and b in DS 11.3).

In any case, comprehensive seismogram analysis should be carried out for strong earthquakes which produce many secondary phases. Unknown phase arrivals should also be reported for further investigations into the structure of the Earth. When reporting both identified and unknown phases to international data centers the IASPEI-proposed international nomenclature should strictly be observed (see IS 2.1).

### 11.5.3.2 Distance range $100^\circ < D \leq 144^\circ$

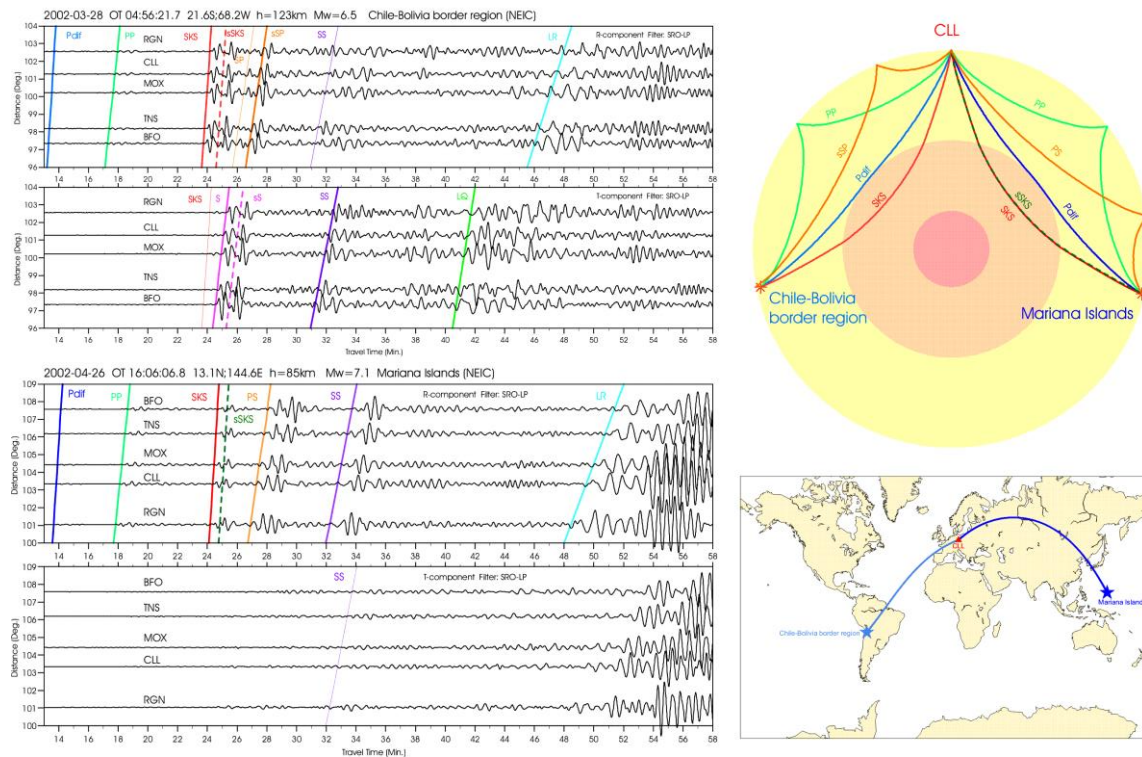
Within this distance range, the ray paths of the P waves pass through the core of the Earth. Due to the large reduction of the P-wave velocity at the core-mantle boundary (CMB) from about 13.7 km/s to 8.0 km/s (Fig. 2.79 in Chapter 2) seismic rays are strongly refracted into the core (i.e., towards the normal at this discontinuity). This causes the formation of a "core shadow". This "shadow zone" commences at an epicentral distance around 100°. The shadow edge is sharper for short-period P and S waves than for long-period P and S waves that are diffracted much further around the curved CMB (compare Figures 6b and c as well as 7a and b in DS 11.3). For strong earthquakes Pdif and Sdif may be observed out to distances of more than 150° (Figs. 11.67, 11.68, 11.70 and 11.74).



**Fig. 11.67** SRO-LP filtered 3-component seismograms at station GRA1, Germany, in  $D = 117.5^\circ$  from an earthquake in Papua New Guinea (17 July 1998,  $M_s = 7.0$ ). The N and E components have been rotated into the R and T directions. Phases Pdif, PP, PPP and a strong PS and SP are visible on the vertical component, whereas the phases SKS, SKKS and PS, which are polarized in the vertical propagation plane, are strong on the radial (R) component (as are PP and PPP). In contrast, Sdif, SS and SSS are strong on the transverse (T) component.

Fig. 11.68 shows rotated (R-T) horizontal component SRO-LP recordings at GRSN stations from two intermediate deep events in the Chile-Bolivia border region and in the Mariana Islands, respectively. The related ray paths are depicted in the upper part. The records cover the transition from the P-wave range into the P-wave core shadow. Magnified cut-outs, also of the related Z-component records, are presented for both earthquakes in Figures 1 and 2 of DS 11.3. They show more clearly the first arriving longitudinal waves and their depth phases. The following conclusions can be drawn from a comparison of these figures:

- Pdif arrives about 4 minutes (at larger distance up to 6 min; Fig. 11.74) earlier than the stronger PP;
- The largest phases (Figs 11.68 and . 11.71) are usually PP, PPP, PS, SP, Sdif, SKS, SKKS, SS and SSS;
- SKS is the first arriving shear wave, followed by SKKS, SP or PS (and the related depth phases), all on the R component;
- S/Sdiff and SS may be strong(est) in T or R, or even in both components, depending on the SV/SH ratio of shear-wave energy radiated by the source.

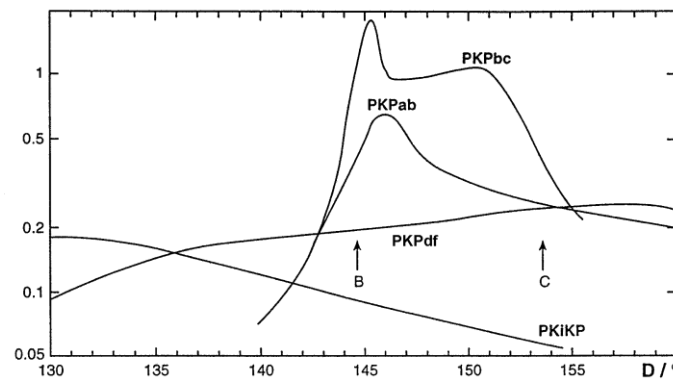


**Fig. 11.68** SRO-LP filtered records of GRSN stations on R and T components in the distance range between  $96^\circ$  and  $109^\circ$  from two earthquakes on opposite backazimuth (Chile-Bolivia Border region and Mariana Islands, respectively; source data according to the NEIC).

If no Pdif is observed, PKiKP is the first arrival in short-period records up to  $113^\circ$  (Figure 3b in DS 11.3). For distances beyond  $114^\circ$  PKiKP follows closely after PKPdf (alternatively termed PKIKP). The latter has traveled through the outer and inner core and arrives as first onset for  $D \geq 113^\circ$ . PKPdf is well recorded in short-period seismograms but usually with emergent onsets and, up to about  $135^\circ$  distance, still with weaker amplitudes than PKiKP. Fig. 11.69 shows theoretical amplitude-distance relationship between PKiKP, PKPdf and the other

direct core phases PKPab and PKPbc, which appear with largest amplitudes beyond  $143^\circ$ . Fig. 11.70 depicts the ray paths and travel-time curves of Pdif, PKiKP, PKPdf, PKPab and PKPbc (for more complete ray paths see IS 2.1). Also PKKP (with its branches ab and bc) is often clearly recorded between  $110^\circ$  and  $125^\circ$  (Figures 3c and d in DS11.3).

Fig. 11.71 presents records of GRSN stations in the distance range  $121^\circ$  to  $127^\circ$  from an earthquake at intermediate depth ( $h = 138$  km) in the region of New Britain (see file 8 in IS 11.3 and animation CD). They show the PKPdf arrivals about 3.5 min after Pdif together with the dominant phases in this range, namely PP, PPP, PS, PPS and the Rayleigh-wave arrival LR in the Z component and the SS, SSS and the Love-wave arrival LQ in the T component. Also shown, together with the ray paths, is the core phase P4KPbc, which has been reflected 3 times at the surface of the Earth, and which is recognizable only on the short-period filtered vertical component. Between about  $128^\circ$  and  $144^\circ$  some incoherent waves, probably scattered energy from “bumps” or heterogeneities at the CMB or in the D” layer above it (Chapter 2, Fig. 2.79), may arrive as weak forerunners up to a few seconds before PKPdf. They are termed PKPpre (old PKhKP). PKPdf is followed by clear PP, after about 2 to 3 minutes, with SKP or PKS arriving about another minute later (Fig. 11.72).



**Fig. 11.69** Theoretical smoothed amplitude-distance relationships for the core phases PKiKP, PKPdf, PKPab and PKPbc as calculated for the model 1066B in the distance range  $130^\circ$  to  $160^\circ$  (modified from Houard et al., 1993, *Bull. Seism. Soc. Am.*, Vol. 83, No. 6, Fig. 4, p. 1840, © with permission Seism. Soc. Am.).

SKP/PKS have their caustics at about  $132^\circ$  and thus, near that distance, usually have rather large amplitudes in the early part of short-period seismograms (Fig. 11.72). For medium-sized earthquakes these phases may even be the first ones to be recognized in the record and be mistaken for PKP. Note that for near-surface events SKP and PKS have the same travel time, but with the former having relatively larger amplitudes in the Z component whereas PKS is larger in the R component. For earthquakes at depth, PKS and SKP separate, with the latter arriving the earlier the deeper the source (Fig. 11.72). Beyond  $135^\circ$  there are usually no clear phases between SKP and SS. Misinterpretation (when Pdif is weak or missing) of PP and SKS or PS waves as P and S may in this distance range result in strong underestimation of D (up to  $70^\circ$ ). This can be avoided by looking for multiple S arrivals (SS, SSS) and for surface waves which follow more than 40 min later (Table 5 in DS 3.1). For more records see DS 11.3, examples 1 to 7.

Figure 1 displays seismic tomography results for the New Britain region. The figure includes four panels of seismic waveforms (BFO, FUR, GRFO, MOX, CLL) and a map of the region.

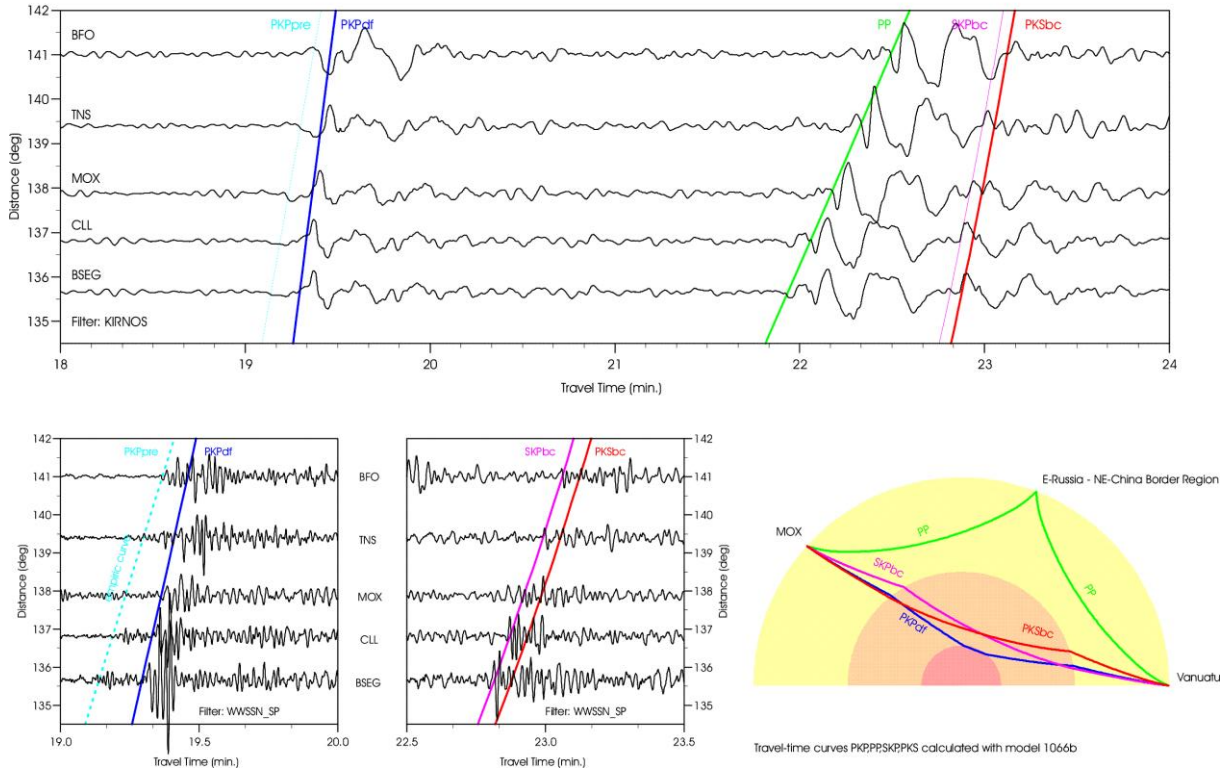
The top-left panel shows Z-component filtered SRO\_LP waveforms with phases PdF, PKPcF, JP, PFP, PS, PPS, and LR marked. The top-right panel shows T-component filtered SRO\_LP waveforms with phases SS and SSS marked. The bottom-left panel shows Z-component filtered SRO\_LP waveforms with phases PdF, PKPcF, and 4(PKPbc) marked. The bottom-right panel shows Z-component filtered WWSBL\_LP waveforms with phases PdF, PKPcF, JP, PFP, PS, PPS, SS, SSS, LQ, and LR marked.

The map on the right shows the New Britain region with various seismic phases and their travel times. The map includes labels for Central Russia, Mongolia, E China Sea, Philippine Sea, W of Isonaka Is., New Britain, Southern Pacific Ocean, and Brazil.

99



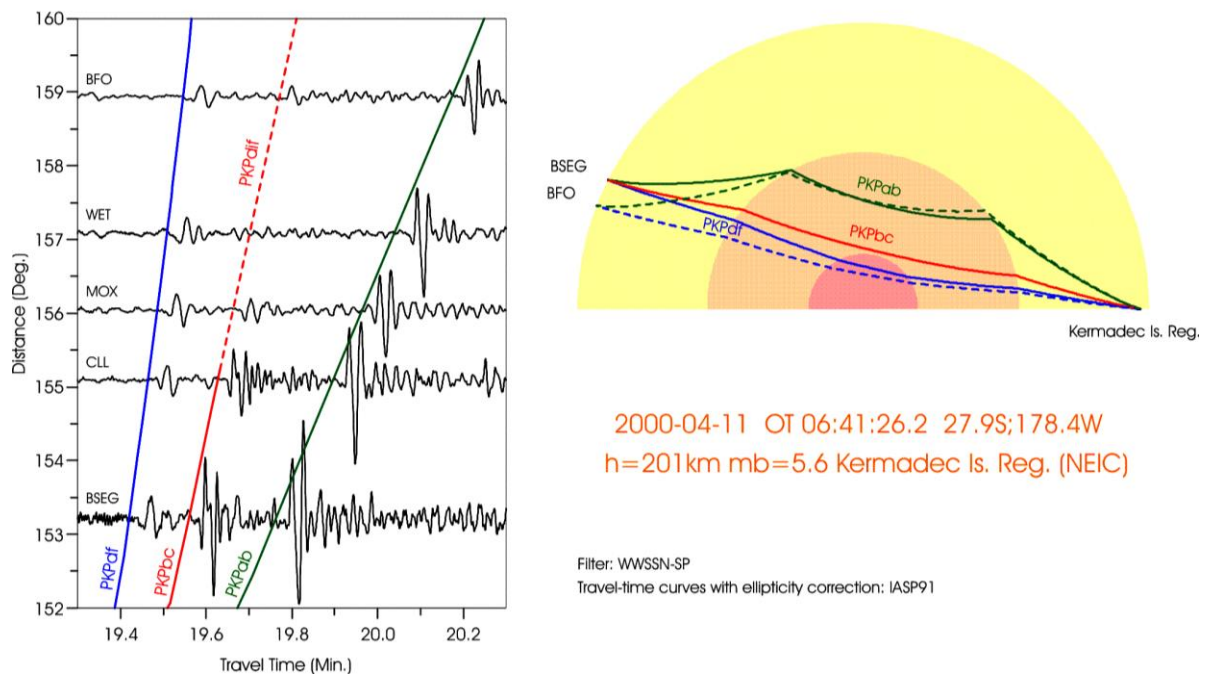
1998-09-21 OT 12:09:39.7 13.6S;166.8E h=33km mb=6.0 Vanuatu Islands (NEIC)



**Fig. 11.72** Vertical-component seismograms at GRSN stations recorded in the distance range 135° to 141°. **Top:** Kirnos SKD BB-displacement; **Bottom:** from left to right - WWSSN-SP filtered records showing the onsets of PKPpre, PKPpdf, SKPbc and PKSbc as well as the ray paths of the plotted phases.

### 11.5.3.3 Core distance range beyond 144°

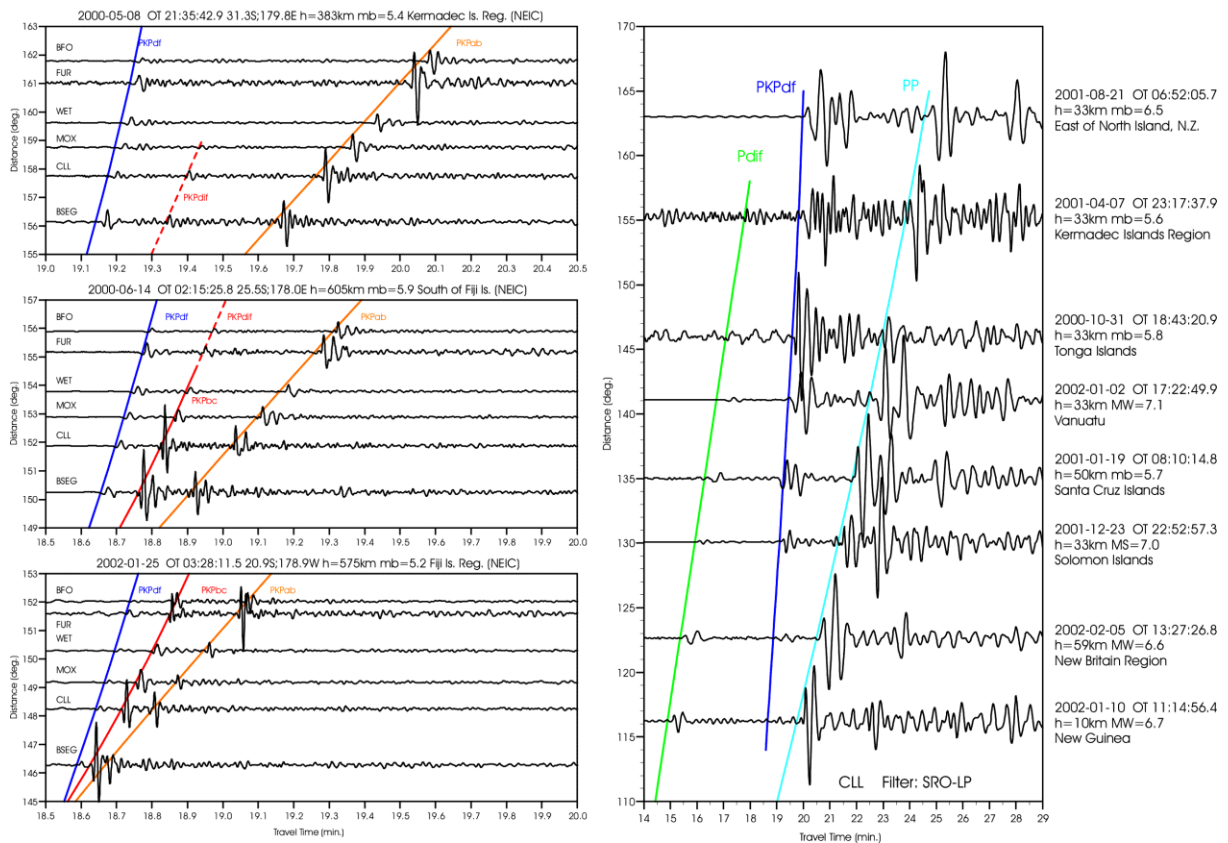
Between 130° and 143° the first onsets of longitudinal core phases are relatively weak and complex in short-period records, but their amplitudes increase strongly towards the caustic around 144°. At this epicentral distance three PKP waves, which have traveled along different ray paths through the outer and inner core, namely PKPpdf, PKPbc (old PKP1) and PKPab (old PKP2) arrive approximately at the same time (Fig. 11.70 above) so their energies superimpose to give a strong arrival with amplitudes comparable to those of direct P waves at epicentral distances around 40° (compare with Fig. 11.10). Beyond the caustic the travel-time curves of these three PKP waves split into the branches AB (or ab), BC (or bc) and DF (or df) (Fig. 11.70). Accordingly, the various arrivals can be identified uniquely by attaching to the PKP symbol for a direct longitudinal core phase the respective branch symbol (IS 2.1 and Figs. 11.70 and 11.73). Note that the PKPbc branch shown in Figs. 11.70 between the point B and C is ray-theoretically not defined beyond 155°. However, in real seismograms one often observes weak onsets between PKPpdf and PKPab up to about 160° or even slightly beyond in the continuation of the PKPbc travel-time curve. This phase is a PKP wave diffracted around the inner core boundary (ICB) and named PKPdif (IS 2.1 and Fig. 11.76 and 11.77).



**Fig. 11.73 Left:** Records of the direct core phases PKPdf, PKPbc and PKPab as well as of the diffracted phase PKPdif from a Kermadec Island earthquake at stations of the GRSN in the distance range between 153° and 159°; **right:** PKP ray paths through the Earth.

The relative amplitudes between the three direct longitudinal core phases change with distance. In SP records these three phases are well separated beyond 146°, and PKPbc is the dominant one up to about 153° though the separation between these three phases is not clear within this range in LP records (Fig. 11.74 right and Figures 9a-c in DS 11.3).

On short-period records the phases PKPdf, PKPbc and PKPab are easy to identify on the basis of their typical amplitude and travel-time pattern. D can be determined with a precision better than 1.5° by using differential travel-time curves for the different PKP branches (see EX 11.3). On records of weaker sources, PKPbc is often the first visible onset because the PKPdf, which precedes the PKPbc, is then too weak to be observed above noise. On long-period records the different superimposed onsets may be recognizable only at distances larger than 153°. Then PKPab begins to dominate the PKP-wave group on short-period records (compare Figs. 11.74). Towards the antipode, however, PKPdf (PKIKP) becomes dominating again whereas PKPab disappears beyond 176°.



**Fig. 11.74** Short-period (left side) and long-period (SRO-LP; right side) filtered broadband records of PKP phases at GRSN stations in the distance range  $116^{\circ}$  to  $163^{\circ}$ . In the LP records additionally the onsets of Pdif and PP have been correlated with their travel-time curves.

On LP and BB records the dominant phases on vertical and radial components are PKP, PP, PPP and PPS while on the transverse component SS and SSS are dominant. For deep sources, their depth phases sSS and sSSS may be strongest (Fig. 11.75). Besides PP, which has traveled along the minor arc (epicentral distance  $D$ ) the phase PP2, which has taken the longer arc to the station ( $360^{\circ} - D$ ), may be observable, as well as phases such as PcPPKP and others (Fig. 11.76 as well as the animation file 10 in IS 11.3). SKKS, SKKKS, SKSP etc. may still be well developed on radial component records (Chapter 2, Figs. 2.60 and 2.61). The whole length of BB or LP seismograms in this distance range between the first onsets and the surface wave maximum is more than an hour (Tab. 5 in DS 3.1).

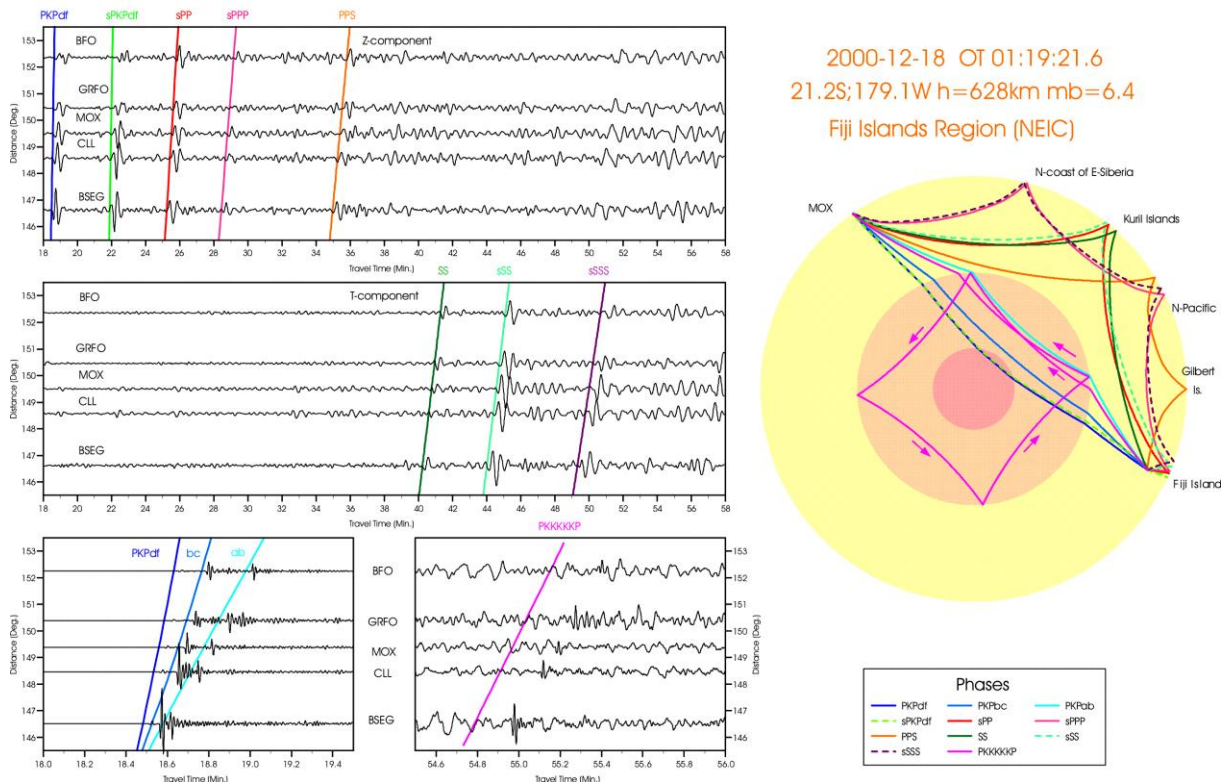
### 11.5.4 Late and very late core phases

For large magnitude sources, core phases reflected at the Earth's surface or at the core-mantle-boundary (CMB) may be observed in addition to the direct ones, sometimes with up to 4 (or even more) repeated reflections. These phases may be observed at practically all teleseismic distances with delays behind the first arriving P or PKP onsets ranging from about 10 minutes up to about 80 minutes, depending on the number of multiple reflections (for travel-time curves Chapter 2, Fig. 2.59). These phases are clearly discernible only in high-magnifying SP (or appropriately filtered BB) records (Fig. 2.57). Most frequently observable are the single surface

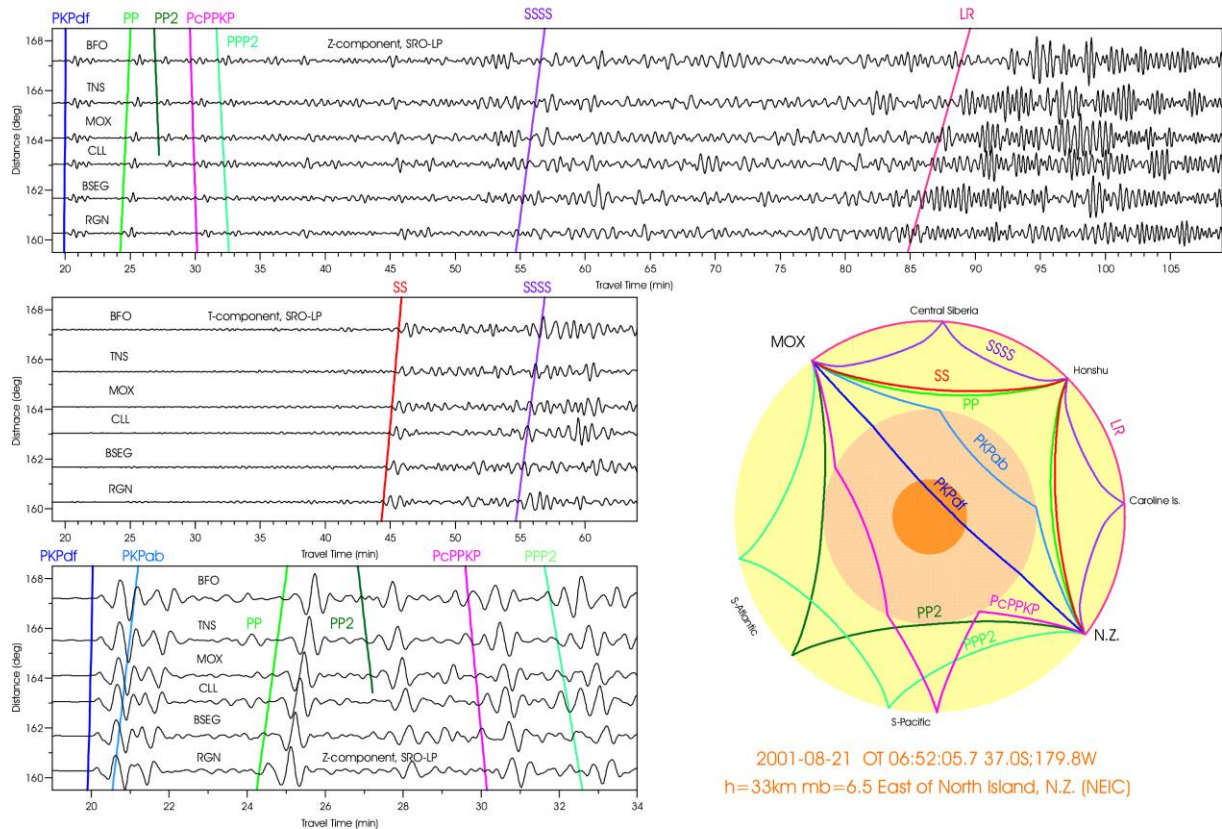


reflection P'P' (also termed PKPPKP), and the single reflection from the inner side of the core-mantle boundary, PKKP. As for the direct core phases, these multiple reflections develop different travel-time branches according to their different penetration depth into the outer core (see also ray tracing figures in IS 2.1). Figs. 11.77 and 11.78 show the ray paths for P'P' and PKKP waves, respectively, together with their related IASP91 travel-time curves (Kennett and Engdahl, 1991) for sources at depth  $h = 0$  km and  $h = 600$  km. Where there is more than one reflection the respective phases are often written P'N or PNKP, respectively, with the number of reflections being  $N-1$ . Ray paths and short-period record examples for P'N with  $N = 2$  to 4 are shown in Figs. 11.79 to 11.81 and for PNKP with  $N = 2$  to 5 in Figs. 11.81 and 82. Fig. 11.75 shows a P5KP (PKKKKKP), which has been four times reflected from the inner side of the CMB. It is observed nearly 37 min after PKP. The phase P7KP has been found in a record at Jamestown, USA, of an underground nuclear explosion on Novaya Zemlya in 1970. Complete ray diagrams of many direct, converted and multiply reflected core phases have been presented in IS 2.1.

Additionally, the figures presented below show that these late arrivals may still have a significant SNR. Since they appear very late and thus isolated in short-period records, station operators may wrongly interpret them as being P or PKP first arrivals from independent events. This may give rise to wrong phase associations and event locations, which, particularly in a region of low seismicity, may give a seriously distorted picture of its seismicity. This was demonstrated by Ambraseys and Adams (1986) for West Africa.

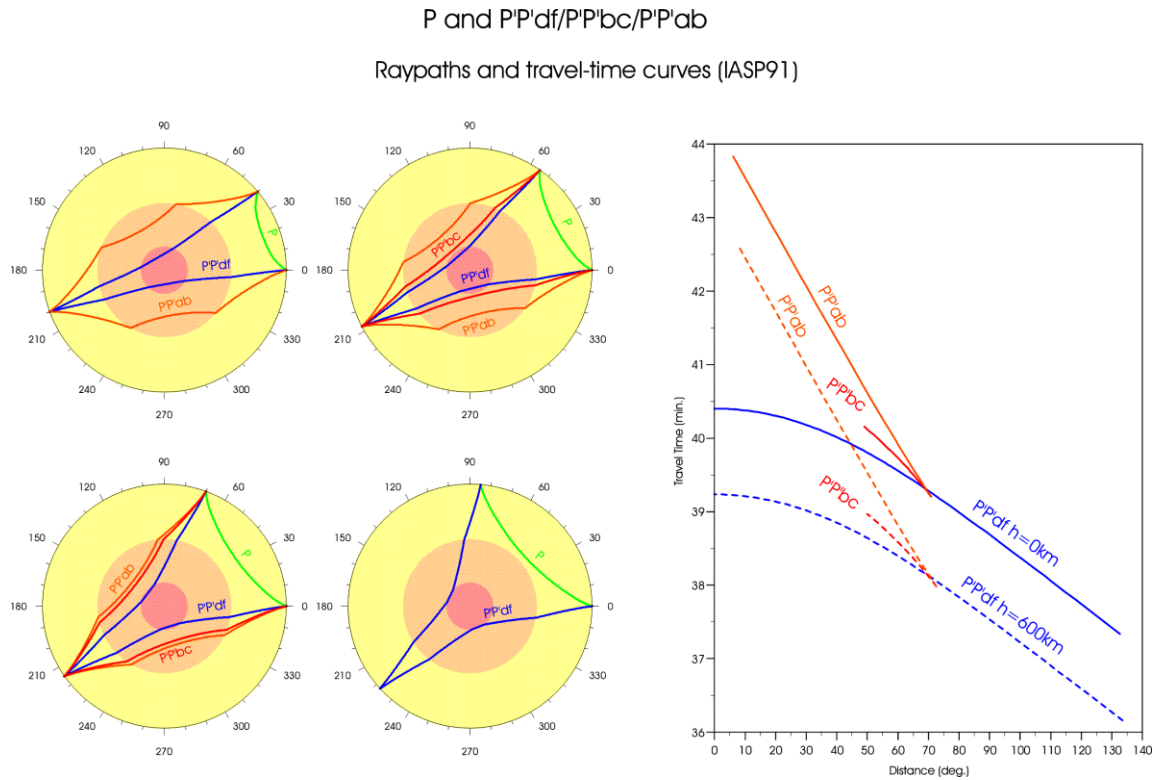


**Fig. 11.75** Records of GRSN stations of a deep earthquake in the Fiji Island region. **Right:** ray paths and source data; **Left:** Records on the Z component (LP top left and SP lower left) and T component (LP middle left) (see also animation file 9 in IS 11.3).

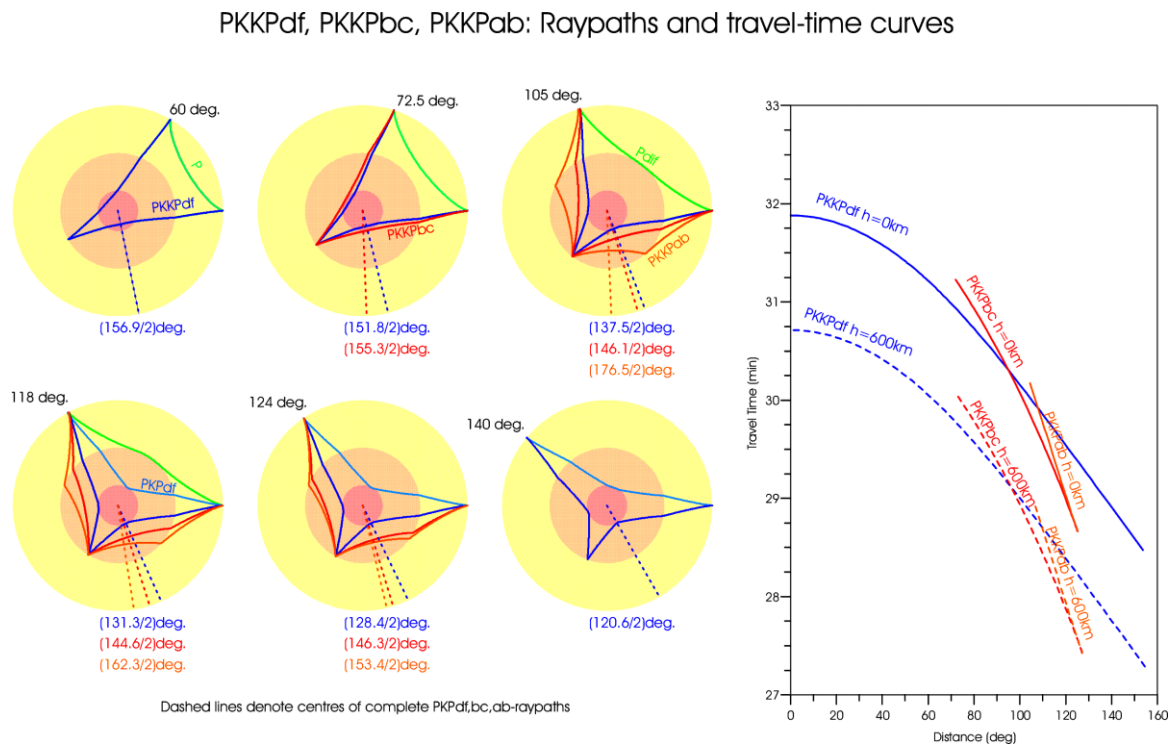


**Fig. 11.76** Records of GRSN stations of a shallow (crustal) earthquake east of the North Island of New Zealand. **Lower right:** ray paths to station MOX in Germany and source data; **Top and middle left:** Records on SRO-LP components Z and T). **Lower left:** The first 14 min of the SRO-LP Z-component P-wave record. For animation see file 10 in IS 11.3).

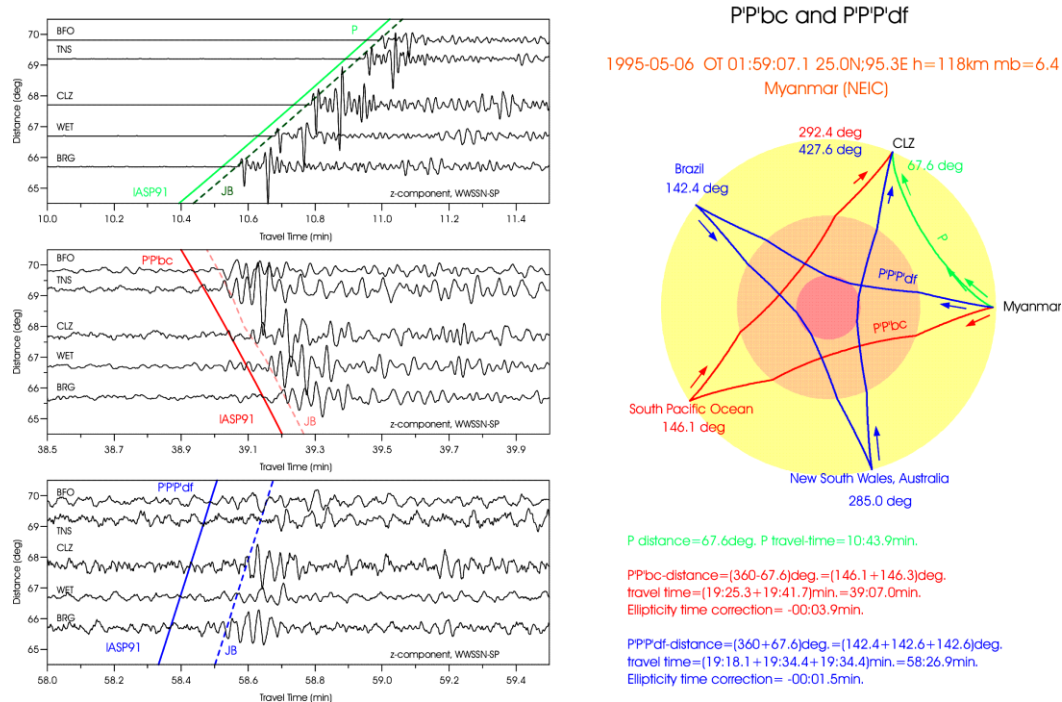
In Figs. 11.73 to 11.82 the theoretical travel-time curves, respectively arrival times for core phases have been plotted. For consistency, they are based on the travel-time model IASP91 (Kennett and Engdahl, 1991), as in all earlier record sections shown for the teleseismic distance range. An exception is Fig. 11.79, which shows additionally the theoretical travel-time curve for the JB model (Jeffreys and Bullen, 1940) as in the lower middle record segment too for P5KP. One recognizes that the model IASP91 yields onset times for core phases that tend to be up to several seconds earlier than the real onsets in the seismograms (see also Fig. 9b and c in DS 11.3). This applies to both direct and multiple reflected core phases. The agreement between real and theoretical onsets of core phases is better when using the JB model. Also S onsets arrive with respect to the IASP91 model a bit too late. The more recent model AK135 (Kennett et al., 1995) is more appropriate than IASP91. The JB model has regularly been used until recently for the location of teleseismic sources at the international data centers in Boulder (NEIC), Thatcham (ISC) and Moscow whereas the IDC of the CTBTO uses IASP91. AK135 is now the default model used by the ISC and the NEIC.



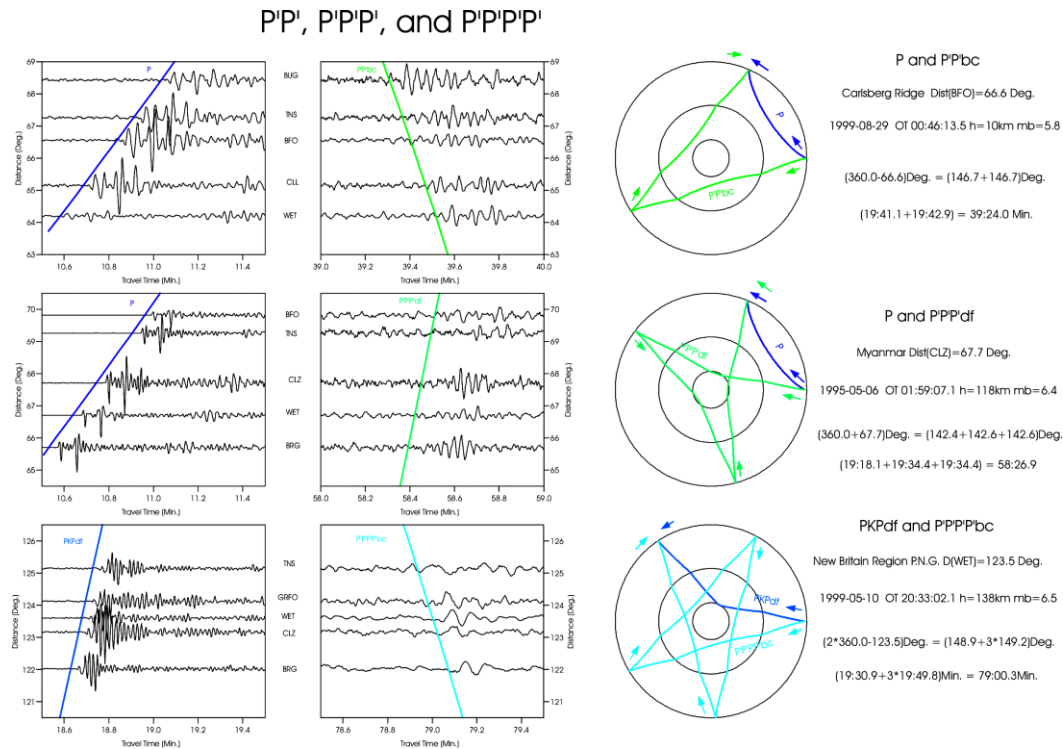
**Fig. 11.77** Ray paths and travel-time curves for P'P' according to the Earth model IASP91 (Kennett and Engdahl, 1991).



**Fig. 11.78** Ray path and travel-time curves for PKKP according to the Earth model IASP91 for epicentral distances of the source between  $60^\circ$  and  $140^\circ$ .



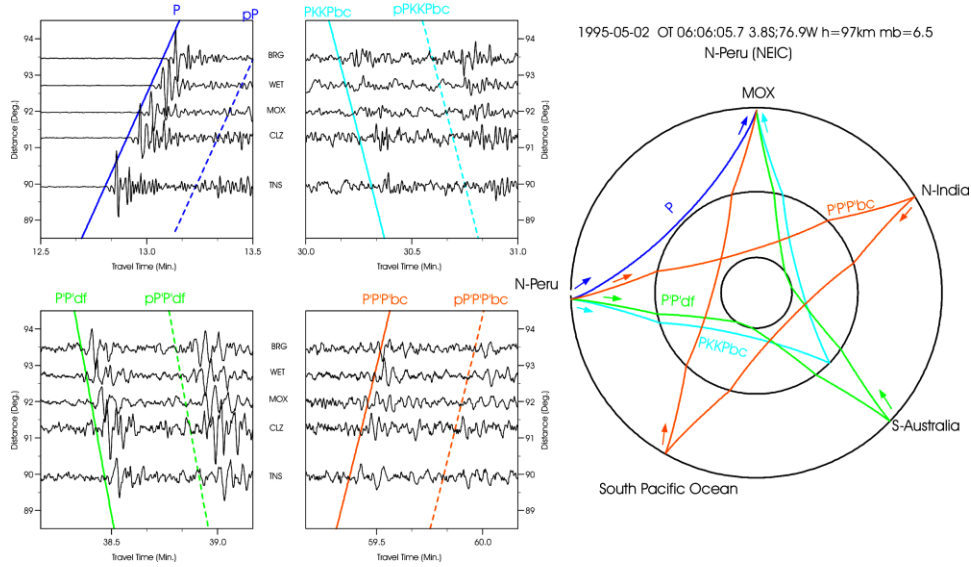
**Fig. 11.79 Right:** Source data of the Myanmar earthquake and ray paths to station CLZ in Germany. **Left:** Short-period Z-component record segments of GRSN stations with P, P' and P'3 arrivals at  $65^\circ < D < 70^\circ$  together with their theoretically expected arrival times according to the IASP91 and JB tables. Note the much better fit with the JB times.



**Fig. 11.80 Left:** 1 min long SP record segments showing P and PKPdf (outer left) together with P'2, P'3 and P'4 (middle left) at GRSN stations. **Right:** Related ray paths and source data. Note the negative slowness for P'2 and P'4. The travel-time curves relate to IASP91.

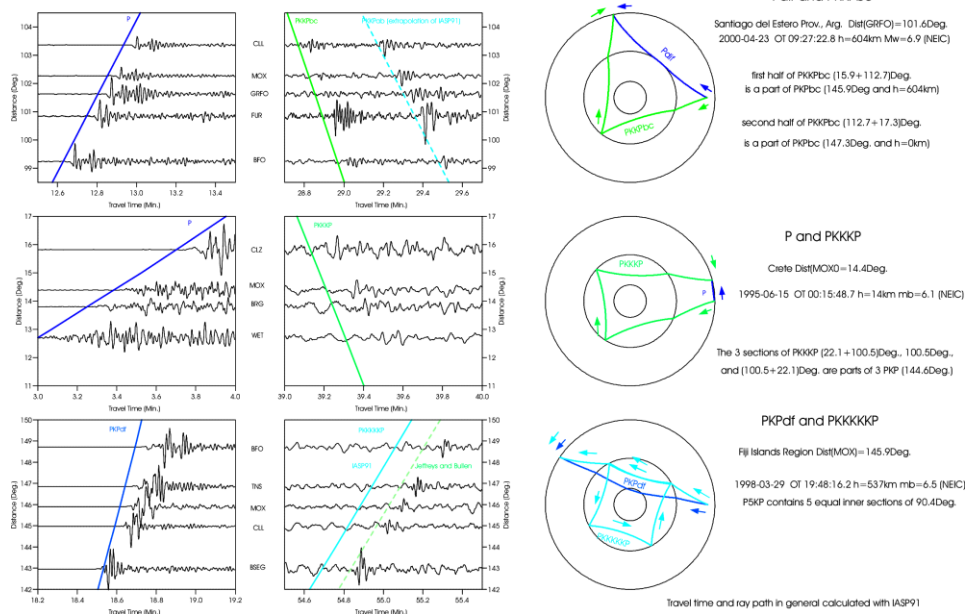


### Late and very late core phases



**Fig. 11.81** Late and very late multiple core phases PKKP, P'2 and P'3, respectively, together with their depth phases in short-period filtered record segments of GRSN stations from an earthquake in Northern Peru at epicentral distances around 92°. For an animation of ray propagation and seismogram formation from this source see file 6 in IS 11.3.

### PKKP, P3KP, and P5KP



**Fig. 11.82 Left:** Short-period 1 min record segments of P and PKP (outer left) together with those of PKKP, P3KP and P5KP (middle left) at GRSN stations. Note in the upper middle segment the excellent fit of the second onset with the extrapolation of the IASP91 PKKPbc branch and the much better fit of the P5KP onset in the lower middle segment with the JB-times **Right:** Related source data and ray paths. Note the negative slowness for PKKP and P3KP. The other theoretical travel-time curves relate to IASP91.

Note that the difference between the azimuth of the P wave and that of P'P' and PKKP, respectively, is  $180^\circ$  (Figs. 11.77 and 11.78). The related angular difference of the surface projections of their ray paths is  $360^\circ - D$  where  $D$  is the epicentral distance. Accordingly, the slowness of P'P' as well as of any even number P'N is negative, i.e. their travel time decreases with  $D$ . This also applies to PKKP and P3KP, as can be seen from Fig. 11.82. The surface projection of the travel paths of P'3 is  $360^\circ + D$  and that of P'4 is  $2 \times 360^\circ - D$ . PKPPKP (= P'P') is well observed between about  $40^\circ < D \leq 105^\circ$ . In this range the phase follows the onset of P by 33 to 24 min (Figure 10 in EX 11.3 with observed data). The existence of P'N is not limited to PKPbc. Fig. 11.79 shows an example of P'3df, recorded at a distance of about  $67^\circ$ . P'4 is sometimes observed in the distance range  $112^\circ$  to  $136^\circ$ . An example is given in Fig. 11.80.

Similar ray paths can be constructed for PNKP, the phase with  $(N-1)$  reflections from the inner side of the core-mantle boundary (Fig. 11.82). Figure 9 in EX 11.3 gives the differential travel-time curves for PKKP to the first arrivals P or PKP, respectively, in the distance range between  $80^\circ$  and  $130^\circ$  together with the observed data. In this range PKKP arrives 13 to 19 min behind P or 9.5 to 12 min behind PKP<sub>df</sub>. Higher multiple reflections from the inner side of the CMB such as P3KP, P4KP and P5KP are observed, if at all, at  $37 \pm 1$  min after the first arriving wave. The latter is true for P3KP following P at around  $10^\circ$ , for P4KP following P between  $45^\circ < D < 75^\circ$  and for P5KP following the onset of PKP<sub>df</sub> between about  $130^\circ < D < 150^\circ$  (called “37- min” rule- of- thumb).

The travel-time difference of multiple reflected core phases to the first arriving P and PKP does not change much with source depth. Therefore, you may try it just for fun although there is no more *need* for it in the present time world of near real-time global network locations and data dissemination, but the identification of these phases and knowing their travel-time curves, allows rather good distance estimates to be made even from single station records, irrespective of whether you know the source depth or not. Because of the inverse differential travel-time curves of PKPPKP and PKKP with respect to P and PKP their identification can be facilitated by comparing the onset times at neighboring stations (e.g., upper left two panels in Fig. 11.81).

The polarization of both the first arrival and the possible PKKP or PKPPKP onset, determined from 3-component records, can also aid identification because their azimuths should be opposite to that of P or PKIPK, respectively. Sometimes, also converted core reflections such as SKPPKP or SKKP can be observed in short-period recordings. However, direct or reflected core phases, which have traveled along the two ray segments through the mantle as S waves (such as SKS, SKKS, etc.) are mostly observed in broadband or long-period records.

### 11.5.5 Final remarks on the recording and analysis of teleseismic events

Box 2 summarizes the key criteria that should be taken into account when recording and analyzing seismograms from sources at teleseismic distances (see also 11.2.6.1).

## Box 2: General rules for recording and analyzing teleseismic events

- The overall duration of teleseismic records at epicentral distances larger than  $15^\circ$  (or  $20^\circ$ ) ranges from tens of minutes to several hours. It increases both with epicentral distance and the magnitude of the source.
- High frequencies, of S waves in particular, are more strongly attenuated with distance. Therefore recordings at teleseismic distances are usually of lower ( $f \approx 0.01 - 1$  Hz) frequency than local or regional recordings although for some events frequencies up to about 10 Hz have also been observed in P and PKP waveforms.
- Usually only longitudinal waves, both direct or multiple reflected P and PKP phases, which are much less attenuated than S waves, are well recorded by short-period, narrow-band seismographs (or their simulated equivalents) with high magnification of frequencies around 1 Hz. However, S waves from deep earthquakes may sometimes be found also in SP teleseismic records.
- Because of the specific polarization properties of teleseismic body and surface waves, polarization analysis is an important tool for identifying the different types of wave arrivals.
- According to the above, teleseismic events are best recorded by high-resolution 3-component broadband seismographs with large dynamic range and with sampling rates  $f = 20\text{-}40$  Hz.
- The main types of seismic phases from teleseismic sources are (depending on distance range) the longitudinal waves P, Pdif, PKP, PcP, ScP, PP, and PPP and the shear waves S, Sdif, SKS, ScS, PS, SS, and SSS. The longitudinal waves are best recorded on vertical and radial components whereas the shear waves appear best on transverse and/or radial components.
- Multiple reflected core phases such as P'N and PNKP, which appear on SP records as isolated wavelets, well separated from P or PKP, may easily be misinterpreted as P or PKP arrivals from independent seismic sources if no slowness data from arrays or networks are available. Their proper identification and careful analysis helps to avoid wrong source association, improves epicenter location and provides useful data for the investigation of the physical properties of the deeper interior of the Earth.
- Several body wave phases such as PP, PS, SP, SS, PKPab, SKKSac, SKKSdf, PKPPKpab, SKSSKSab and their depth phases undergo phase shifts and wavelet distortions at internal caustics (see 2.5.4.3). This reduces the accuracy of their time and amplitude picks and their suitability for improving source location by waveform matching with undistorted phases. Therefore it is recommended that seismological observatories correct these phase shifts prior to parameter readings by applying the inverse Hilbert-transformation, which is available in modern software for seismogram analysis.
- Surface waves of shallow events have by far the largest amplitudes but surface wave amplitudes from deep earthquakes and large (nuclear) explosions are small at teleseismic distances.
- At seismic stations or network centers the following parameter readings should be obligatory during routine analysis: onset time, SNR, and, if possible, polarity of the first arriving phase; maximum P amplitude A [nm],

period  $T$  [s] and/or ground velocity  $V$  [nm/s]; onset time of secondary phases; and, if existing, additionally also for surface waves the maximum amplitude  $A$  [nm], period  $T$  [s] and/or  $V$  [nm/s], irrespective of source depth.

- P-wave amplitudes  $A_{\max}$  for the determination of the short-period body-wave magnitude  $m_b$  have to be measured on standard short-period (WWSSN-SP simulated) records at periods  $0.2 \text{ s} < T < 3 \text{ s}$  and for broadband  $mB\_BB$   $V_{\max}$  in the period range between 0.2 s and 30 s, whereas the surface-wave amplitudes  $A_{\max}$  for the determination of the surface-wave magnitude  $M_s_{20}$  have to be measured on standard long-period filtered (WWSSN-LP or SRO-LP simulated) records, typically in the period range between 18 and 22 s, whereas  $V_{\max}$  for broadband  $M_s\_BB$  has should be measured between 3 s and 60 s period.
- For more guidance on magnitude determination, using also other phases and records/filters, consult Chapter 3 and IS 3.3 in particular.
- Networks and arrays should additionally measure and report slownesses and azimuths for analysed onsets.
- For improved determination of epicentre distance and source depth the measurement and reporting of onset time readings for secondary onsets such as S, depth phases (pP, sP, sS, etc.) and of core reflections (PcP, ScP, etc.) in particular, are very important.
- Picking and reporting of onset-times, amplitudes and periods of other significant phases, including those not identified, are encouraged by IASPEI within the technical and personnel facilities available at observatories and analysis centers as being a useful contribution to global research. These extended possibilities for parameter reporting are now well supported by the recently adopted IASPEI Seismic Format (ISF), which is much more flexible and comprehensive than the traditional Telegraphic Format (see 10.2 as well as IS 10.1 and 10.2), or by QuakeML (see IS 3.2, section 2).
- For reporting of seismic phases (including onsets not identified) one should exclusively use the new IASPEI phase names. For the definition of seismic phases and their ray paths see IS 2.1. Amplitude data should be reported according to the agreed IASPEI nomenclature for standard magnitudes (IS 3.3) and also other non-standard amplitude measurements (see IS 2.1).

## 11.6 Is fully automatic seismogram analysis, event location and classification already possible and feasible?

### 11.6.1 Introduction

Despite the potential benefits which digital data and interactive analysis software nowadays offer it is a matter of fact that digital records are increasingly less frequently read manually in a comprehensive way by seismologically trained experienced people. As outlined in Chapter 1, one main reason for this development is the tremendous increase in the number of high-gain seismic stations and networks deployed now all over the world for a steadily growing number of tasks which nowadays go even far beyond classical earthquake monitoring. They include volcano- seismology (Chapter 13 and Hammer et al., 2012); monitoring of induced seismicity (e.g., McGarr et al., 2002; Beyreuther et al., 2012) related to industrial activities such as the exploitation of geothermal, gas and oil reservoirs, mining activities (e.g., Kwiitek



et al., 2010), waste water disposal, impoundment and level-changes of huge water reservoirs (e.g., Gupta and Rastogi, 1976), or mass movements in mountain areas (e.g., Hammer et al., 2013). Accordingly, the amount of data to be analyzed has grown in recent decades by several orders of magnitude as compared to the times of classical analog earthquake seismology. Another reason is the increasing unwillingness of decision makers and funding institutions to pay fair salaries for a much larger number of well-educated people required to do all these jobs. Pressing demands for reducing labor cost both in the public sector and in industry and the often non-realistic belief that in the computer age automated procedures could do the job already not only faster but even better than well trained and experienced human beings are dominating nowadays.

The latter is surely not yet true. Nevertheless, we are now in a rapidly advancing process of increasingly better modeling of human intelligence, pattern recognition, and knowledge-based decision making. Machine-learning methods for the automatic classification and analysis of seismic waveforms gain momentum (e.g., Riggelsen and Ohrnheimer, 2012). Automatic algorithms, advanced or not yet advanced, have the advantage of being objective. Their results are fully reproducible, but not the analysis results of a group of human analysts (see IS 11.5). This even makes it difficult to correctly assess, in a unique manner, the performance of automatic algorithms as compared to human seismic data analysis. An example for such a comparison see is section 3.2.3.2 in Chapter 3. Moreover, both the acceptable ratio between missed events and false alarms on the one hand and between the timeliness and required accuracy/completeness of the data and derived information on the other hand are task-dependent. This also applies to the proper balance to be found between automatic and human analysis of seismological data.

Of greatest importance is in this context the fact of rapid population growth and urbanization since the mid 20<sup>th</sup> century, combined with a dramatic increase of the vulnerability and related human and material losses due to earthquakes and volcanic eruptions. This necessitates the development of earthquake, tsunami and volcanic eruption early warning systems (EEWS; TEWS; VEEWS) for enabling fastest possible appropriate automatic or human controlled response aimed at mitigating the disastrous consequences of such events. Highest priority has the timeliness of sufficiently accurate information for triggering and guiding response actions. This can only be realized with fully or highly automated systems of seismological data analysis and is clearly different from the need for most complete and precise/accurate information required for related future basic research.

Early efforts of automatic seismological detection and classification techniques were based on attributes of specific seismic phases. However, when monitoring volcano-induced seismicity, due to the rather different source processes and complex propagation conditions of the mostly high-frequency waves in a highly heterogeneous medium, a clear identification of individual seismic phases may not be possible. Therefore methods that make use of the whole seismogram are required for event type classification. The majority of algorithms applied to event-based data sets have been based so far on artificial neural networks (ANN). These algorithms, however, are “black boxes”, as are support vector machines that focus on the output only and do not allow any insight into the physical processes which control the relationship between input and output parameters. Alternative approaches will be sketched below. Good summaries with many references one finds, e.g., in Hammer et al. (2012).

A state-of-the-art overview of concepts, theory and approaches to automated seismic event and phase identification gives Chapter 16. Already operational or still experimental tsunami

or earthquake early warning systems such as the Pacific Tsunami Warning System (PTWC; <http://ptwc.weather.gov/>), GITEWS (<http://www.gitews.org/index.php?id=6>), and similar ones in Alaska, Japan, Mexico, Taiwan, Turkey and elsewhere (see Gasperini et al., 2007; Allen et al., 2009; Fleming et al., 2009) are examples of how far developments have already gone this way. In the following we add a few more case studies published in recent years that illustrate the different approaches chosen and the progress made in the automation of seismogram analysis and the classification of seismic events.

### 11.6.2 Amplitude/energy event triggers, phase identification and event location

Allen (1978, 1982) has been the first to propose an automatic P-wave picker for earthquake recognition and timing from short-period narrow-band single trace seismic records. The trigger is based on the long-term/short-term average amplitude, respectively energy ratio STA/LTA (Fig. 11.86; for details Chapter 16 and IS 8.1). It is still the most common and widely applied event trigger algorithm which has to be tuned to the specific signal and noise conditions.

Still most common is the application of the STA/LTA trigger in seismic network recordings. A seismic event is then declared if at least 4 stations have coincident observations in time of trace amplitudes above the set detection threshold that are consistent with seismic wave propagation models. Then, assuming that all these signal detections above noise relate to the first P-wave onset, they can be associated with each other to calculate the 4 unknowns of the hypocenter coordinates and origin time. Various such automatic association programs have been developed, e.g., by Engdahl and Gunst (1966), Slunga (1980), and a knowledge-based system by Bache et al. (1993) to produce automatically teleseismic event bulletins. However, as emphasized by Tong and Kennett (1996), most association programs use only P and S (including also Pn, Pg, Sn, Sg and Lg) as identifying phases. Other frequently observed phases such as PcP, ScS, PKP, PKiPK, PP, PS, SS, SKS, SKP, depth phases and multiple core phases (Fig. 11.16 and Figs. 2.57-2.61 in Chapter 2) are usually not taken into account. Withers et al. (1998a) compared selected trigger algorithms for onset time picking both in the time and frequency domain and also for particle motion and adaptive processing while Withers et al. (1998b) described an automated local and regional seismic event detection and location system using waveform correlation. All of these algorithms require a multi-station network, as do small aperture seismic arrays used for event detection and location of regional events (Chapter 9).

If, however, the number of available stations is small and/or the SNR at many sites is low then signals may be missed on the single station level and as a consequence probably a significant number of weak events will not appear in the automatic draft event bulletins. But then there will also be no visual inspection of records concerned and those weak events remain undetected. On the other hand, when lowering the detection threshold too much then noise fluctuations may cause a great number of spurious detections, *false alarms*, not related to true seismic events or even being wrongly associated to *fake events*, which all have to be reviewed in the interactive visual control step, thus increasing again the analyst workload.

An alternative approach for event detection and location is to rely not only on the automatically picked single component first onset time readings, i.e., just measuring one

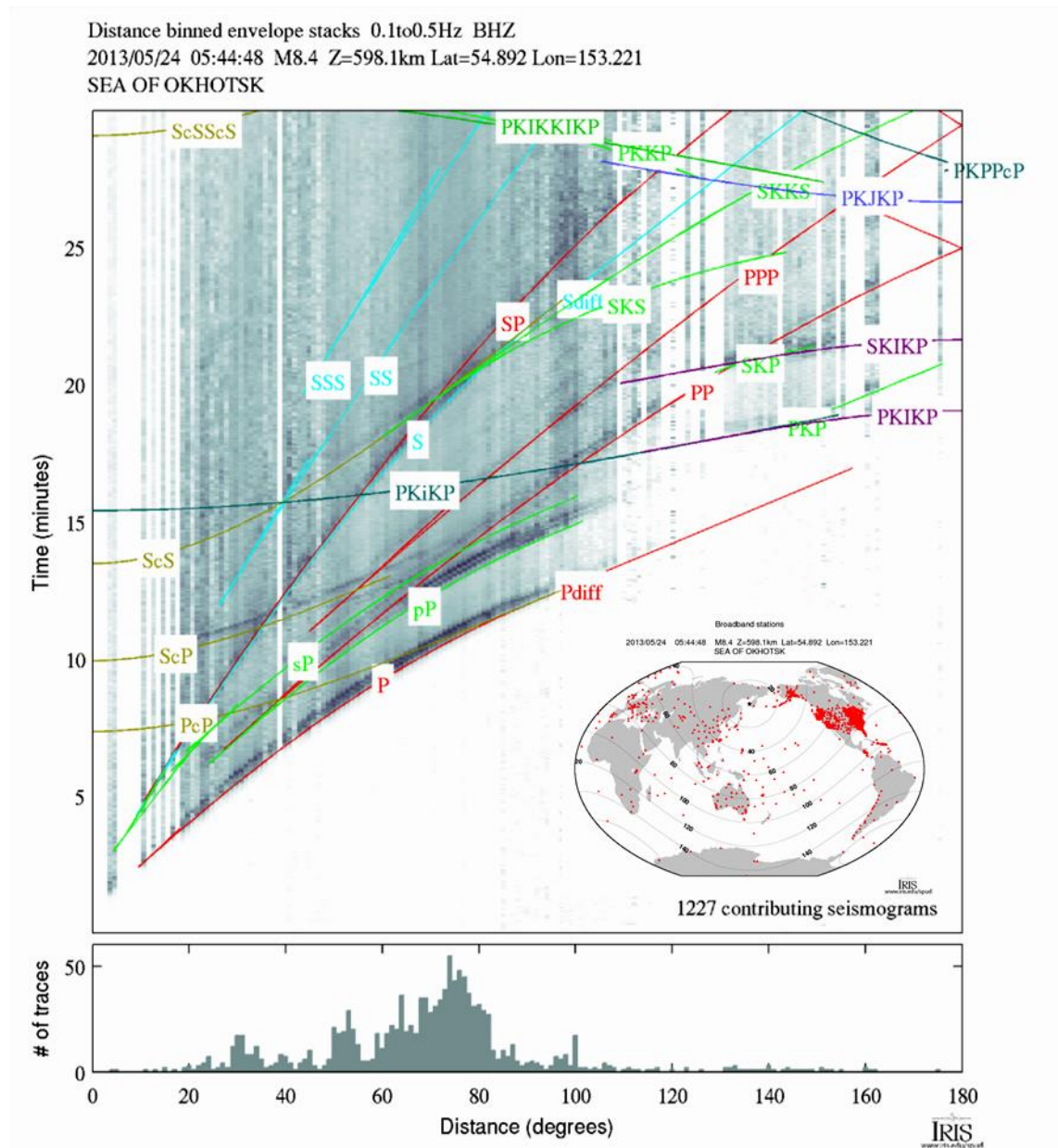
parameter at many stations. Rather one should improve also the accuracy and completeness of waveform attributes measured at 3-component broadband records of single stations for many seismic phases,. Such differences may relate to polarization, amplitude ratios, frequency differences and others. Such comprehensive automatic real-time single-station record analysis may yield rather good starting locations for a subsequent iterative location refinement. This particularly holds for short-period seismic arrays because of the inferior distance control of P-wave slowness data as compared to distance estimates based on differential travel times between P and later phases with rather different slowness. But it will also apply when network data are available only from stations with rather bad azimuthal and distance coverage. The latter is often the case for near coastal events because of lacking or insufficient stations in the ocean environment or for seismic events in areas with no co-operation and data exchange between neighboring countries at all. Because of the globally rapidly growing number of stations with free data excess and exchange via Internet this will, however, in future increasingly less be the case. On the other hand, there exist but no near real-time data exchange with others, such as. In such cases the larger information content of 3C BB stations has necessarily to be exploited. Also the International Monitoring System (IMS) of the CTBTO with its rather autonomous global network of a relatively limited number of high quality stations has shown great interest in the performance of 3C-station based automatic event detection, classification and location schemes (Riggelsen and Ohrnberger, 2012).

Saari (1981) described an algorithm for automatic event location by using single-site, 3C seismograms from near earthquakes ( $D < 100$  km) and Cichowicz (1993) an automatic S-wave picker. Earle and Shearer (1994) extended the STA/LTA trigger concept to 3C broadband velocity records and their bandpass-filtered versions in all distance ranges. Astiz et al. (1996) additionally used axes rotated records from the NEIC data base and the Global Seismic Network (GSN) in order to take into account also the different polarization properties of the various seismic phases that appear in different frequency bands and distances ranges. These authors took the STA/LTA along an envelope function generated from the seismogram, thus adapting the trigger parameters to the variable amplitude-frequency content of seismic noise and signal in the band-pass filtered records. The algorithm returns arrival times and related pick qualities which can then be associated in plots of travel-time versus distance easily with theoretically expected travel-time curves. Some of the travel-time-period-polarization-distance plots of these authors have been reproduced as Figs. 2.57, 2.58 and 2.61, together with the related IASP91 theoretical travel-time curve for short-period and long-period phases in Figs. 2.59 and 2.60.

With this automatic procedure it has been possible to pick practically all seismic phases which could be expected according to the 1D global travel-time model and which an experienced analyst would be able to recognize and identify with different relative frequency according to their different average SNR in seismic records of different type and in different components (see, e.g., Fig. 11.16). Picks made from LP records are less precise but good enough to reveal the discrepancy in SV and SH travel-times due to upper mantle anisotropy, being larger for SS than for S. Moreover, the travel-time residuals from the short-period automatic picks were comparable with those reported to the ISC from mostly human analysis. This indicates that human picking errors are not the main reason for the observed data scatter but rather travel-time inhomogeneities due to local site effects and velocity heterogeneities along the travel paths as compared to the 1D Earth model used in the computations.

Trabant et al., (2012) developed the Astiz et al. (1996) procedure further for real-time applications to all events with magnitude  $\geq 6.0$ . Their “Global Body-Wave Envelope Stacks”

are generated by stacking of envelope functions in  $1^\circ$ -distance bins. The envelopes have the mean pre-P-wave noise removed, and the square roots of the amplitudes are then summed. The stacks are nonlinearly weighted by the number of contributing seismograms to enhance visualization of the wave field. They are among the about 70 different automatically generated “Event Plots” published by the IRIS Data Management Center (DMC) about 2 hours after origin time, including, amongst others, also global surface- and body-wave record sections, virtual array P-waves and vespagrams and P-wave coda stacks (see <http://www.iris.edu/spud/eventplot>). An example of a body-wave envelope stack plot for the 2013 Mw(GCMT)8.3 deep Sea of Okhotsk earthquake is given in Fig. 11.83.

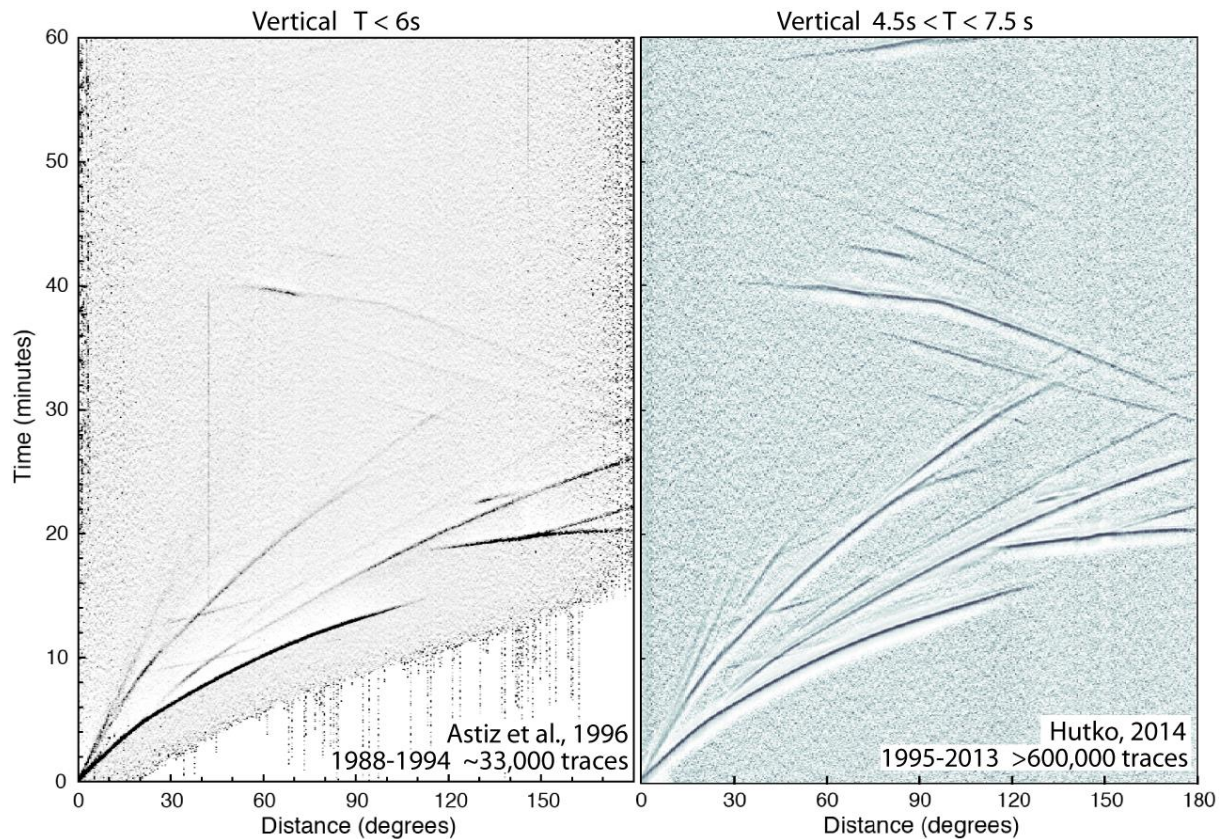


**Fig. 11.83** IRIS DMC global body-wave envelope stacks for the May 24, 2013, Mw(GCMT)8.3 Sea of Okhotsk deep earthquake ( $h = 598$  km) with inserted station distribution map and superposed theoretical AK 135 body-wave travel-time curves. Courtesy of A. Hutko, 2014.



For the same event we presented in Fig. 11.15 an unfiltered STS2 broadband record of the German station RUE at  $D = 67.9^\circ$ . However, for the envelop travel-time stack a 0.1-0.5 Hz bandpass filter had been applied, as compared to the unfiltered 0.0083-20 Hz STS2 record in Fig. 11.15. Therefore, in the DMC vertical component plot the more long-period and usually also more horizontal component phases SS, sSS, as well as the inversely dispersed mantle surface waves are not present. Also the usually distinct core phases PKP, SKP and SKS are underrepresented do to the lack of stations at  $D > 100^\circ$  for this specific event (see inserted map of station distribution). In any event, such near real-time plots allow a quick assessment of data availability and quality for almost all sufficiently large and thus hazardous earthquakes. Moreover, similar to vespagrams, these plots are especially useful for quickly scanning events to determine whether in the considered bandwidth range specific phases are visible in the data or not.

Fig. 11.84 compares long-term Astiz et al. (1996) STA/LTA stacks with the respective new global IRIS stacks based on about 20-times more data. The improvement is obvious.



**Fig. 11.84** Comparison of the long-term Astiz et al. (1996) vertical component STA/LTA stacks for  $T < 6$  s with the respective new global IRIS stacks in the period range  $4.5 \text{ s} < T < 7.5 \text{ s}$ , based on more than 600,000 event records. For more and updated information see <http://www.iris.edu/dms/products/globalstacks>. Compiled by courtesy of A. Hutko, 2014.

Chapter 16 gives many more examples on automatic event detection, phase picking and phase identification based on single component, 3-component-, network and array records. Amongst these examples the authors brief also on essentials of the rather sophisticated 3C algorithms developed by Tong and Kennett (1995, 1996) and Bai and Kennett (2000). These algorithms

combine automated detection and phase association for first and later arrivals and have been implemented in an experimental expert systems for automatic later phase identification and event location by means of single station 3-component broadband records. One key element was the introduction of a complexity measure for better discrimination between seismic signals and noise. It mimics the performance of the human eye, which uses the local change in frequency as a key criterion for a phase arrival. Also polarization analysis proved to be quite successful in distinguishing body waves from background noise. Assessing the performance of this robust and near real-time automatic event-recognition and classification system for teleseismic earthquakes Tong and Kennett (1996) conclude that "...the more phases detected and identified when recognizing a seismic event, the better discrimination between the possible solutions ...". In most cases their estimated event distances were within 7% and the related azimuths within 5°, thus constraining the location within 4° caps. This compares well with what also an experienced analyst could achieve off-line with some more time delay even with simple analog 3C data analysis (section 11.2.6.2 and EX 11.2), or even better and more comfortably nowadays interactively with modern PC-assisted analysis programs (see example 6 in IS 11.6). Surely, no single station location cannot compete at all with modern seismic network locations and are not thought to be propagated here as being a viable alternative for the latter. But they illustrate what would be achieved already with competent comprehensive data analysis in near real-time with just one good 3C-BB station in the case that network locations are not timely available, for what reason ever. And for "Seismology at Schools", undergraduate geoscience courses or public demonstrations on how seismology works (concept Education and Outreach) such on-the-spot single-station performance demonstrations are much more attractive and convincing than the reference to the near real-time communication about a seismic event by the agency processing the data of an anonymous regional or global network consisting of dozens to hundreds of stations.

In a follow-up paper Bai and Kennett (2000) proposed for arrival recognition and detection of a variety of P and S type phases a combination of three different techniques with different sensitivity to noise. They may be applied with different weightings to account for specific station site effects such as temporal/seasonal variations in the nature and power of seismic noise. The authors applied this method to unfiltered records in a wide range of epicentral distances from far-regional to teleseismic. Thus they followed the precept of Douglas (1997): "...the less filtering, the better the algorithm." The procedure works successful even when noise is quite low and weak phases are buried in the coda of earlier events. For the used set of windows and thresholds they achieved an effective precision of onset-time picking of about 1 s, thus constraining the differential time between P and S to within 2 s. This corresponds to about 0.3° in the distance range between 20° and 60°. Once a phase detection has been made even more sophisticated picking algorithms may be applied subsequent filtering can help to clarify the nature of phases. such as the tensor AR method by Leonard and Kennett (1999). S

While the above approach focusses on the comprehensive picking and interpretation of later phase onsets in 3C-records at single stations VanDecar and Crosson (1990) aimed at the determination of teleseismic relative phase arrival times using multi-channel regional network records, their cross-correlation and least squares. They found for quality records of the Washington Regional Seismograph network that the rms uncertainty in arrival time estimates was of the order of the sampling interval of 0.01 s. Also the reproducibility of delay anomalies was excellent for events from the same geographic location, even if their waveform character differed significantly.

Outlining further the details of these methods would go, however, well beyond the scope of the Manual. Here the original literature needs to be consulted. The same applies to the machine learning approaches by Ohrnberger and his team (e.g., Riggelsen et al., 2007; Riggelsen and Ohrnberger (2012); Hammer et al., 2012). Therefore, we will introduce them only briefly in the next and last section.

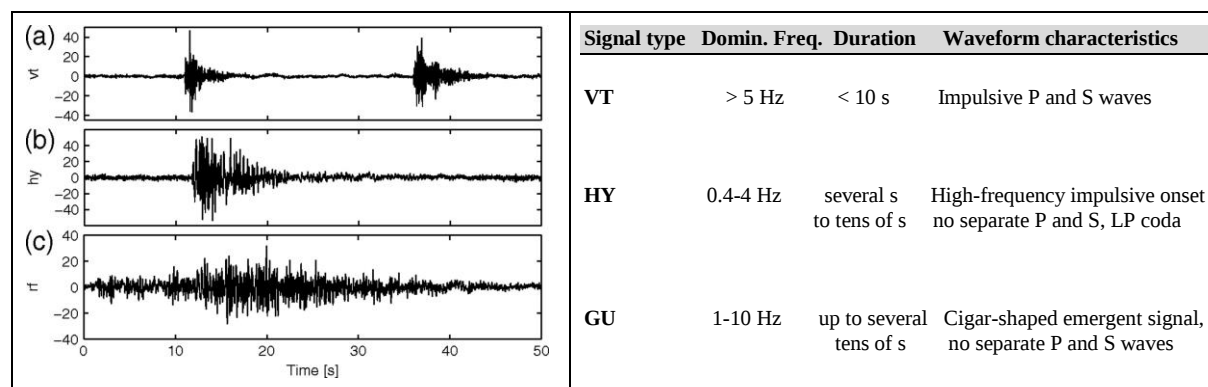
### **11.6.3 Machine learning approaches for improving event detection and classification at single 3-component seismic stations**

Ohrnberger (2001) was the first who applied Hidden Markov Model (HMM) based speech recognition techniques to parameters deduced from the continuous analysis of seismic data recorded at the Mt. Merapi volcano (Chapter 13; also: Hidden Markov Model Toolkit HMT, <http://htk.eng.cam.ac.uk/>; Python toolbox ObsPy, <http://www.python.org/> and <http://docs.obspy.org/index.html>). Riggelsen et al. (2007) introduced the method of Dynamic Bayesian networks (DBNs) to supervised learning in seismology. One of the key elements in their detection procedures is the adaptation to drifts in the noise characteristics. This results in distinct improvement of performance.

In order to cope with the situation of sparse networks, e.g., at volcanos and in particular also during volcanic crises, Beyreuther et al. (2012) designed an earthquake detector with a probabilistic architecture. It is based on the generation of characteristic functions and the derivation of a multi-dimensional feature vector via principal component analysis and inverse weighting. It performs already well when applied to just a single station. In contrast to the analysis of teleseismic earthquake records, where the emphasis lies on the detection and identification of as many as possible phase onsets for a given event, the main task in volcano seismology is to detect and discriminate/classify different types of volcano-seismic events of different origin such as volcano-tectonic event (VT), Hybrids (HY) with characteristics of both “high-frequency” and “long-period” earthquakes, or rock falls (GU). In periods of reactivation or ongoing eruption of a volcano they may occur in dense and mixed sequences in different relative proportions. The relative frequency and strength of occurrence of these different types of events have both predictive power and assessment quality for the formulation of a likely pending volcanic threat or the objective characterization and documentation of an ongoing activity. Applying the procedure to almost 4 months of continuous data sampling during the activity period of Mt. Merapi, Indonesia, in 1998, the authors concluded that even single channel single station multi-parameter HMM triggers work well, provided that they have been well-trained with a sufficient number of representative training events for the various classes. In terms of objectivity and reproducibility of the classification results the procedure may be equivalent or even outperform those of an experienced volcanologist who bases his decisions on the waveforms recorded at many more sensors.

However, often training set data are not available and their compilation and testing is a tedious process which takes quite some time. Therefore, Hammer et al. (2012) took a HMM *learning-while-recording* approach for volcano monitoring. In a first step the algorithm converts the seismic waveforms into a sequence of n-dimensional parameter vectors. This compressed signal representation increases the robustness of the procedure. After creation of the draft model for each particular event type, the classification process can be started on an unknown data stream and begin to retrain the original classification model as more and more

data become available. Moreover, new event type classes can be created if incoming data segments cannot be associated with any of the previously specified classes. The procedure has been applied to the detection and classification of volcano-tectonic events, hybrids and rockfalls in records at Soufrière Hills volcano, Montserrat (Fig. 11.85). The initial recognition accuracy of 84% could be increased to 97% by retraining the system on newly detected events.



**Fig. 11.85** Left: Typical examples of VT, HY and GU events recorded at Soufrière Hills volcano, Montserrat. Right: Characteristics of these three considered signal classes. Compiled from Figure 6 and Table 2 in Hammer et al. (2012), BSSA, **102**(3), pp. 954 and 955; © Seismological Society of America.

The same algorithm has been applied by Hammer et al. (2013) to small earthquakes with magnitudes  $M_l < 2.5$  and source depth  $\leq 20$  km and thus record durations less than 20 s in an area of 2700 km<sup>2</sup> in Switzerland. They had to be detected and discriminated from quite many quarry blasts and rare rockfalls in records of the only seismic station, FUSIO, in this area. Over the 9 years of the Swiss Earthquake Survey (SED) event catalog analyzed for the small area under investigation with the continuously re-trained self-learning algorithm, 97% of all events (earthquakes, quarry blasts and rockfalls) have been detected and none of the earthquakes or quarry blasts been misclassified as rockfall. But there is disagreement between the automatic and human classification for about 11 % of the detected events whether these are earthquakes or explosions.

These examples give an idea of the development of theoretical concepts and procedures for automatic detection and classification of seismic events. They show, as of now, what can be achieved with them and what not. But the various discussions in this section and in Chapter 11 as a whole also reveal what still remains to be the permanent duty of well-educated and practically trained seismologists who work with compassion and zeal over many years in the field of comprehensive seismogram and event analysis. Without their expertise also the automation of seismological data analysis procedures will not progress in the right direction, may result in too many uncontrolled and by users not well understood “black box”-procedures and data products whose completeness and accuracy may not satisfy the requirements for long-term fundamental and applied seismological research.



## Acknowledgments

The authors are very grateful to A. Douglas and R. D. Adams for their careful reviews which helped to significantly improve the original draft of the first edition. Thanks go also to staff members of the Geophysical Survey of the Russian Academy of Science in Obninsk who shared in reviewing the various sections of this Chapter (Ye. A. Babkova, L. S. Čepkunas, I. P. Gabsatarova, M. B. Kolomiyez, S. G. Poygina and V. D. Theophylaktov). Many of their valuable suggestions and references to Russian experience in seismogram analysis were taken into account. Thanks go to M. Ohrnberger and his team for making their recent publications available. We are grateful also to K. Stammler, for useful information and clarifications with respect to the procedures of the SZO Hannover and the analysis program SHM as well as to B. Weber and J. Saul for the same with respect to SeisComp3. Our greatest thanks, however, go to the four reviewers of the largely new and amended Chapter 11, L. Ottemöller, D. DiGiacomo, J. Schweitzer and W. Strauch. Their comments helped greatly to clarify several issues and to streamline and shorten the first draft. W. Strauch additionally contributed the subsection on Earthworm and A. Hutko from IRIS the Figs. 11.83 and 11.84 together with related references and links.

## Recommended overview readings

Bowers and Selby (2002)  
Dahlmann et al. (2009)  
Gasperini et al. (2007)  
Havskov and Ottemöller (2010)  
Kennett (2002)  
Kulhanek (1990 and 2002)  
Payo (1986)  
Richards and Wu (2011)  
Richter (1958)  
Richards (2002)  
Richards and Wu (2011)  
Scherbaum (2001 and 2002)  
Simon (1981)  
Willmore (1979)

## References

- Allen, R. V. (1978). Automatic earthquake recognition and timing from single traces. *Bull. Seism. Soc. Am.*, **68**, 1521-1532.
- Allen, R. V. (1982). Automatic phase pickers: Their present and future prospects. *Bull. Seism. Soc. Am.*, **72**, 225-242.
- Allen, Q. M., Gasperini, P., Kamigaichi, O., and Böse, M. (2009). The status of earthquake early warning around the world. An introductory overview. *Seism. Res. Lett.*, **80**, 682-693; doi: 10.1785/gssrl.80.5.682.
- Ambraseys, N. N., and Adams, R. D. (1986). Seismicity of West Africa. *Annales Geophysicae*, **86**, 6, 679-702.

- Astiz, L., Earle, P., and Shearer, P. (1996). Global stacking of broadband seismograms. *Seism. Res. Lett.*, **67**, 8-18.
- Barsch, R. (2009). *Web-based technology for storage and processing of multi-component data in seismology*. PhD thesis, LudwigMaximilians Universität München, Munich, Germany.
- Baumbach(†), M. (1999). SEIS89 - A PC tool for seismogram analysis. In: Bormann (Ed.) International Training Course 1999 on Seismology, Seismic Hazard Assessment and Risk Mitigation, Lecture and exercise notes, Vol. 1, *GeoForschungsZentrum Potsdam, Scientific Technical Reports STR99/13*, 423-452.
- Bache, T., Bratt, W. R., Swanger, H. J., Beall, G. W., and Dashiell, F. K. (1993). Knowledge-based interpretation of seismic data in the Intelligent Monitoring System. *Bull. Seism. Soc. Am.*, **83**, 1521-1532.
- Bai, C. Y., and Kennett, B. L. N. (2000). Automatic phase-detection and identification by full use of a single three-component broadband seismogram. *Bull. Seism. Soc. Am.*, **90**, 187-198.
- Beyreuther, M., Barsch, R., Krischer, L., Megies, T., Behr, Y. and Wassermann, J. (2010). ObsPy: A Python toolbox for seismology. *Seism. Res. Lett.*, **81**(3); doi:10.1785/gssrl.81.3. 530.
- Beyreuther, M., Hammer, C., Wassermann, J., Ohrnberger, M., and Megies, T. (2012). Constructing a Hidden Markov Model based earthquake detector: application to induced seismicity. *Geophys. J. Int.*, **189**(1), 602-610; doi: 10.1111/j.1365-246X.2012.05361.x.
- Bisztricsany, E. (1958). *A new method for the determination of the magnitude of earthquake*. *Geofiz. Kozl.*, **7**, 69-96 (in Hungarian with English abstract).
- Bock, G., Grünthal, G., and Wylegalla, K. (1996). The 1985/86 Western Bohemia earthquakes: Modelling source parameters with synthetic seismograms. *Tectonophysics*, **261**, 139-146.
- Bokelmann, G. (1995). Azimuth and slowness deviations from the GERESS regional array. *Bull. Seism. Soc. Am.*, **85**, 5, 1456-1463.
- Bondár, I. and D. Storchak. (2011). Improved location procedures at the International Seismological Centre. *Geophys. J. Int.*, **186**, 1220-1244, doi:10.1111/j.1365-246X.2011.05107.x
- Bormann, P. (1971a). Statistische Untersuchungen zur Ortung teleseismischer Ereignisse aus Raumwellenregistrierungen der Station Moxa. *Veröff. Zentralinstitut für Physik der Erde* **9**(1971), 103 pp.
- Bormann, P. (1971b). Location of teleseismic events by means of body-wave records at station Moxa (in German: Ortung teleseismischer Ereignisse aus Raumwellenregistrierungen der Station Moxa), *Monatsberichte der ADW*, **13**, 10-12, 847-852.
- Bormann, P. (1972a). A study of travel-time residuals with respect to the location of teleseismic events from body-wave records at Moxa station. *Gerl. Beitr. Geophys.*, **81**, 117-124.
- Bormann, P. (1972b). A study of relative frequencies of body-wave onsets in seismic registrations of the station Moxa. In: Seismological Bulletin 1967, Station Moxa (MOX). *Akademie-Verlag*, Berlin 1972, 379-397
- Bormann, P. (1972c). Identification of teleseismic events in the records of Moxa station. *Gerl. Beitr. Geophys.*, **81**, ½, 105-116.

- Bormann, P., Wylegalla, K., Strauch, W. and Baumbach, M. (1992). Potsdam seismological station network: processing facilities, noise conditions, detection threshold and localization accuracy. *Physics Earth Planet. Int.*, **69**, 311-321.
- Bowers, D., and Walter, W. R. (2002). Discriminating between large mine collapses and explosions using teleseismic P Waves. *Pure and Applied Geophysics*, **159**, 803–830.
- Bowers, D., and Selby, N. D. (2009). Forensic Seismology and the Comprehensive Nuclear-Test-Ban Treaty. *Annual Reviews of Earth and Planetary Sciences*, **37**, 209–236.
- Buchbinder, G. R., and Haddon, R. A. W. (1990). Azimuthal anomalies of short-period P-wave arrivals from Nahanni aftershocks, Northwest territories, Canada, and effects of surface topography. *Bull. Seism. Soc. Am.*, **80**, 5, 1272-1283.
- Cassidy, F., Christoffersson, A., Husebye, E. S., and Ruud, B. O. (1990). Robust and reliable techniques for epicentre location using time and slowness observations. *Bull. Seism. Soc. Am.*, **80**, 1, 140-149.
- Chernobay, I. P., and Gabsatarova, I. P. (1999). Source classification in the Northern Caucasus. *Phys. Earth Planet. Int.*, **113**, 183-201.
- Chernobay, I. P., and Gabsatarova, I. P. (1999). Source classification in the Northern Caucasus. *Phys. Earth Planet. Int.*, **113**, 183-201.
- Chernobay, I. P., and Gabsatarova, I. P. (1999). Source classification in the Northern Caucasus. *Phys. Earth Planet. Int.*, **113**, 183-201.
- Christoffersson, A., Husebye, E. S., and Ingate, S. F. (1988). Wavefield decomposition using ML-probabilities in modeling single-site 3-component records. *Geophys. Journal*, **93**, 197-213.
- Cichowicz, A. (1993). An automatic S-wave picker. *Bull. Seism. Soc. Am.*, **83**, 180-189.
- Dahlman, O., Mykkeltveit, S., and Haak, H. (2009). *Nuclear Test Ban: Converting Political Visions to Reality*. Berlin: Springer.
- Davies, D., Kelly, E. J., and Filson, J. R. (1971). The VESPA process for the analysis of seismic signals. *Nature*, **232**, 8-13.
- Douglas, A., and Rivers, D. W. (1988). An explosion that looks like an earthquake. *Bull. Seism. Soc. Am.*, **78**, 2, 1011-1019.
- Douglas, A. (1997). Bandpass filtering to reduce noise on seismograms: is there a better way? *Bull. Seism. Soc. Am.*, **87**, 770-777.
- Douglas, A., Hudson, J. A., Marshall, P. D., and Young, J. B. (1974). Earthquakes that look like explosions. *Geophys. J. R. astr. Soc.*, **36**, 227-233.
- Douglas, A., Stewart, R. C., and Richardson, L. (1984). Comments on “Analysis of broadband seismograms from the Chile-Peru area” by R. Kind, and D. Seidl. *Bull. Seism. Soc. Am.*, **74**, 2, 773-777.
- Douglas, A., and Rivers, D. W. (1988). An explosion that looks like an earthquake. *Bull. Seism. Soc. Am.*, **78**, 2, 1011-1019.
- Douglas, A., Bowers, D., and Young, J. B. (1997). On the onset of P seismograms. *Geophys. J. Int.*, **129**, 681-690.
- Earle, P. S., and Shearer, P. M. (1994). Characterization of global seismograms using an automatic-picking algorithm. *Bull. Seism. Soc. Am.*, **84**, 366-376.
- Engdahl, E. R., and Gunst, R. H. (1966). Use of a high speed computer for the preliminary determination of earthquake hypocenters. *Bull. Seism. Soc. Am.*, **56**, 325-336.

- Engdahl, E. R., Van der Hilst, R. D., and Buland, R. P. (1998). Global teleseismic earthquake relocation with improved travel times and procedures for depth determination. *Bull. Seism. Soc. Am.*, **88**, 722-743.
- Fleming, K., Picozzi, M., Milkereit, C., Kühnlenz, F., Lichtblau, B., Fischer, J., Zulfikar, C., Özel, O., and the SAFER and EDIM working groups (2009). The self-organizing seismic early warning information network (SOSEWIN). *Seismol. Res. Lett.*, **80**(5), 755-771; doi: 10.1785/gssrl.80.5.755,
- Flinn, E. (1965). *Signal analysis using rectilinearity and direction of particle motion*. Proc. IEEE, **53**, 1874-1876.
- Gasperini, P., Manfredi, G., and Zschau, J. (2007). *Earthquake early warning systems*. Springer Berlin-Heidelberg, 349 pp.
- Geiger, L. (1910). Herdbestimmung bei Erdbeben aus den Ankunftszeiten. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*, Mathematisch-Physikalische Klasse, 331-349. (In 1912 translated to English by Peebles, F. W. L., and Corey, A. H.: see Geiger, L. (1912).
- Geiger, L. (1912). Probability method for the determination of earthquake epicenters from the arrival time only (translation of the 1911 article). *Bull. St. Louis Univ.*, **8**, 60-71.
- Goldstein, P., and Snoke, A. (2005). SAC Availability for the IRIS Community. Incorporated Institutions for Seismology Data Management Center Electronic Newsletter. <http://www.iris.edu/dms/newsletter/vol7/no1/sac-availability-for-the-iris-community/>
- Goldstein, P., Dodge, D., Firpo, M., Lee, M. (2003). 85.5 SAC2000: Signal processing and analysis tools for seismologists and engineers. In: W.H. K Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger (Eds.). *International Handbook of Earthquake and Engineering Seismology*, Vol. 81, Part B. 1613-1614: Academic Press, London. DOI: 10.1016/S0074-6142(03)80284-X
- Gupta, H. K., and Rastogi, B. K. (1976). *Dams and earthquakes*. Elsevier.
- Gutenberg, B., and Richter, C. F. (1956a). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Hammer, C., Beyreuther, M., and Ohrnberger, M. (2012). A seismic-event spotting system for volcano fast-response systems. *Bull. Seism. Soc. Am.*, **102**(3), 948-960; doi: 10.1785/0120110167..
- Hammer, C., Ohrnberger, M., and Fäh, D. (2013). Classifying seismic waveforms from scratch: a case study in the alpine environment. *Geophys. J. Int.*, **192**(1), 425-439; doi: 10.1093/gji/ggs036.
- Hanka, W., and R. Kind (1994). The GEOFON program. *Annali di Geofisica*, **37** (5), 1,060-1,065.
- Hanka, W., Heinloo, A., and Jäckel, K.-H. (2000). Networked seismographs: GEOFON real-time data distribution. *ORFEUS Electronic Newsletter*, **2**, 3, <http://www.orfeus-eu.org>.
- Hanka, W., Saul, J., Weber, B., Becker, J., Harjadi, P., Fauzi, and GITEWS Seismology Group (2010). Real-time earthquake monitoring for tsunami warning in the Indian Ocean and beyond. *Nat. Hazards Earth Syst. Sci.*, **10**, 2611-2622; doi: 10.5194/nhess-10-2611-2010.
- Harjes, H.-P., Jost, M. L., and Schweitzer, J. (1994). Preliminary Calibration of Candidate Alpha Stations in the GSETT-3 Network. *Annali di Geofisica*, **37**, 382-396.

- Havskov, J. (Ed.) (1996). The SEISAN earthquake analysis software for the IBM PC and SUN, Version 5.2. *Institute of Solid Earth Physics, Univ. of Bergen*, August 1996.
- Havskov, J., and Ottemöller, L. (1999). Electronic Seismologist - SeisAn Earthquake Analysis Software. *Seism. Res. Lett.*, **70**, 532-534.
- Havskov J., and Ottemoller, L. (2010). Routine Data Processing in Earthquake Seismology: With Sample Data, Exercises and Software, Springer (May 25, 2010), ISBN-10: 904818696X, ISBN-13: 978-9048186969
- Herrin, E. (1968). Introduction to “1968 Seismological Tables for P phases”. *Bull. Seism. Soc. Am.*, **58**(44), 1193-1195.
- Houard, S., Plantet, J. L., Massot, J. P., and Nataf, H. C. (1993). Amplitudes of core waves near the PKP caustic, from nuclear explosions in the South Pacific recorded at the “Laboratoire de Detection et Geophysique” network in France. *Bull. Seism. Soc. Am.*, **83**, 6, 1835-1854.
- Hwang, L. J., and Clayton, R. W. (1991). A station catalog of ISC arrivals: Seismic station histories and station residuals. *U.S. Geological Service Open-File Report* 91-295.
- IASPEI (2013). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data.  
[http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)
- IDC Documentation (May 1998). Technical Instructions: Analyst instructions for seismic, hydroacoustic, and infrozonic data, 103pp.
- Jeffreys, H., and Bullen, K. E. (1940, 1948, 1958, 1967, and 1970). Seismological Tables. British Association for the Advancement of Science, *Gray Milne Trust*, London, 50 pp.
- Johnson, C. E., Bittenbinder, A., Bogaert, B., Dietz, L., and Kohler, W. (1995). Earthworm: A flexible approach to seismic network processing, *IRIS Newsletter* **14**, 2, 1–4.
- Kennett, B. L. N. (Ed.) (1991). IASPEI 1991 Seismological Tables. Research School of Earth Sciences, Australian National University, 167 pp. Kennett, B. L. N. (2002). The seismic wavefield. Volume II: Interpretation of seismograms on regional and global scales. *Cambridge University Press*, Cambridge, x + 534 pp, ISBN 0-521-00665-1, 163-170.
- Kennett, B. L. N., and Engdahl, E. R. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, **105**, 429-465.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, **122**, 108-124.
- Kim, W.-Y., Simpson, D. W., and Richards, P. G. (1993). Discrimination of earthquakes and explosions in the Eastern United States using regional high-frequency data. *Geophysical Research Letters*, **20**, 1507–1510.
- Kulhánek, O. (1990). *Anatomy of seismograms*. Developments in Solid Earth Geophysics **18**, Elsevier, Amsterdam, 78 pp.
- Kulhanek, O. (2002). The structure and interpretation of seismograms. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). International Handbook of Earthquake and Engineering Seismology, Part A. *Academic Press, Amsterdam*, 333-348.
- Kwiatek, G., Plenkers, K., Nakatani, M., Yabe, Y., Dresen, G., and JAGUARS-Group (2010). Frequency-magnitude characteristics down to magnitude -4.4 for induced seismicity recorded at Mponeng gold mine, South Africa. *Bull. Seism. Soc. Am.*, **100**(3), 1165-1173 ; doi: 10.1785/0120090277.

- Lee, W. H. K. (1995). Realtime seismic data acquisition and processing. *IASPEI Software Vol.1*, 2<sup>nd</sup> Edition.
- Leonard, M., and Kennett, B. L. N. (1999). Multi-component autoregressive techniques for the analysis of seismograms. *Phys. Earth Planet Int.*, **113**, 247-263.
- Lockett, R., Ottemöller, L., and Whitmore, P. (2008). A Tsunami Warning System for the Northwest Atlantic. *ORFEUS Newsletter*, **8**, 1; <http://www.orfeus-eu.org>.
- McGarr, A., Simpson, D., and Seeber, L. (2002). Case histories of induced and triggered seismicity. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 647-661.
- Megies, T., Beyreuther, M., Barsch, R., Krischer, L., Wassermann, J. (2011). ObsPy – What can it do for data centers and observatories? *Ann. Geophys.*, **54**, 1; doi: 10.4401/ag-4838.
- Mykkeltveit, S., and Bungum, H. (1984). Processing of regional seismic events using data from small aperture arrays. *Bull. Seism. Soc. Am.*, **74**, 2313-2333.
- Ohrnberger, M. (2001). Continuous automatic classification of seismic signals of volcanic origin at Mt. Merapi, Java, Indonesia, *Dissertation, Institute of Geosciences, University of Potsdam*.
- Okal E.A. and Taladier, J. (1989) Mm: A Variable-Period Mantle Magnitude, *Journal of Geophysical Research*, Vol. 94, No B4, p.4169-4193, April 10, 1989
- Olivieri, M., and Clinton, J. (2012). An almost fair comparison between earthworm and SeisComp3. *Seismol. Res. Lett.*, **83** (4), 720-727
- Ottemöller, L., Voss, P., and Havskov, J. (Eds.) (2013). *SEISAN earthquake analysis software for Windows, Solaris, Linux and MacOSx Version 10.0*. Institute of Solid Earth Physics, Univ. of Bergen, July 2013, <http://seis.geus.net/software/seisan/seisan.html>.
- Payo, G. (1986). Introduccion al analisis de sismogramas. Instituto Geográfico Nacional, Madrid, 125 pp.
- Plenkers, K., Kwiatek, G., Nakatani, M., Dresen, D., and the JAGUARS group (2010). Observation of seismic events with frequencies  $f > 25$  Hz at Mponeng deep gold mine, South Africa. *Seism. Res. Lett.*, **81**, 467-479.; doi: 10.1785/gssrl.81.3.467.
- Pujol, J. (2004). Earthquake location tutorial: graphical approach and approximate epicentral location techniques. *Seism. Res. Lett.*, **75**, 63-74.
- Reymond, D. (2002). Seismic Tool Kit STK. <http://seismic-toolkit.sourceforge.net/>
- Richards, P. G. (2002). Seismological methods of monitoring compliance with the Comprehensive Nuclear Test-Ban Treaty. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 369-382.
- Richards, P., and Wu, Z. (2011). Seismic monitoring of nuclear explosions. In: Gupta, H. (Ed.) (2011): *Encyclopedia of Solid Earth Geophysics*. Springer, Dordrecht, Vol. 1144-1156.
- Richter, C. F. (1958). *Elementary seismology*. W. H. Freeman and Company, San Francisco and London, viii + 768 pp.
- Riggelsen, C., Ohrnberger, M., and Scherbaum, F. (2007). Dynamic Bayesian networks for real-time classification of seismic signals. In Kok, J., Koronacki, J., Lopez de Mantaras, R., Matwin, S., Mladenic, D., and Skowron, A. (Editors). *Knowledge Discovery in*

- Databases: PKDD 2007, Lecture Notes in Computer Science, Springer, Berlin/Heidelberg, **4702**, 565–572.
- Riggelsen, C., and Ohrnberger, M. (2012). A machine learning approach for improving the detection capabilities at 3C seismic stations. *Pure App. Geophys.*, 17pp.; doi: 10.1007/s00024-012-0592-3.
- Roberts, R. G., and Christoffersson, A. (1990). Decomposition of complex single-station three-component seismograms. *Geophys. J. Int.*, **103**, 55-74.
- Ruud, B. O., Husebye, E. S., Ingate, S. F., and Christoffersson, A. (1988). Event location at any distance using seismic data from a single, three-component station. *Bull. Seism. Soc. Am.*, **78**, 308-325.
- Saari, J. (1991). Automated phase picker and source location algorithm for local distances using a single three-component seismic station. *Tectonophysics*, **189**, 307-315.
- Scherbaum, F. (1996). *Of poles and zeros; Fundamentals of Digital Seismometry*, Kluwer Academic Publishers, Boston, 256 pp.
- Scherbaum, F. (2001). *Of poles and zeros: Fundamentals of digital seismology. Modern Approaches in Geophysics*, Kluwer Academic Publishers, 2<sup>nd</sup> edition, 265 pp. (including an CD-ROM by Schmidtke, E., and Scherbaum, F. with examples written in Java).
- Scherbaum, F. (2007). *Of poles and zeros; Fundamentals of Digital Seismometry*. 3<sup>rd</sup> revised edition, *Springer Verlag*, Berlin und Heidelberg.
- Scherbaum, F., and Johnson, J. (1992). Programmable Interactive Toolbox for Seismological Analysis (PITSA). IASPEI Software Library Volume 5, Seismological Society of America, El Cerrito.
- Scherbaum, F., and Bouin (2007). FIR filter effects and nucleation phases. *Geophys. J. Int.*, **130**(3), 661-668; doi: 10.1111/j.1365-246X.1997.tb01860.x
- Schweitzer, J. (2002). Slowness corrections – one way to improve IDC products. *Pure & Appl. Geophys.*, **158**, 375-396.
- Schweitzer, J. (2002). Simultaneous inversion of steep-angle observations of PcP and ScP in Europe – what can we learn about the core-mantle boundary? *J. Geophys. Res.*, **151**, 209-220.
- Seidl, D. (1980). The simulation problem for broad-band seismograms. *J. Geophys.*, **48**, 84-93.
- Seidl, D., and Berckhemer, H. (1982). Determination of source moment and radiated seismic energy from broadband recordings. *Phys. Earth Planet. Inter.*, **30**, 209-213.
- Seidl, D., and Stammer, W. (1984). Restoration of broad-band seismograms (part I). *J. Geophys.*, **54**, 114-122.
- Seidl, D., and Hellweg, M. (1988). Restoration of broad-band seismograms (part II): Signal moment determination. *J. Geophysics*, **62**, 158-162.
- Simon, R. B. (1981). *Earthquake interpretations: A manual of reading seismograms*. William Kaufmann Inc., Los Alto, CA, 150 pp.
- Slunga, R. (1980). International Seismological Data Center. An algorithm for associating reported arrivals to a global seismic network into groups of defining seismic events. FOA Report, C20386-T1.
- Takanami, T., and Kitagawa, G. (1988). A new efficient procedure for estimation of onset times of seismic waves. *J. Phys. Earth*, **36**, 267-290.

- Takanami, T., and Kitagawa, G. (1991). Estimation of the arrival times of seismic waves by multivariate time series model. *Ann. Inst. Stat. Math.*, **43**, 407--433.
- Takanami, T., and Kitagawa, G. (1993). Multivariate time-series models to estimate the arrival times of S waves. *Computers and Geosciences*, **36**, 295-301.
- Tong, C. (1995). Characterization of seismic phases – an automatic analyser for seismograms. *Geophys. J. Int.*, **123**, 937/947.
- Tong, C., and Kennett, B. L. N. (1995). Towards the identification of later seismic phases. *Geophys. J. Int.*, **123**, 948-958.
- Tong, C., and Kennett, B. L. N. (1996). Automatic seismic event recognition and later phase identification for broadband seismograms. *Bull. Seism. Soc. Am.*, **86**, 6, 1896-109.
- Trabant, C, Hutko, A. R., Bahavar, M., Karstens, R., Ahern, T. and Aster, R. (2012). Data products at the IRIS DMC: stepping-stones for research and other applications. *Seismol. Res. Lett.*, **83**(6), 846-854. doi: [10.1785/0220120032](https://doi.org/10.1785/0220120032).
- Vidale, J. D. (1986). Complex polarization analysis of particle motion. *Bull. Seism. Soc. Am.*, **76**, 1393-1405.
- Wagner, G., and Owens, T. (1996). Signal detection using multi-channel seismic data. *Bull. Seism. Soc. Am.*, **86**(1), Part A, 221-231.
- Walck, M. C., and Chael, E. P. (1991). Optimal backazimuth estimation for three-component recordings of regional seismic events. *Bull. Seism. Soc. Am.*, **81**, 2, 643-666.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.
- Withers, M. M., Aster, R. C., and Young, Ch., Beiriger, J., Harris, M., Moore, S., and Trujillo, J. (1998a). A comparison of selected trigger algorithms for automated global seismic phase and event location. *Bull. Seism. Soc. Am.*, **88**, 3, 95-106.
- Withers, M. M., Aster, R. C., and Young, Ch. (1998b). An automated local and regional seismic event detection and location system using waveform correlation. *Bull. Seism. Soc. Am.*, **88**, 3, 657-669.

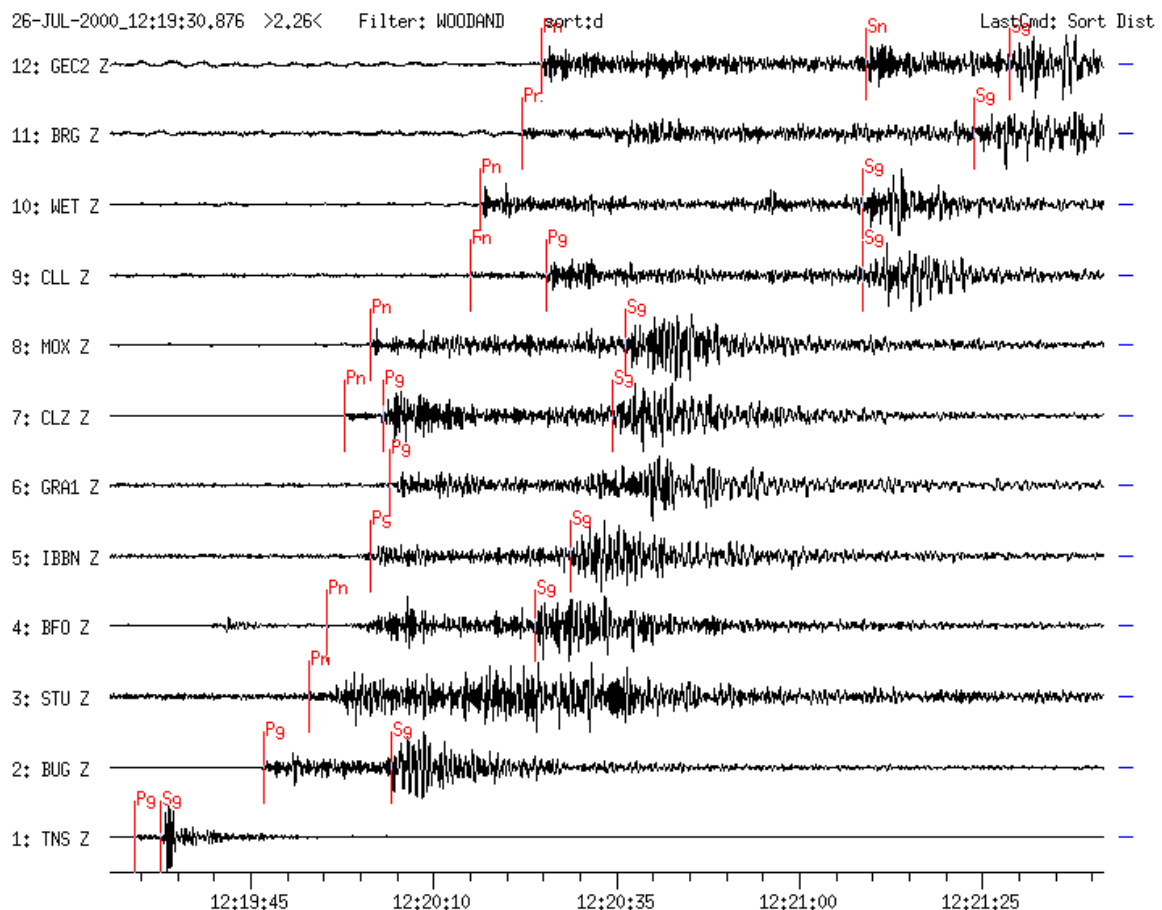


| Topic       | Additional local and regional seismogram examples                                                                                                                                                 |
|-------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| compiled by | Klaus Klinge (formerly Federal Institute for Geosciences and Natural Resources, 30655 Hannover, Germany);<br>E-mail: <a href="mailto:klaus.klinge@googlemail.com">klaus.klinge@googlemail.com</a> |
| Version     | October, 2001; DOI: 10.2312/GFZ.NMSOP-2_DS_11.1                                                                                                                                                   |

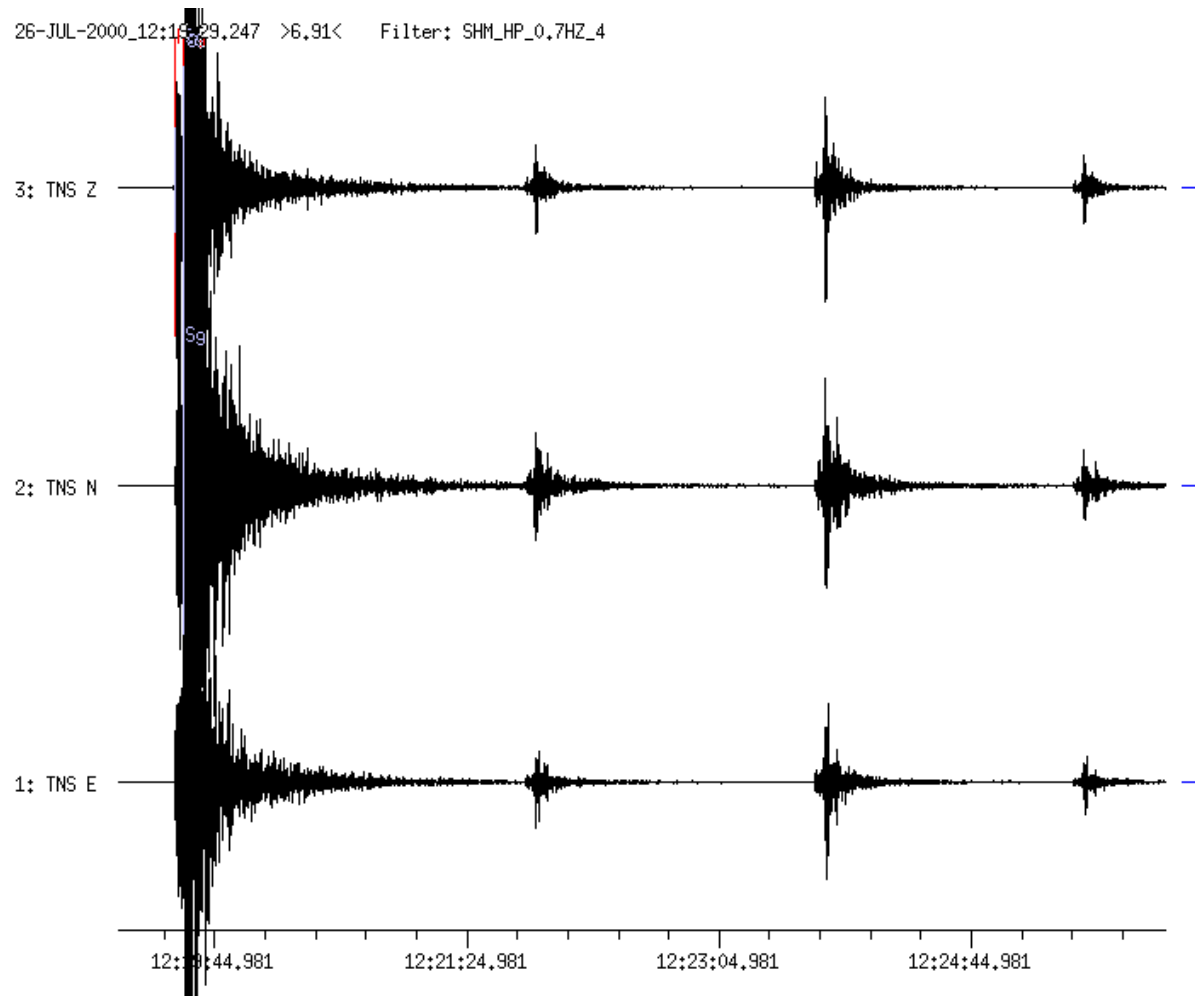
**Note:** “G” after the depth information means that the given figure (in km) is based on the estimate by a geophysicist, “N” means that the depth was assumed to be “normal” and fixed to 33 km. If the depth is given in km it has been calculated based on (depth) phase data. D – epicentral distance in degree, BAZ – backazimuth in degree, h – source depth in km.

### Example 1: Local earthquake south of Limburg/Lahn - Germany

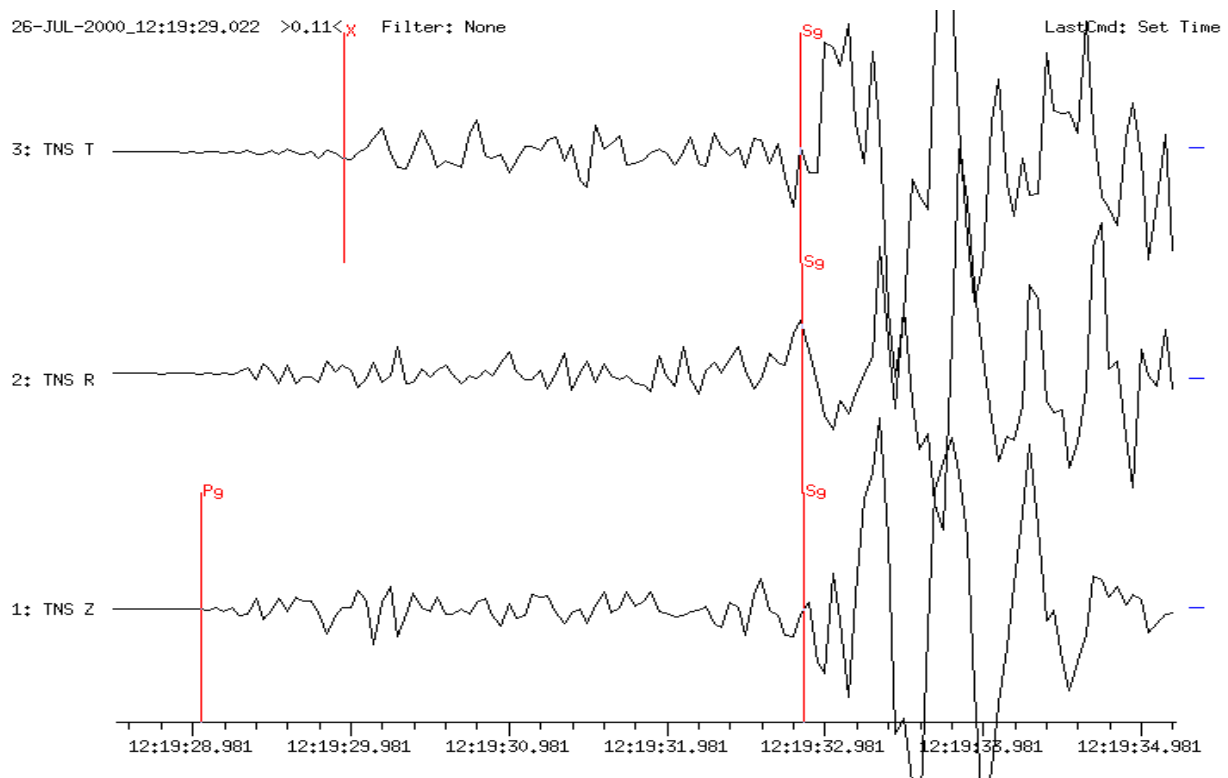
SZGRF-data: 2000-07-26 OT 12:19:23 50.25N 8.04E h = 10G Ml = 3.5



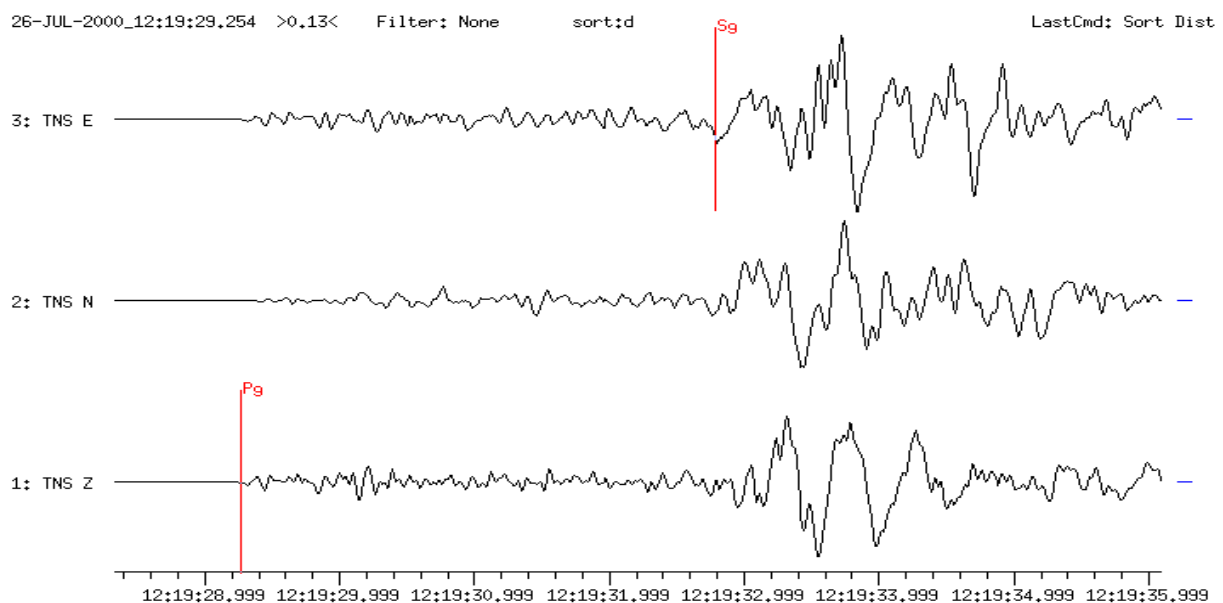
**Figure 1a** Short-period (Wood-Anderson = WA) filtered Z-component seismograms recorded at 12 GRSN/GRF stations (for network position and outlay see Fig. 11.3). Trace amplitudes are normalized and traces are sorted according to increasing epicentral distance from 29 km (TNS) to 447 km (GEC2).



**Figure 1b** High-pass filtered (0.7 to 4 Hz) 3-component record of the local station TNS (D = 29 km). A few minutes after the main shock 3 smaller aftershocks occur.



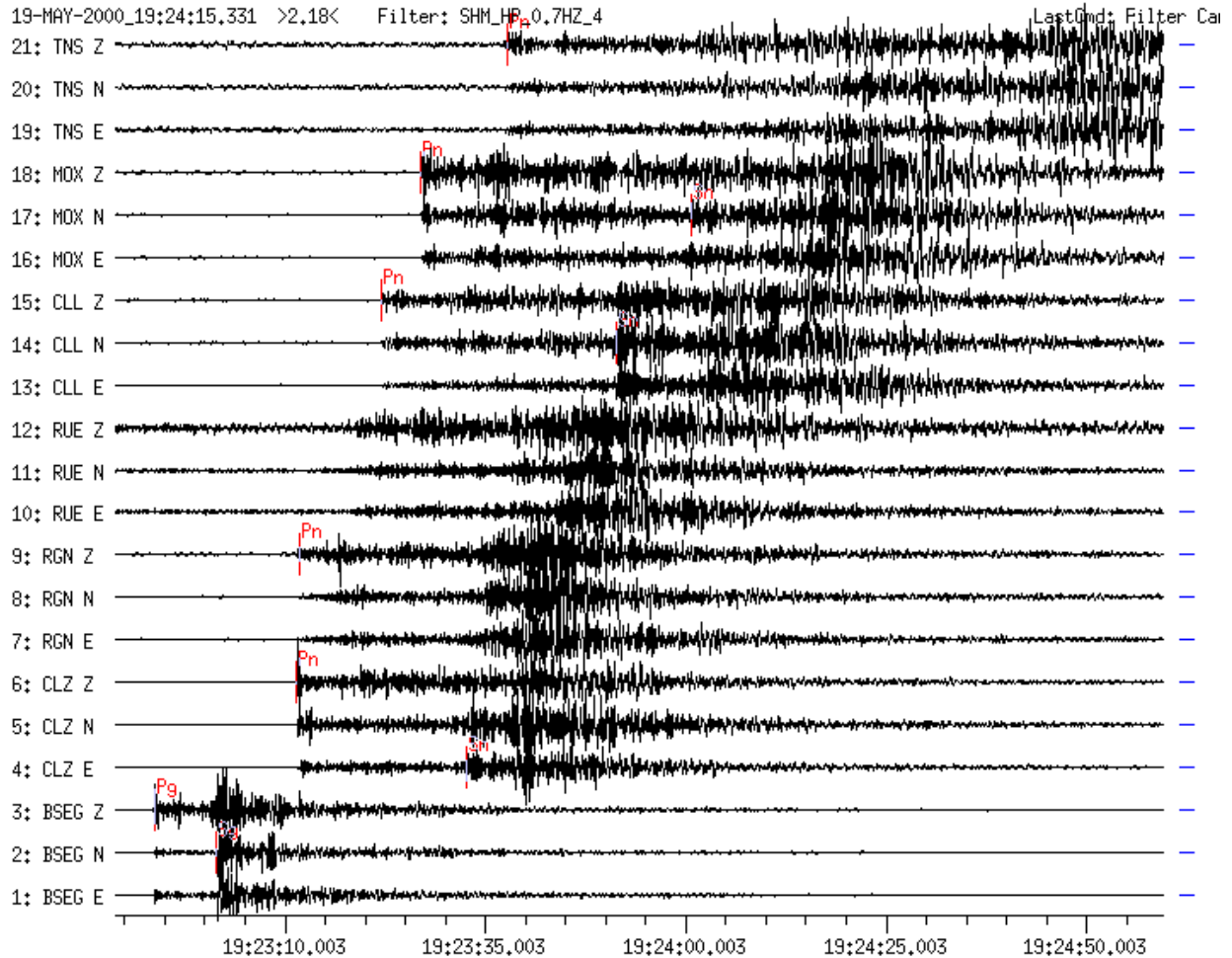
**Figure 1c** Three-component BB record of station TNS at  $D = 29$  km. Horizontal components are rotated in the R (radial) and T (transverse) directions. One second after  $P_g$  a converted P to S wave occurs on the T component (mark X). The sampling rate is 20Hz. Note: “Filter: None” in the uppermost line always means velocity broadband record.



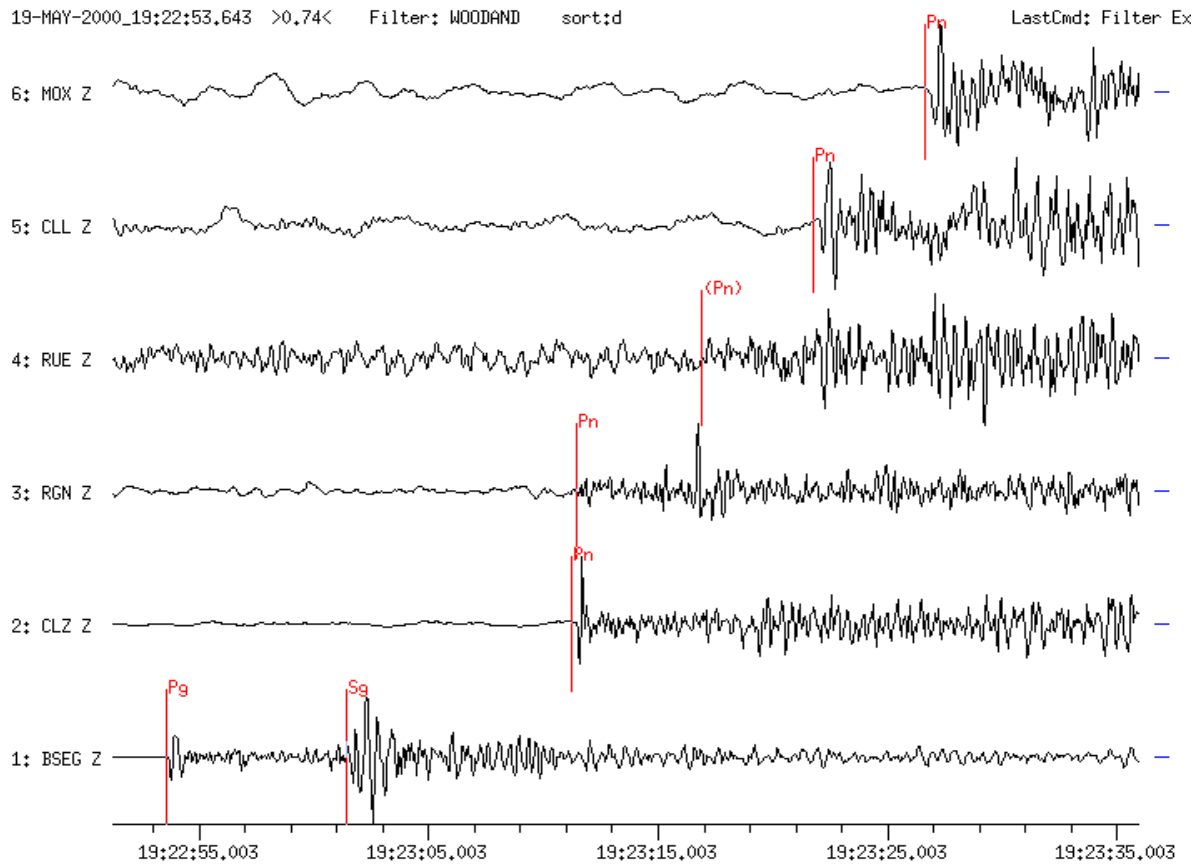
**Figure 1d** Three-component BB record of station TNS. The sampling rate is 80 Hz.

**Example 2: Local earthquake in Northern Germany. Generally, this region is regarded as aseismic.**

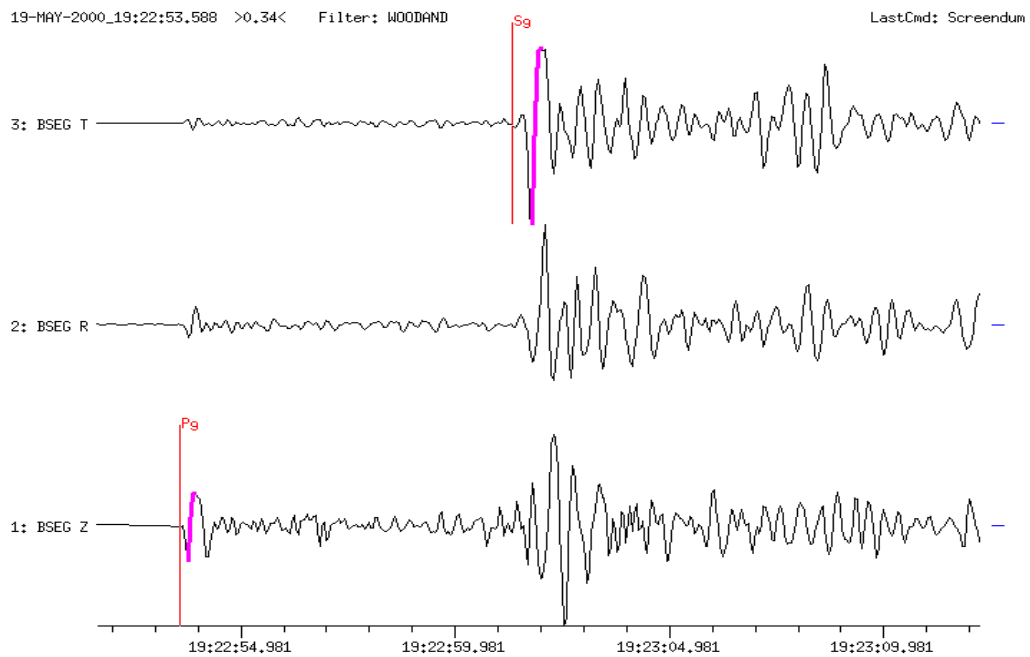
SZGRF-Daten: 2000-05-19 OT 19:22:40.8(UTC) 53.47N 11.10E  $M_l = 3.4$



**Figure 2a** High-pass filtered (0.7 to 4 Hz) 3-component records of 7 GRSN stations. The traces have been sorted according to increasing distance ranging from  $D = 73$  km (BSEG) to 405 km (TNS).



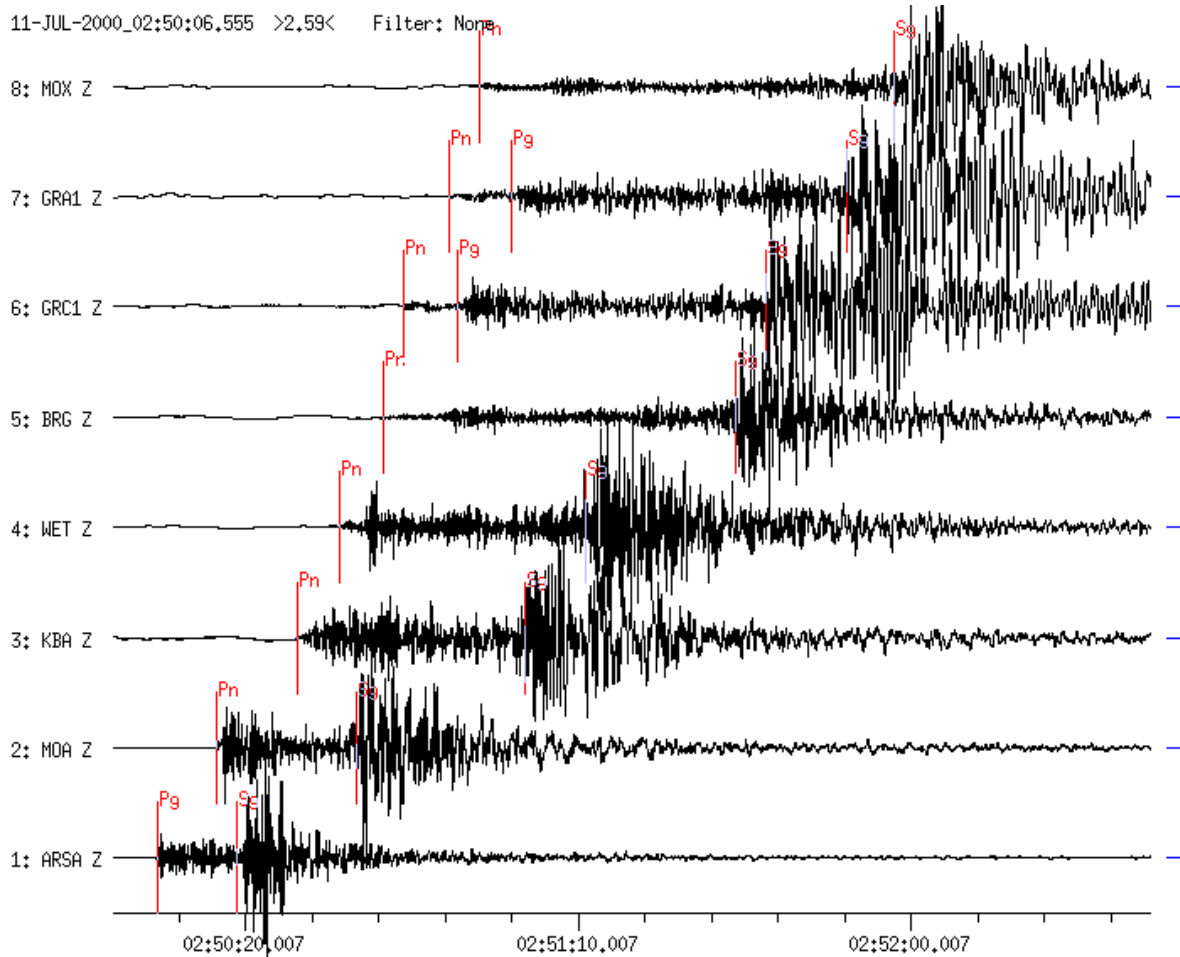
**Figure 2b** Short-period (WA) filtered Z-component seismograms of the same earthquake as in Figure 2a. First motion polarities can be read from traces 1, 2, 5 and 6, only.



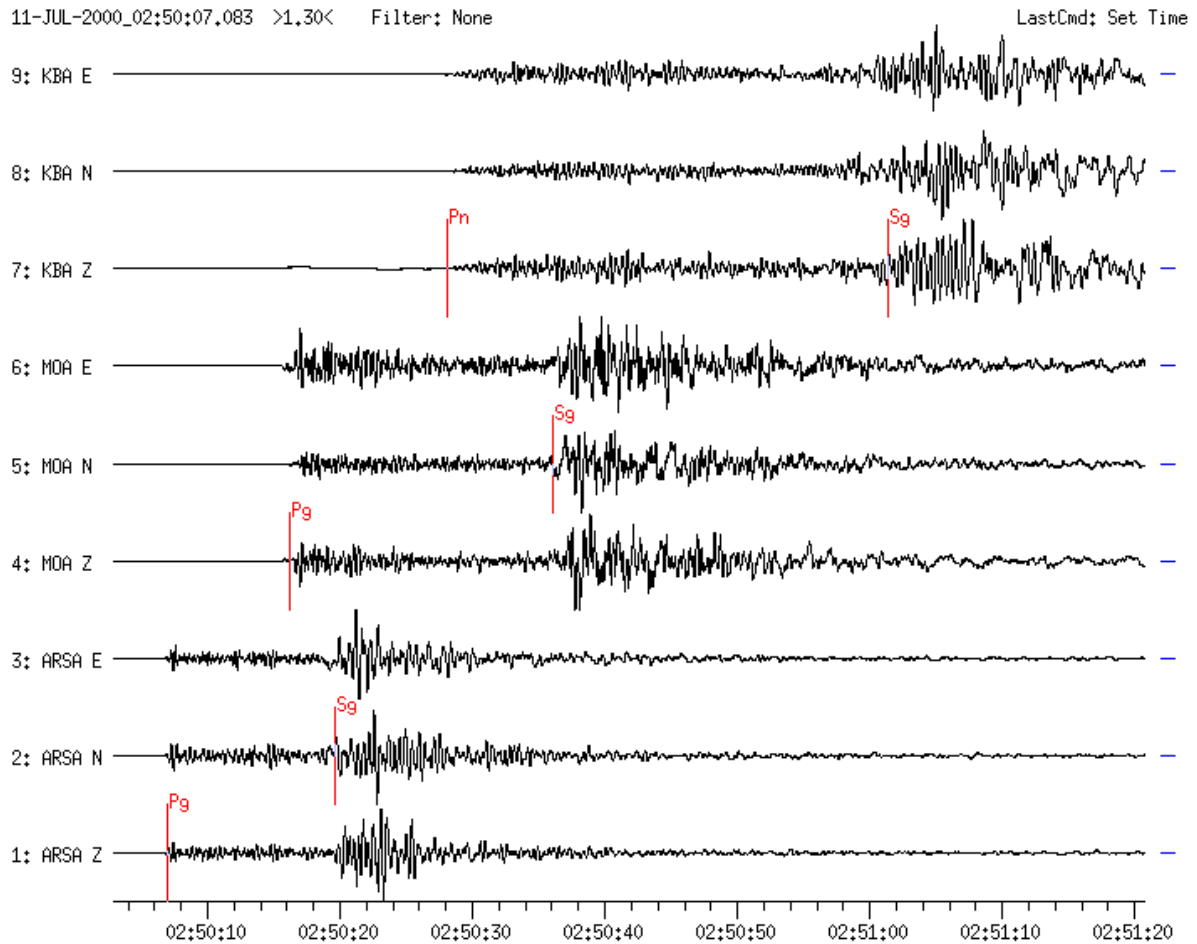
**Figure 2c** Three-component WA record at station BSEG at  $D = 73$  km and  $BAZ = 135^\circ$ . The radial component R shows in the direction of wave propagation, the transversal component T is perpendicular to R.

### Example 3: Regional earthquake south of Wien

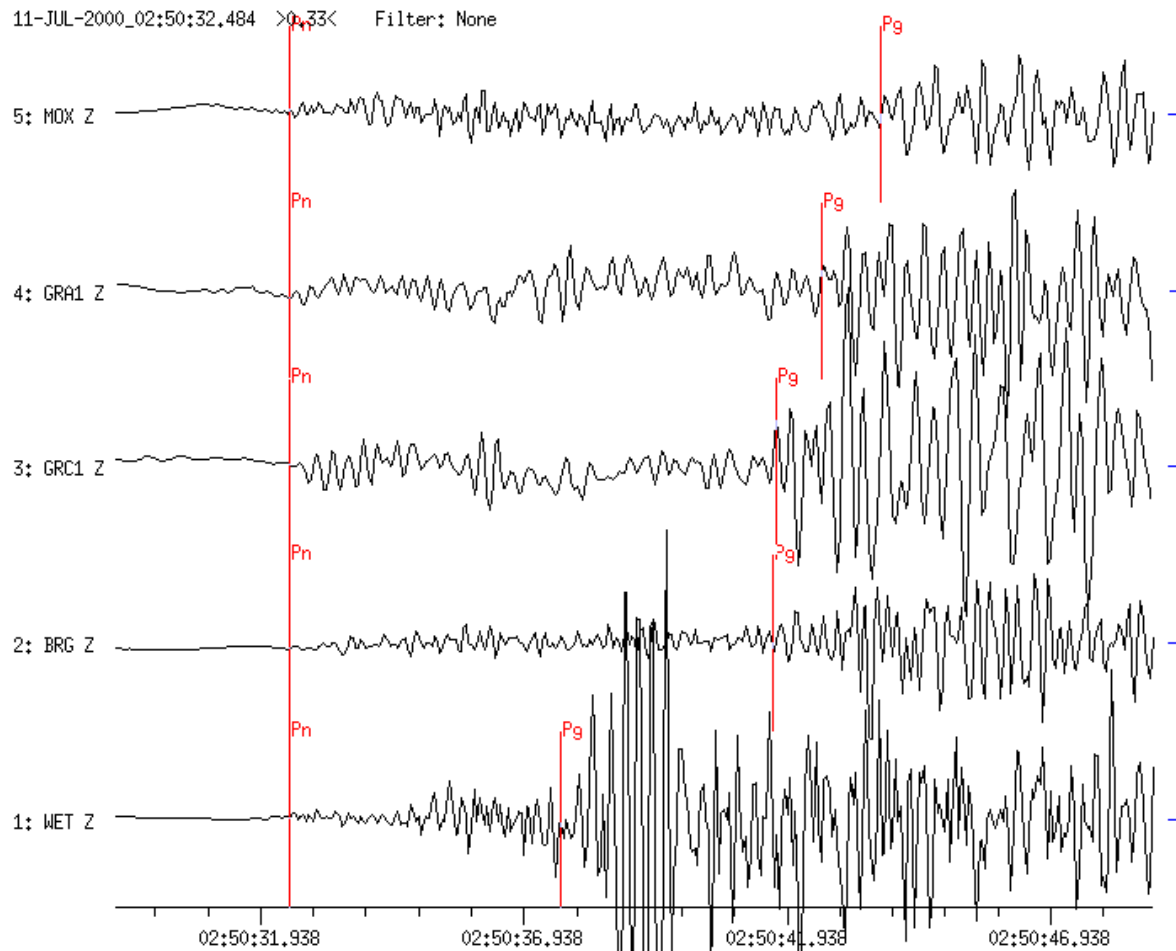
SZGRF-data: 2000-07-11 OT 02:49:51(UTC) 48.10N 16.40E Ml = 5.2



**Figure 3a** Vertical-component BB records of 5 GRSN/GRF stations and 3 Austrian stations (ARSA, MOA, KBA). Seismogram trace amplitudes have been normalized and the traces sorted according to increasing distance (D = 100 km to ARSA is 100 km and 490 km to MOX).



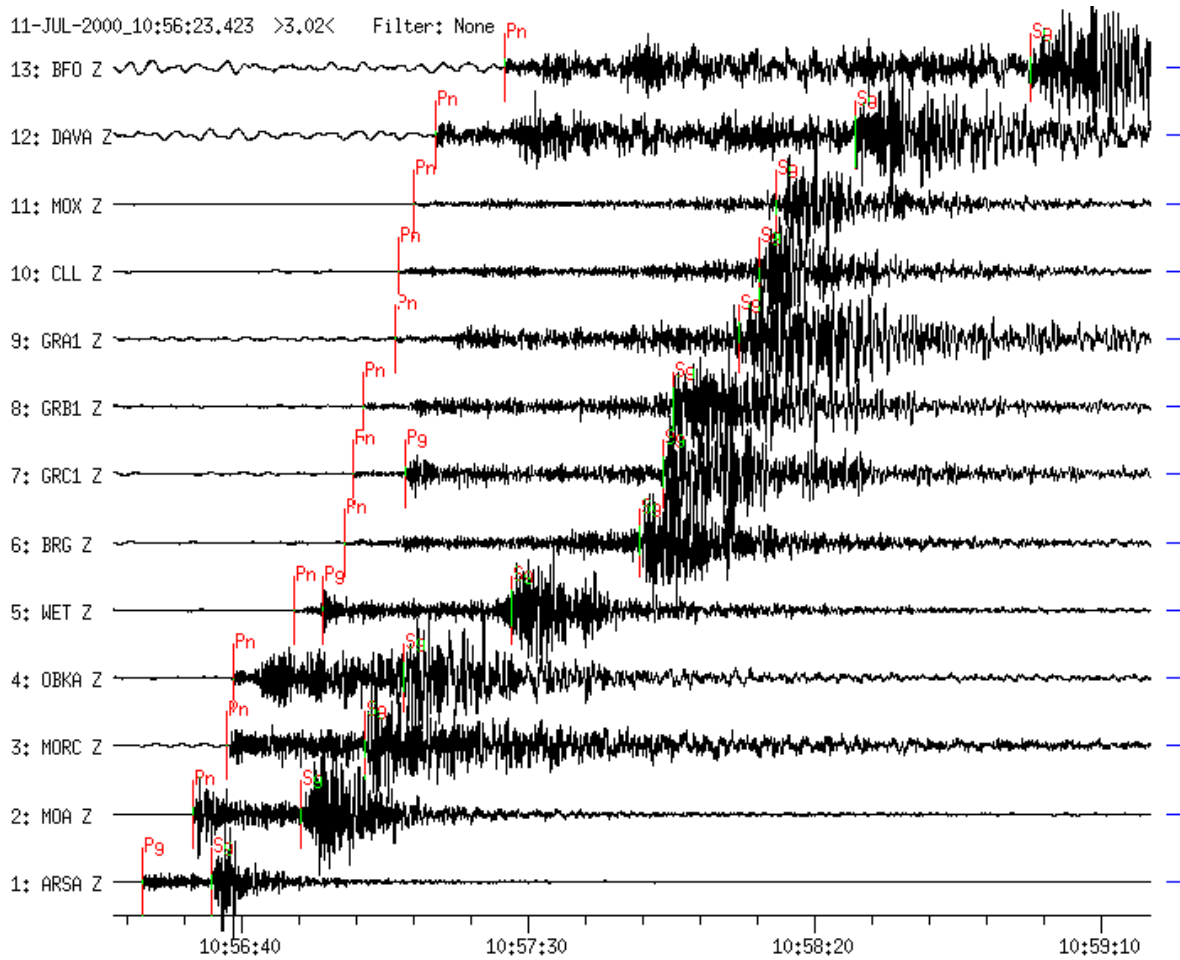
**Figure 3b** Three-component BB records of the stations ARSA ( $D = 100$  km), MOA ( $D = 160$  km) and KBA ( $D = 250$  km) of the Austrian network (ZAMG Wien).



**Figure 3c** Vertical-component BB records of 5 GRSN stations with phases Pn and Pg. Traces are shifted and aligned for Pn according to a reference station (WET) at  $D = 284$  km. Recorded Pn-onsets are weak and polarity readings are impossible



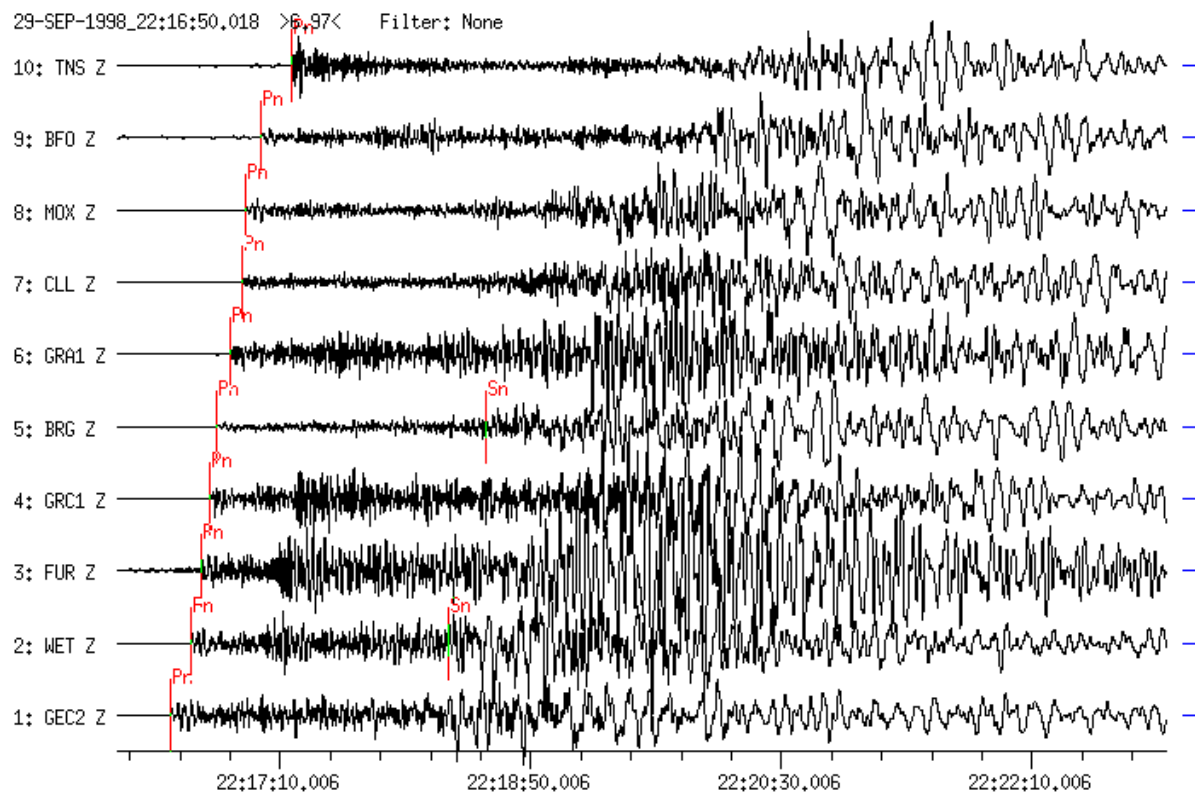
**Aftershock: 2000-07-11 OT 10:56:04.5 (UTC) 48.01N 16.48E MI = 4.7**



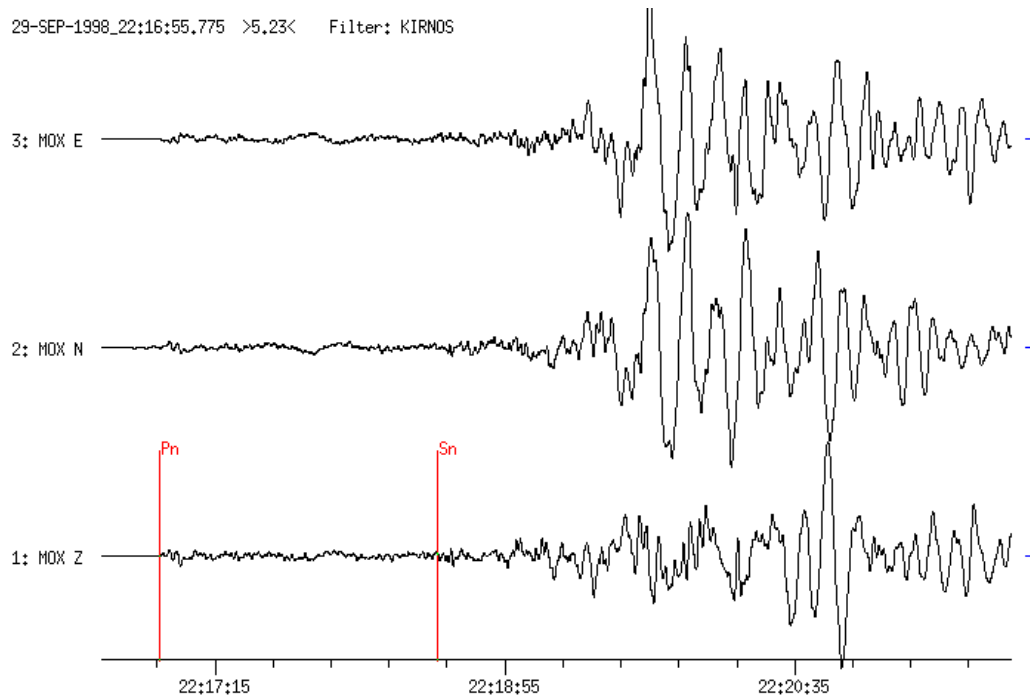
**Figure 3d** Vertical-component BB records of 8 German GRSN/GRF-stations (WET, BRG, GRC1, GRB1, GRA1, CLL, MOX, BFO), 4 Austrian ZAMG-stations (ARSA, MOA, OBKA, DAVA) and 1 Czech GEOFON-station (MORC). Trace amplitudes have been normalized and the stations sorted according to increasing distance ( $D = 96$  km to ARSA and 920 km to BFO).

### Example 4: Earthquake in Yugoslavia

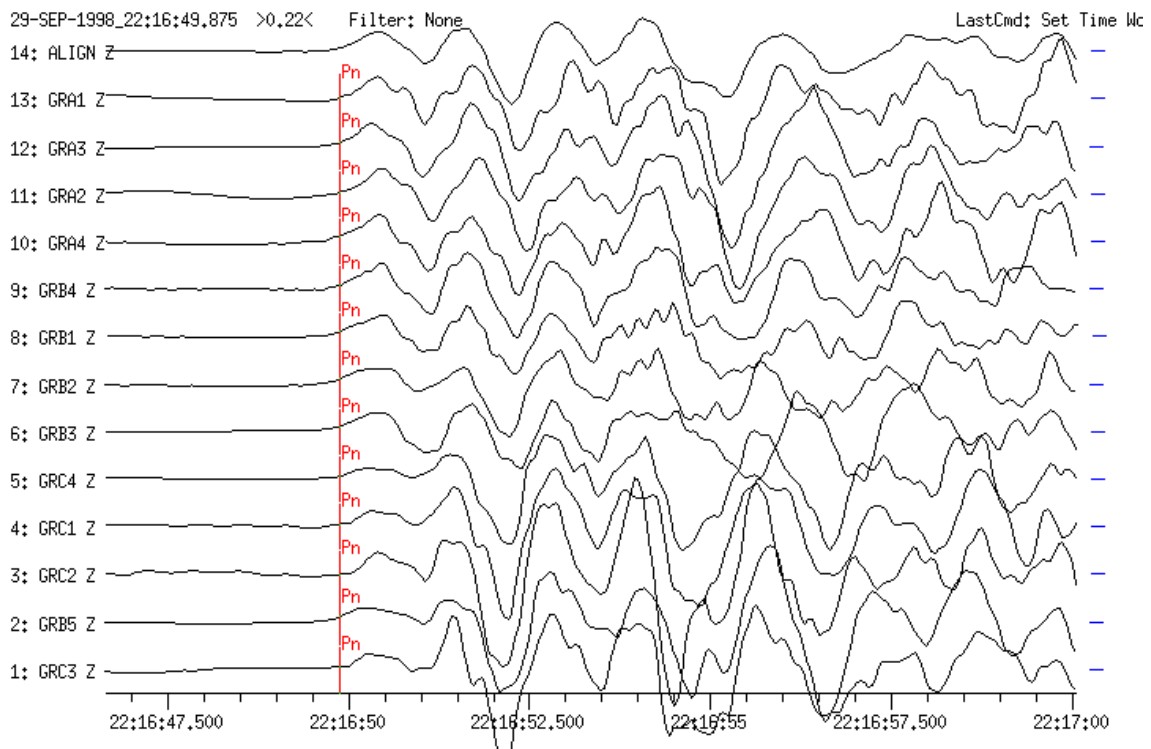
NEIC-data: 1998-09-29 OT 22:14:50 44.11N 20.04E h = 10km mb = 5.2  
(D = 8.2° and BAZ = 130° from GRA1)



**Figure 4a** Vertical-component BB records of 10 GRSN/GRF-stations sorted according to increasing distance (D = 6.45° to GEC2 and 10.0° to TNS).



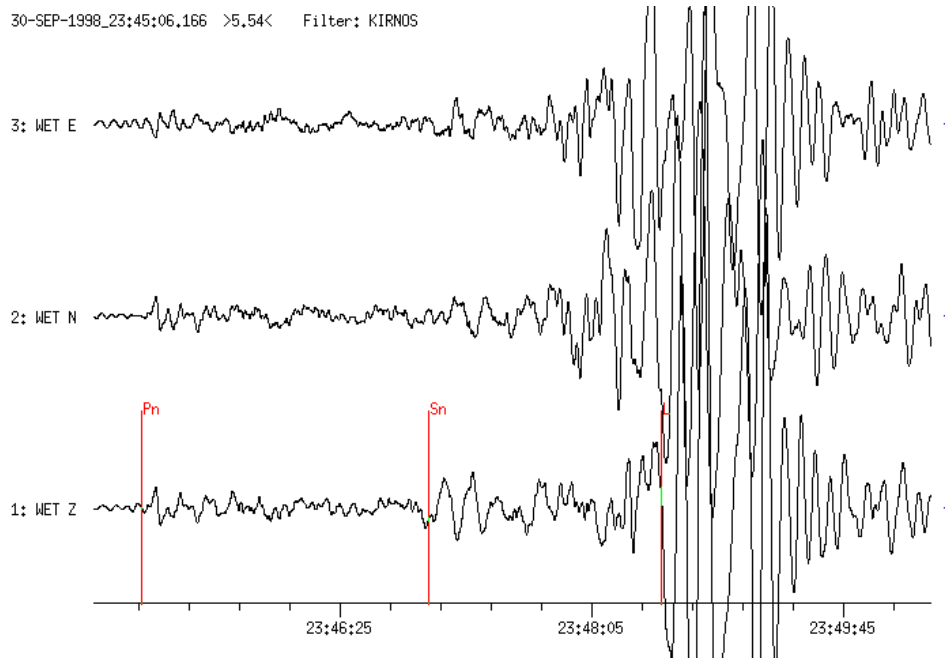
**Figure 4b** Three-component BB-displacement (Kirnos filtered) record of station MOX ( $D = 8.7^\circ$ ,  $BAZ = 136^\circ$ ) with phases Pn, weak Sn and strong dispersed surface waves (LQ onset in N-E around 22:19:10 and  $LR_{max}$  in Z at 22:20:45; note onset-like Lg phases arriving between Sn and LQ).



**Figure 4c** Time-shifted and aligned vertical components BB records of the 13 GRF-array stations (see Fig. 11.3a for array position and outline). Traces are sorted according to increasing distance.

### Example 5: Earthquake in Albania

NEIC-data: 1998-09-30 OT 23:42:54 41.95N 20.39E h = 10km Ms = 5.1  
(D = 10.0° and BAZ = 137° from GRA1).



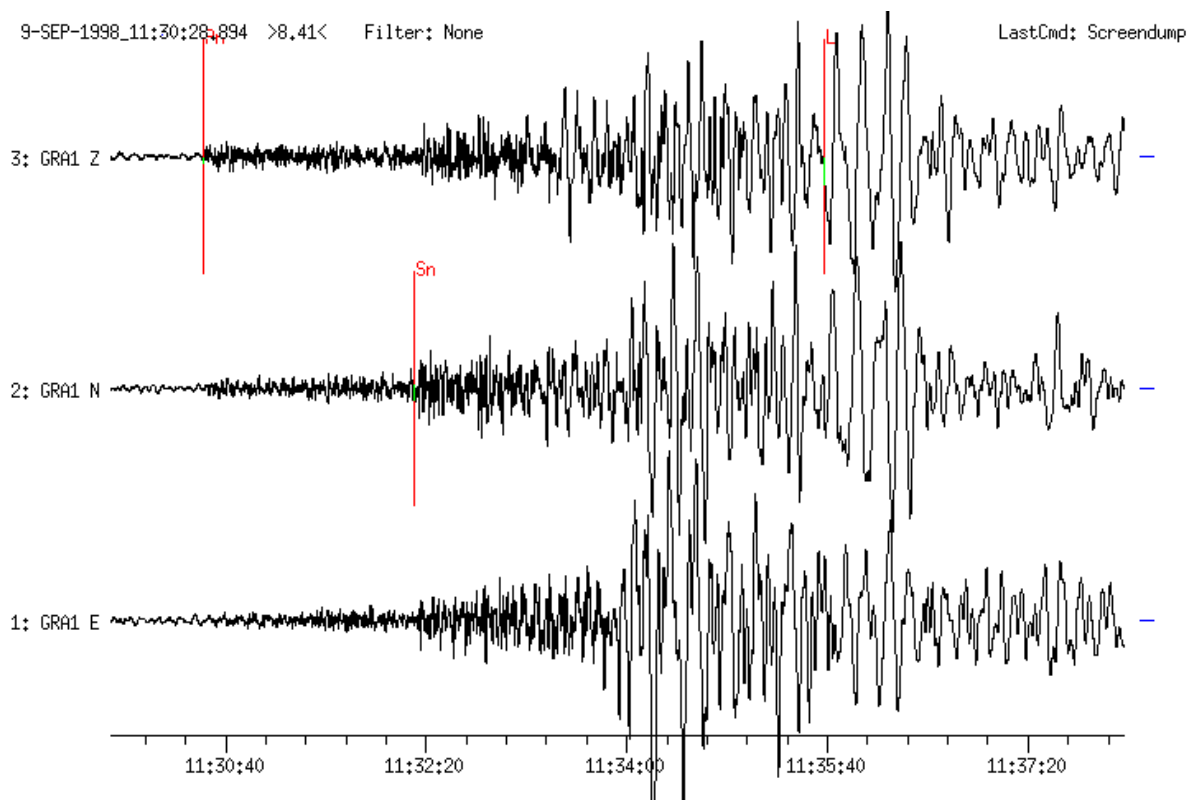
**Figure 5** Three-component BB-displacement (Kirnós filtered) record at the GRSN station WET at D = 8.9° (BAZ = 136°). Note the clear onset of Pn, a very pronounced long-period Sn (as compared to the very weak Sn in the record of the Yugoslavia earthquake in Figure 4b above) and well dispersed surface waves of dominantly Rayleigh (LR) type (because of the strong vertical component).

### Example 6: Earthquake in SOUTHERN ITALY

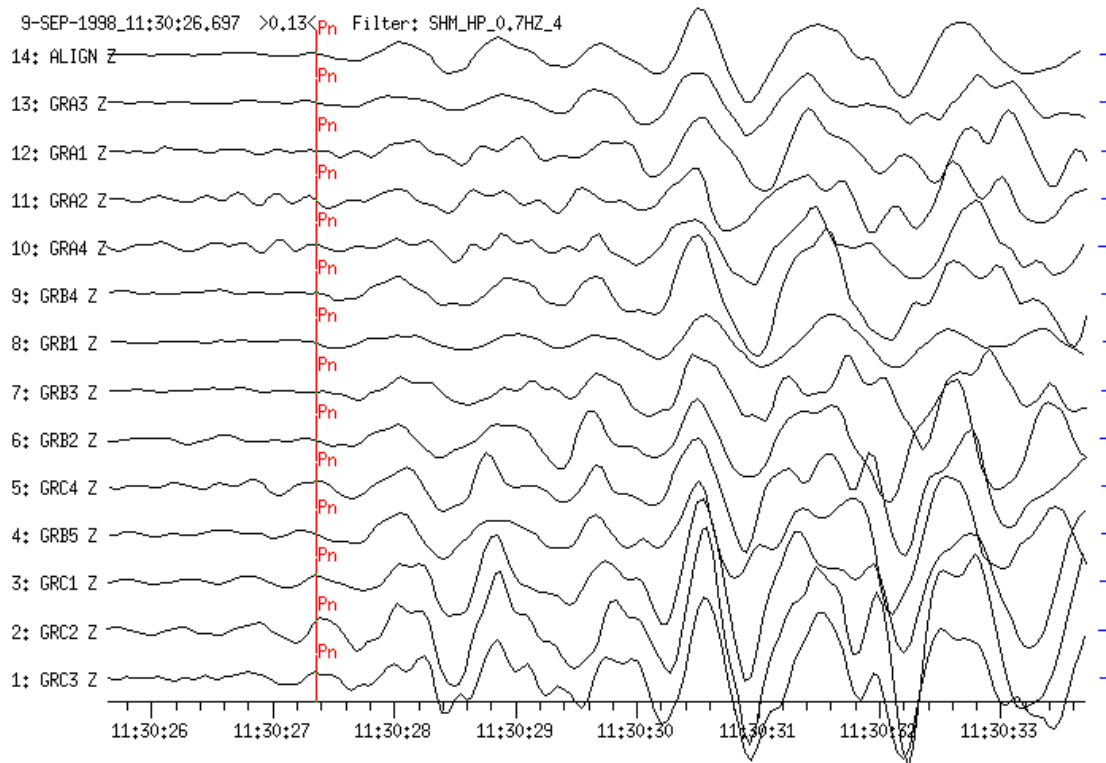
USGS NEIC-data: 1998-09-09 OT 11:27:58.6 39.964N 15.948E h = 10km  
mb = 5.3 Ms = 5.2

SZGRF-data: 1998-09-09 OT 11:28:01.8 40.1N 16.4E Ms = 5.2

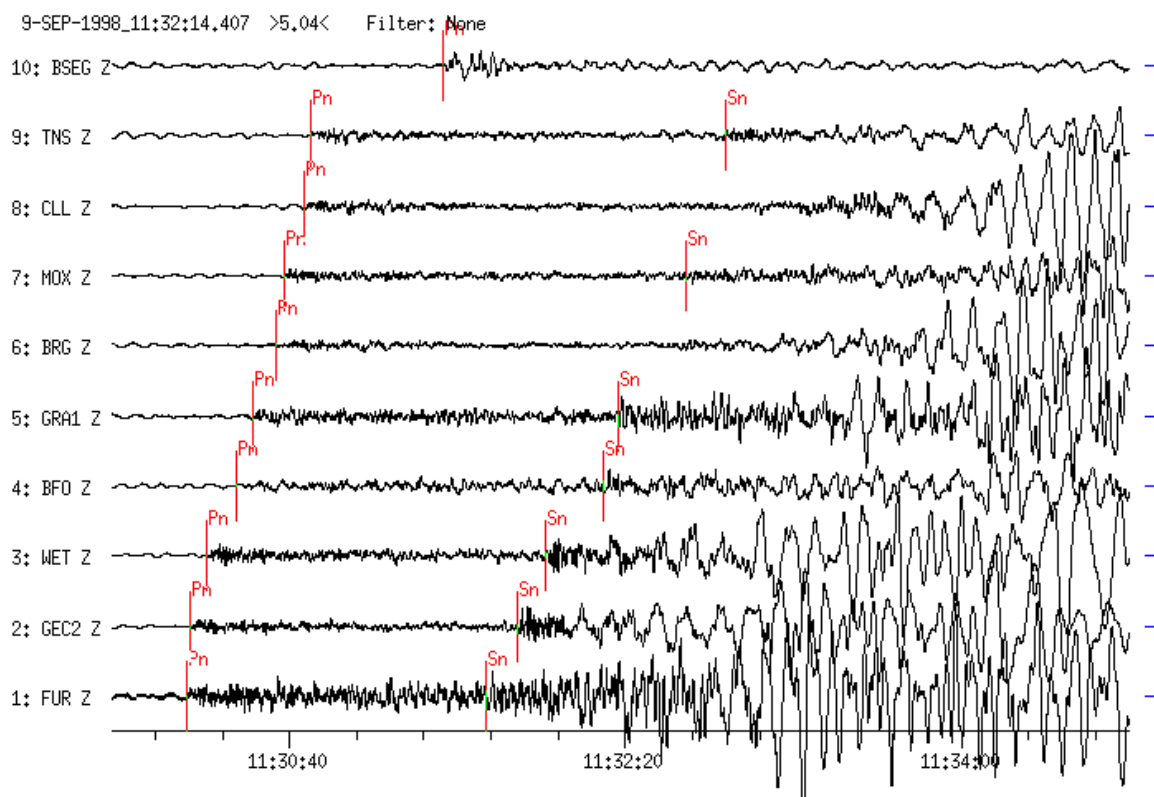
Distance (GRA1) D = 10.3 deg, BAZ = 157 deg



**Figure 6a** Three-component BB record at the GRF-station GRA1 ( $D = 10.3^\circ$ ,  $BAZ = 159^\circ$ ) with clear phases Pn, Sn and surface waves (Lg arriving around 11:34:00 and Rg around 11:35.40).



**Figure 6b** Highpass-filtered (0.7 to 4 Hz) Z-component records with Pn onsets at 13 GRF-array stations. Traces are aligned and sorted according to increasing distance ( $D = 9.45^\circ$  to GRC3 and  $D = 10.33^\circ$  to GRA3). The coherency of Pn is poor at this distance range.



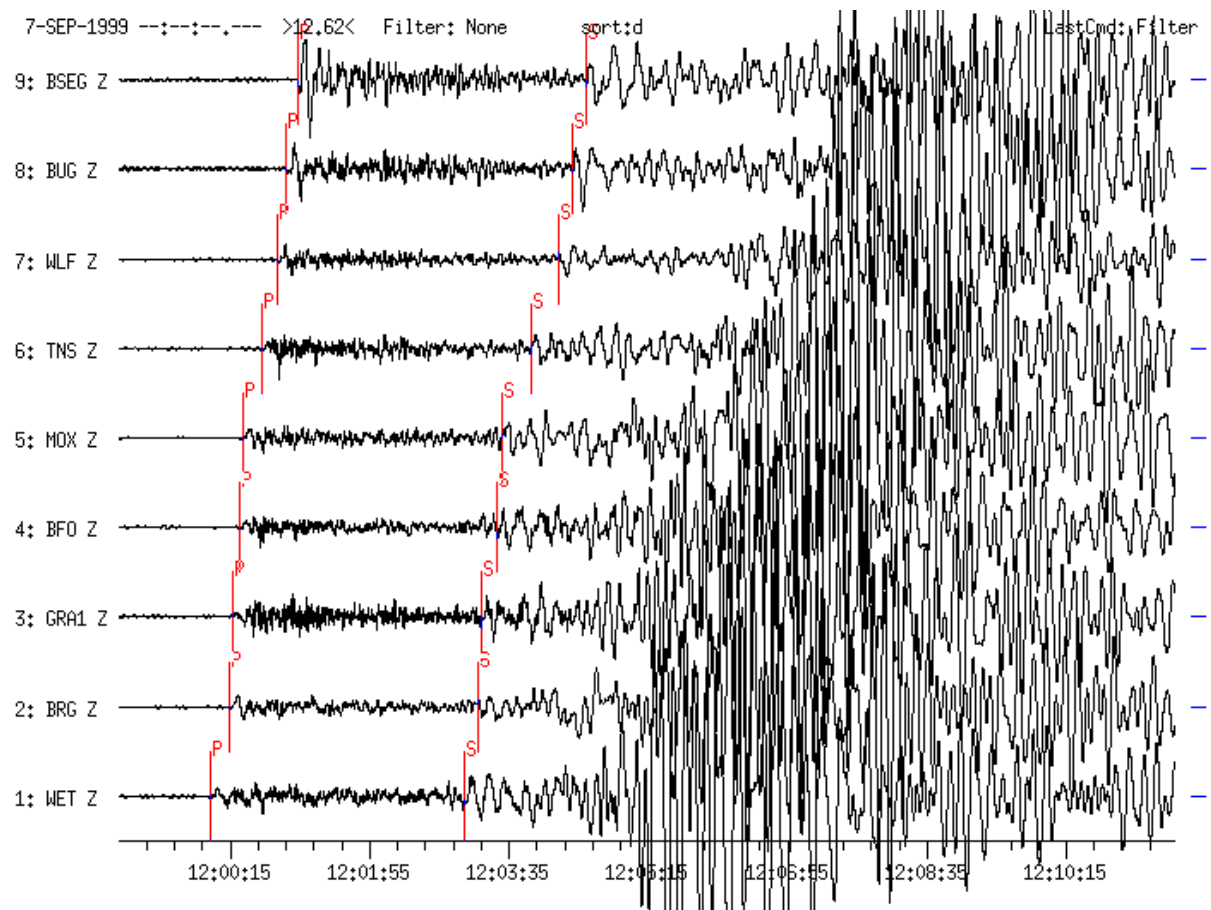
**Figure 6c** Vertical-component BB records with Pn and Sn waves from 10 GRSN-stations. The traces have been sorted according to increasing distance between  $D = 8.86^\circ$  (FUR) and  $14.5^\circ$  (BSEG). Except for stations CLL, BRG and BSEG clear Sn arrivals are visible.

| Topic       | Additional seismogram examples within the distance range 13° - 100°                                                                                                                               |
|-------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| compiled by | Klaus Klinge (formerly Federal Institute for Geosciences and Natural Resources, 30655 Hannover, Germany);<br>E-mail: <a href="mailto:klaus.klinge@googlemail.com">klaus.klinge@googlemail.com</a> |
| Version     | October, 2001; DOI: 10.2312/GFZ.NMSOP-2_DS_11.2                                                                                                                                                   |

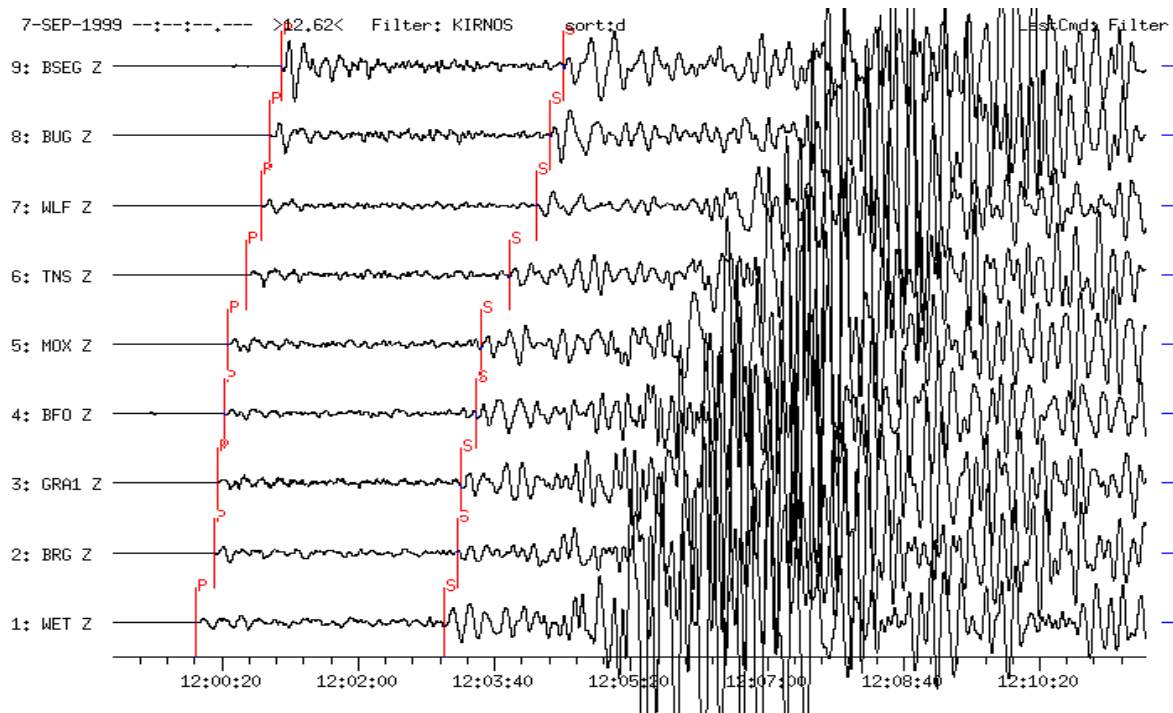
**Note:** Most of the examples given below show either records of the German Regional Seismic Network (GRSN; aperture about 500 x 800 km) or of the Gräfenberg broadband array (GRF; aperture 45 x 110 km; see Figs. 8.14 and 9.4 in the manual Chapters 8 and 9). The following abbreviations have been used: D – epicentral distance in degree, BAZ – backazimuth in degree, h – focal depth in kilometer. Complementary comments have been added by the Editor.

### Example 1: Earthquake in Greece

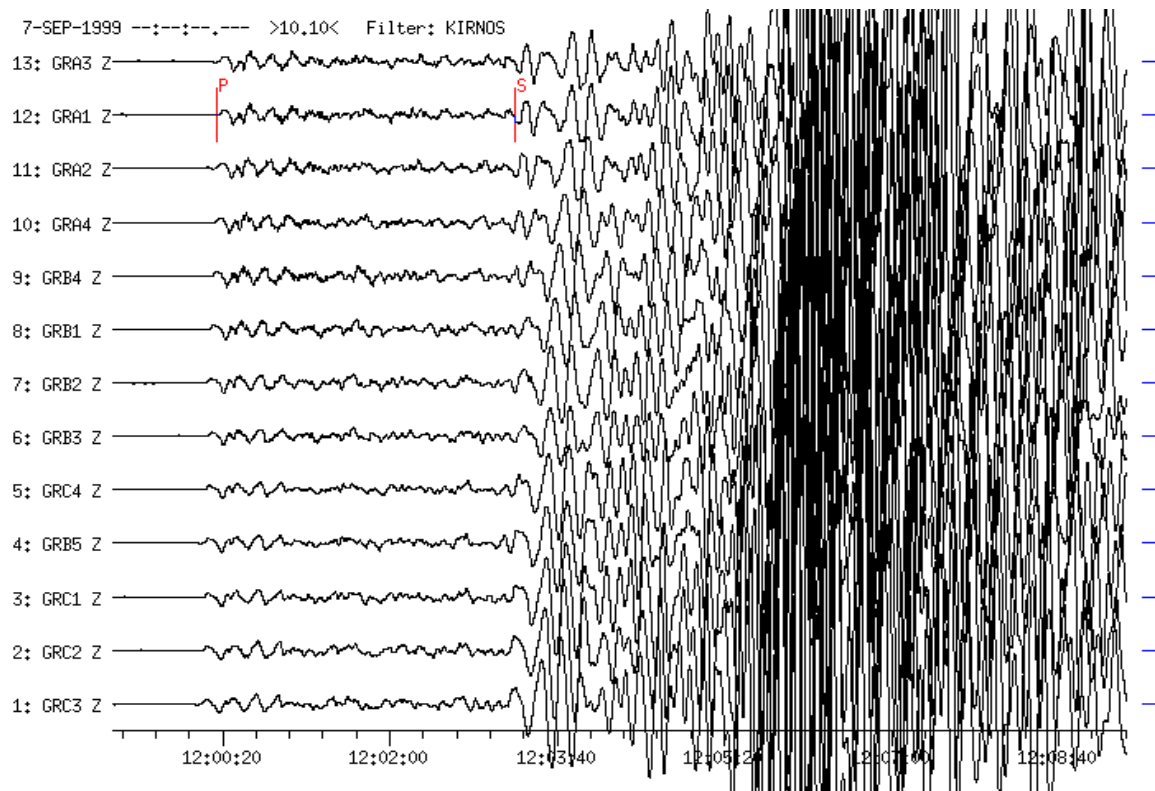
USGS NEIC-data:1999-09-07 OT 11:56:50 38.13N 23.55E h = 10km mb = 5.8



**Figure 1a** Broadband vertical-component seismograms recorded at 9 GRF/GRSN stations within the distance range  $D = 13.4^\circ$  (WET) to  $18.2^\circ$  (BSEG). Traces have been sorted according to increasing distance. A complex P wave is followed by S and surface waves with longer-periods than P. Both body waves are influenced by upper mantle discontinuities. Note the large P-wave amplitude at the most distant station BSEG because amplitudes increase rapidly towards the “20° discontinuity” (see Fig. 3.13 in Chapter 3).

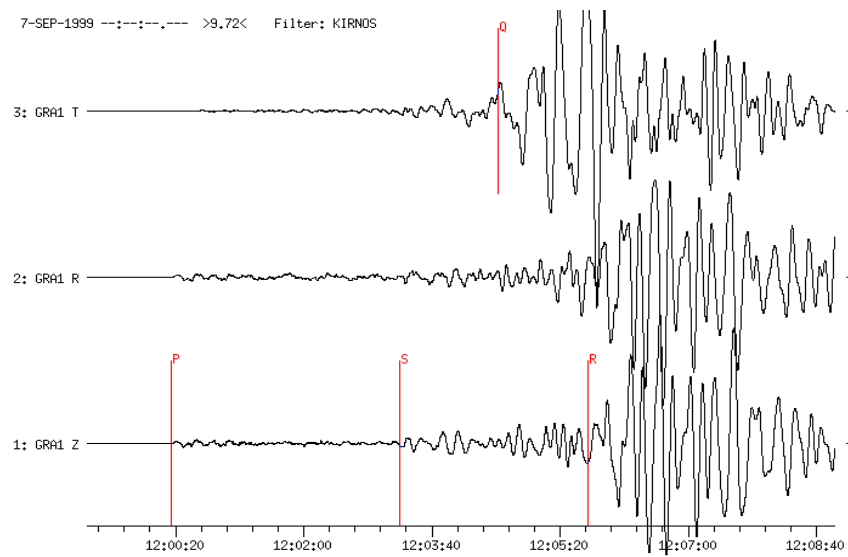


**Figure 1b** The same record as above, however a displacement proportional KIRNOS-filter was applied. Especially S waves are displayed better by using this filter. The network traces are not coherent.

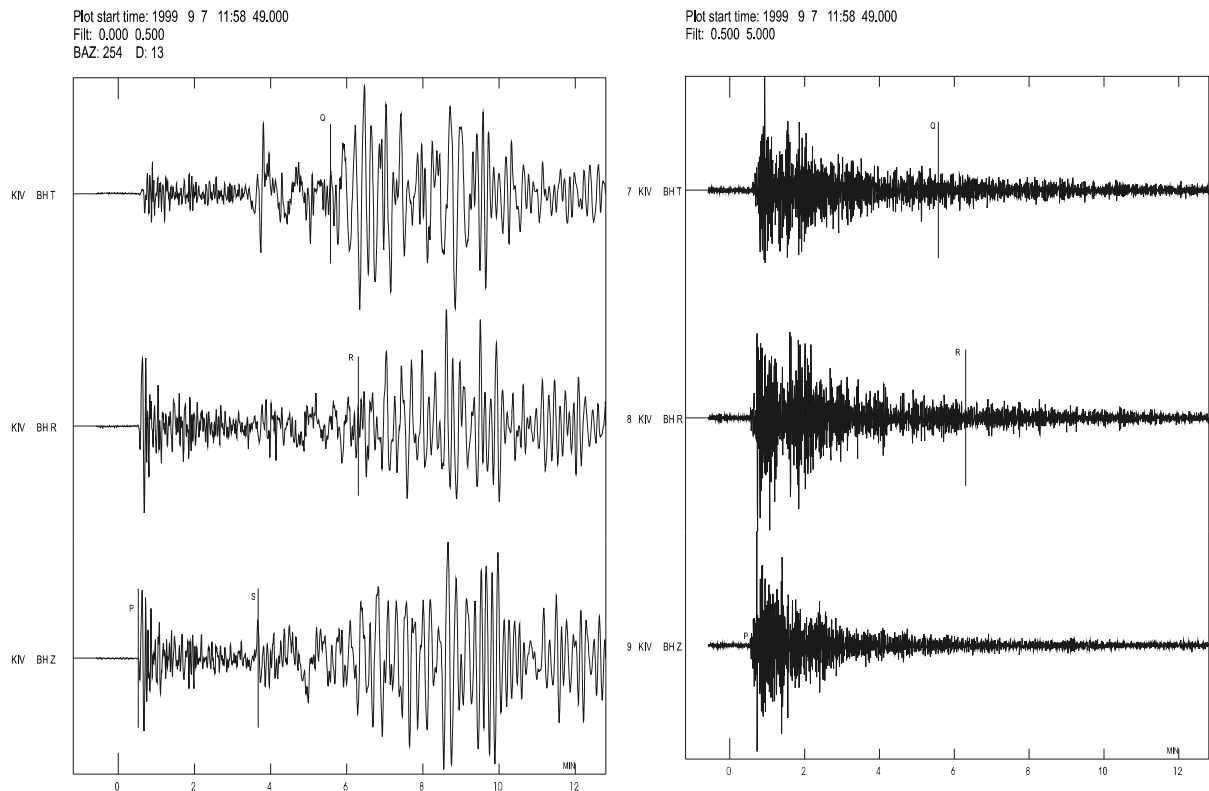


**Figure 1c** KIRNOS-filtered seismograms of vertical-component records at the GRF-array stations sorted according to increasing epicentral distance ( $D = 13.8^\circ$  to GRC3 and  $14.6^\circ$  to GRA3). Note the variability of waveforms within the S-wave group.





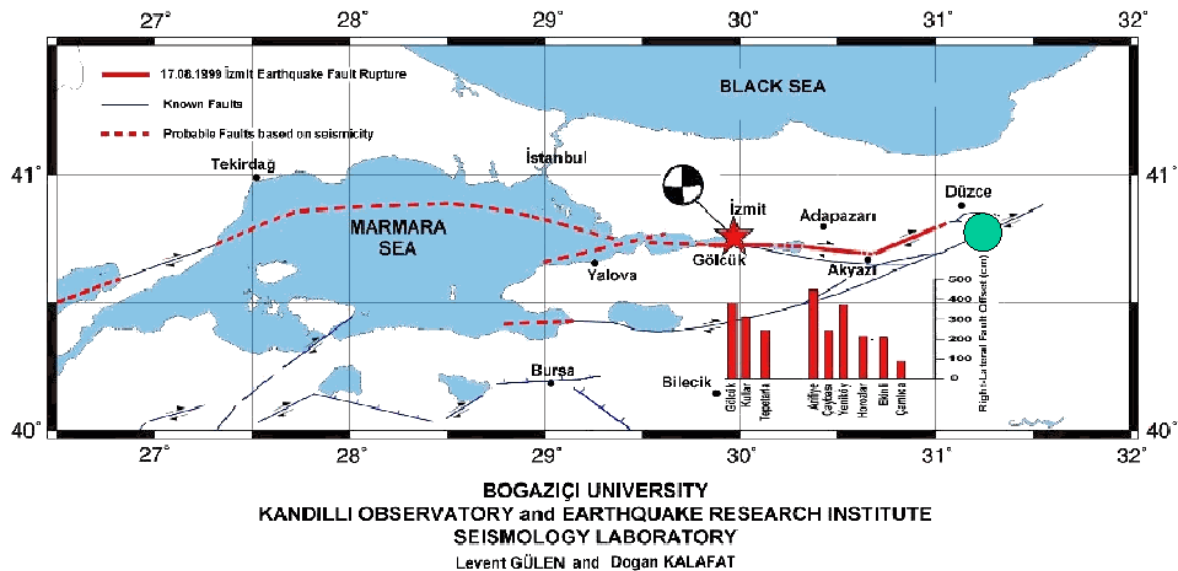
**Figure 1d** KIRNOS-filtered three-component seismogram from the GRF-array station GRA1 ( $D = 14.55^\circ$ ,  $BAZ = 138^\circ$ ). The horizontal components N and E have been rotated into the radial (R) and transverse (T) direction. The onsets of the body waves P and S and of the long-period surface waves LQ (Q) and LR (R) have been marked.



**Figure 1e** Low-pass (left) and high-frequency band-pass filtered (0.5-5 Hz) seismograms of the same earthquake recorded at station KIV (Kislovodsk; Russia) at the distance  $D = 15.2^\circ$ . The horizontal components N and E have been rotated into the R and T direction. No long-period waves are visible in the high-frequency record (courtesy of Lars Ottemöller, 2002).

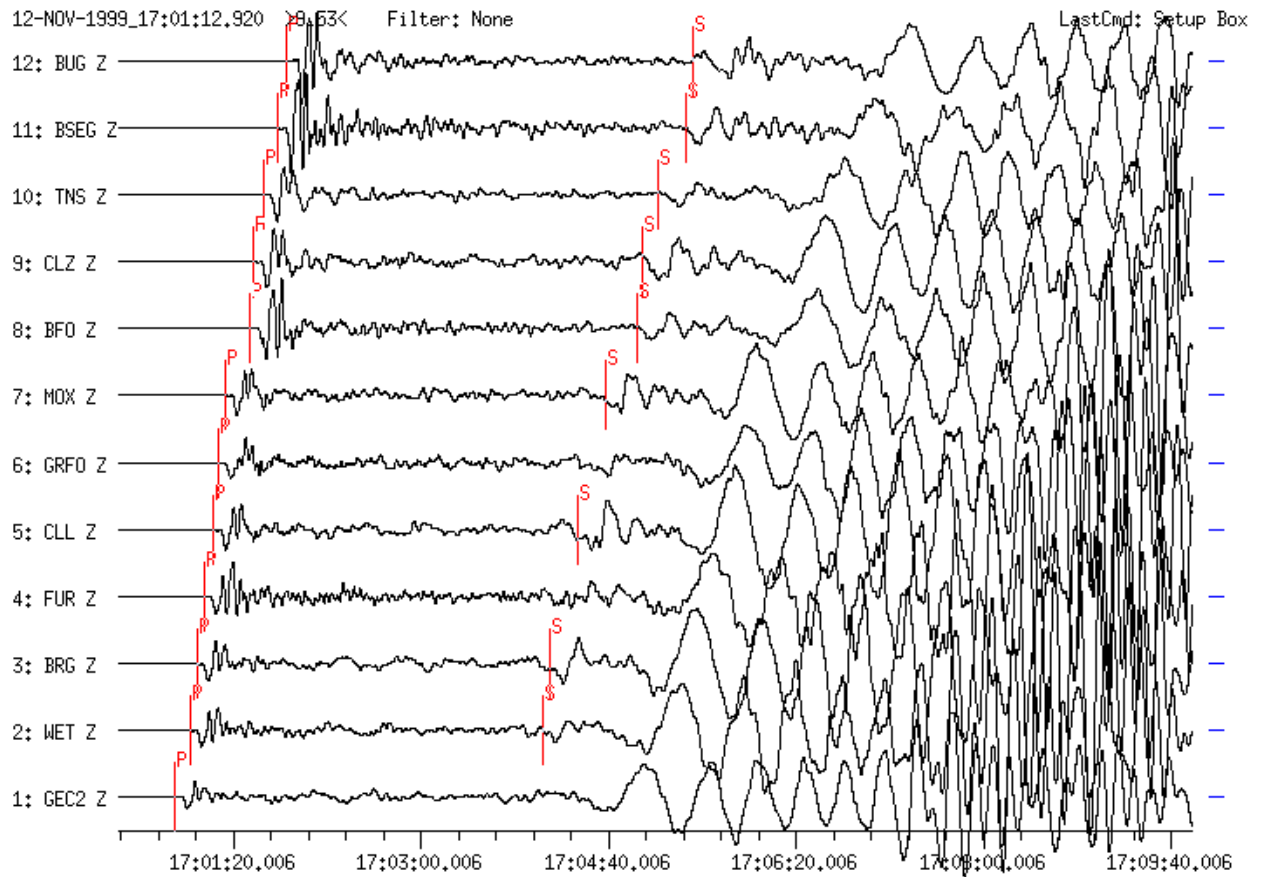
## Example 2: Earthquake in NW-TURKEY

USGS NEIC-data:1999-11-12 OT 16:57:20 40.79N 31.11E h = 10km  $M_w = 7.1$   
( $D = 16.5^\circ$  to GRF,  $BAZ = 115^\circ$ )

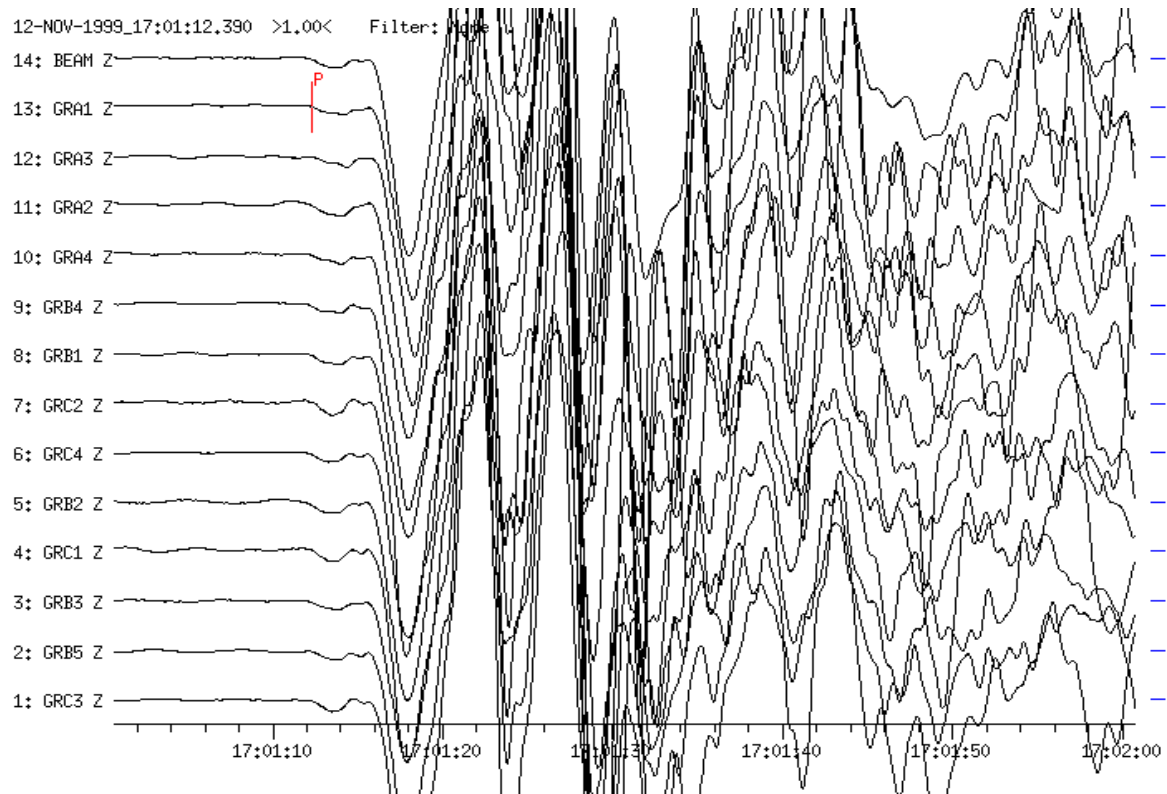


12.11.1999  $M_w 7.1$  (NEIC)

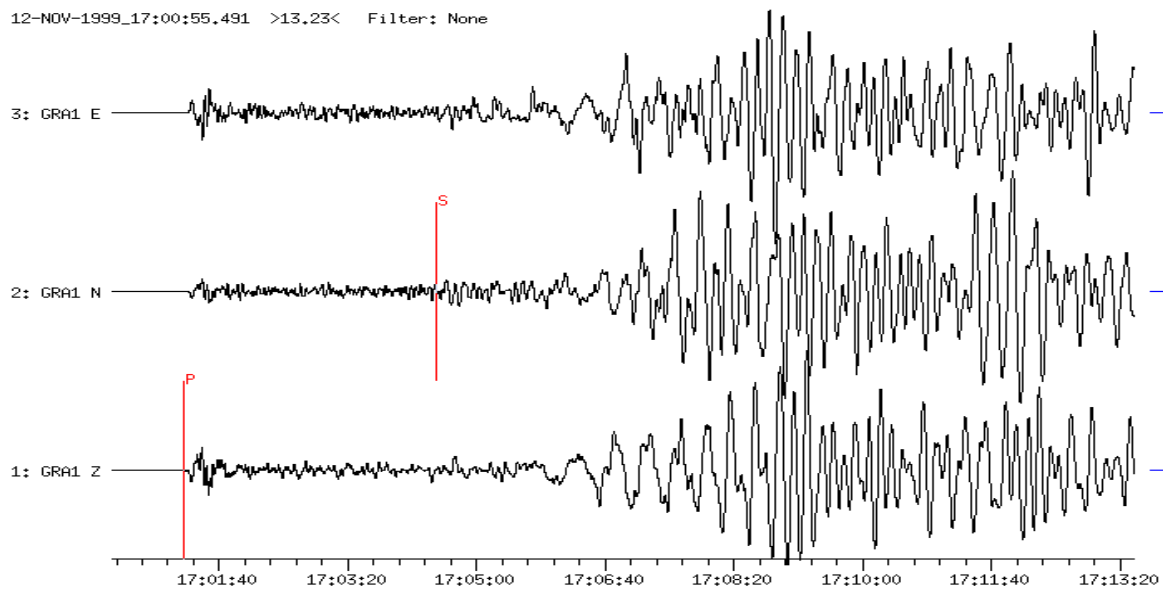
**Figure 2a** The Düzce earthquake of November 12, 1999 occurred about 110 km east of the earlier Izmit earthquake of August 17, 1999. The map shows the epicenter regions of both earthquakes together with one moment-tensor solution (for the Izmit event) and the right lateral surface displacement values (in cm) observed after the first shock.



**Figure 2b** Broadband vertical-component seismograms recorded at 12 GRSN stations. Traces are sorted according to increasing distance ( $D = 14.7^\circ$  from GEC2 and  $19.6^\circ$  from BUG). A clear P wave is followed by a relatively weak S and strong dispersed surface waves with much longer-periods than P. The P and S waveforms, influenced by upper mantle discontinuities, show a complicated structure. Note the growing P-wave complexity and amplitudes with increasing distance towards the “20° discontinuity” (see Fig. 2.29 in Chapter 2 and Fig. 3.13 in Chapter 3).



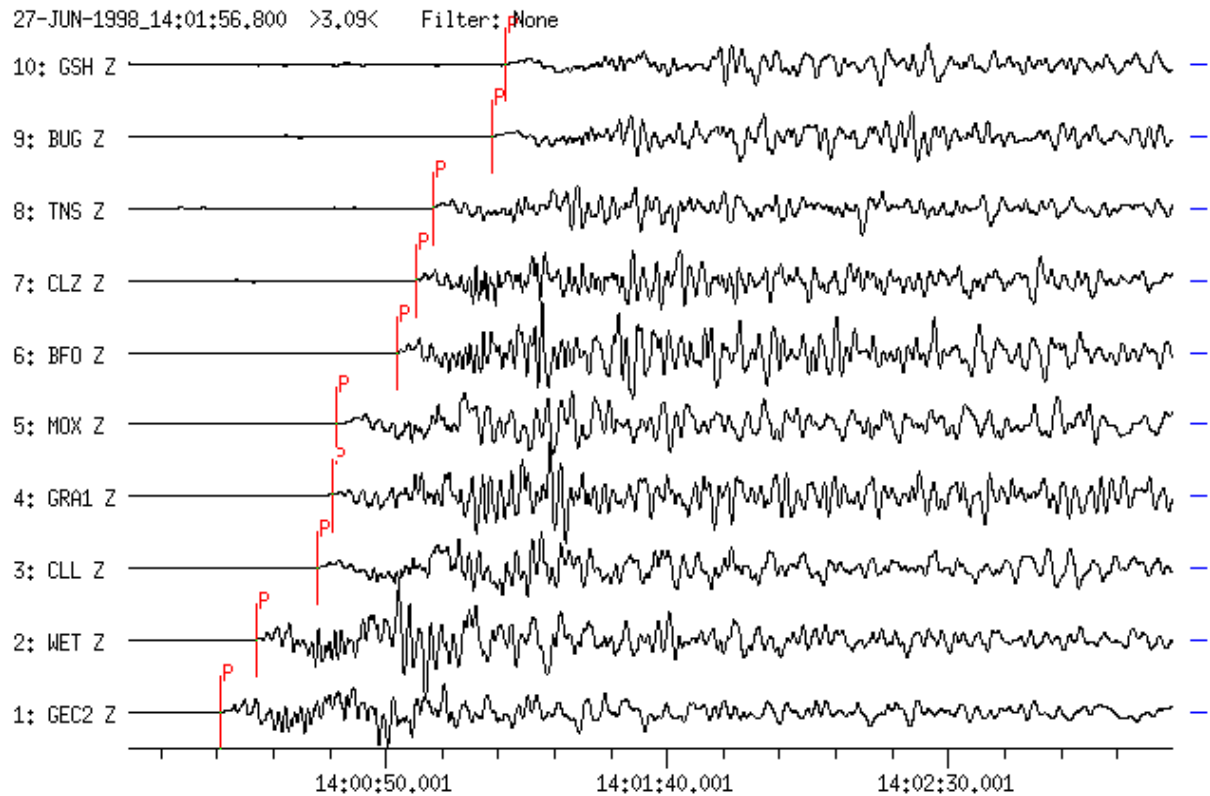
**Figure 2c** Broadband vertical-component seismogram with P-wave onsets recorded at all GRF-array stations. Traces are sorted according to increasing distance ( $D = 12.86^\circ$  from GRC3 and  $13.03^\circ$  from GRA1) and shifted in time according to a reference station (beamforming). All signal onsets are coherent. The weak first arrival of the P-wave onset is marked on the GRF-station GRA1.



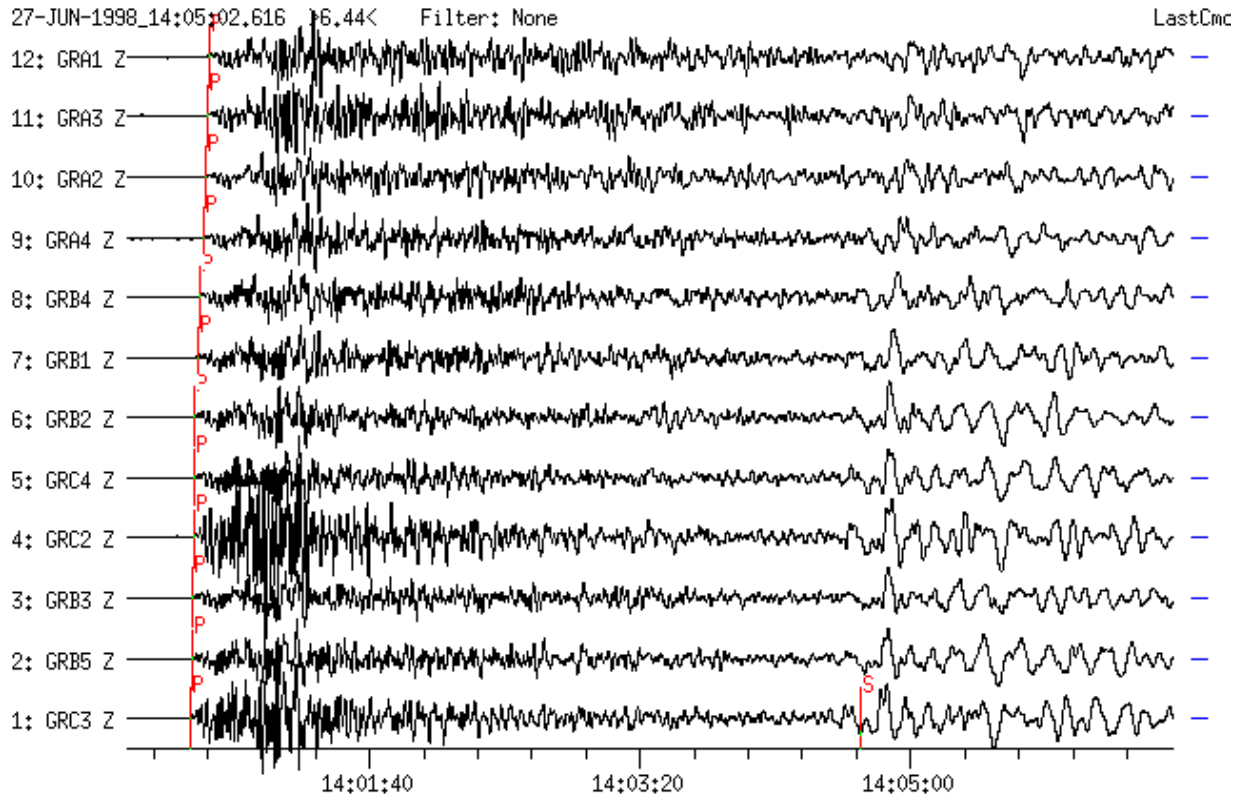
**Figure 2d** Broadband seismogram with P, weak S and long-period surface waves from station GRA1 (at  $D = 13.0^\circ$ ).

### Example 3: Earthquake in Southern TURKEY

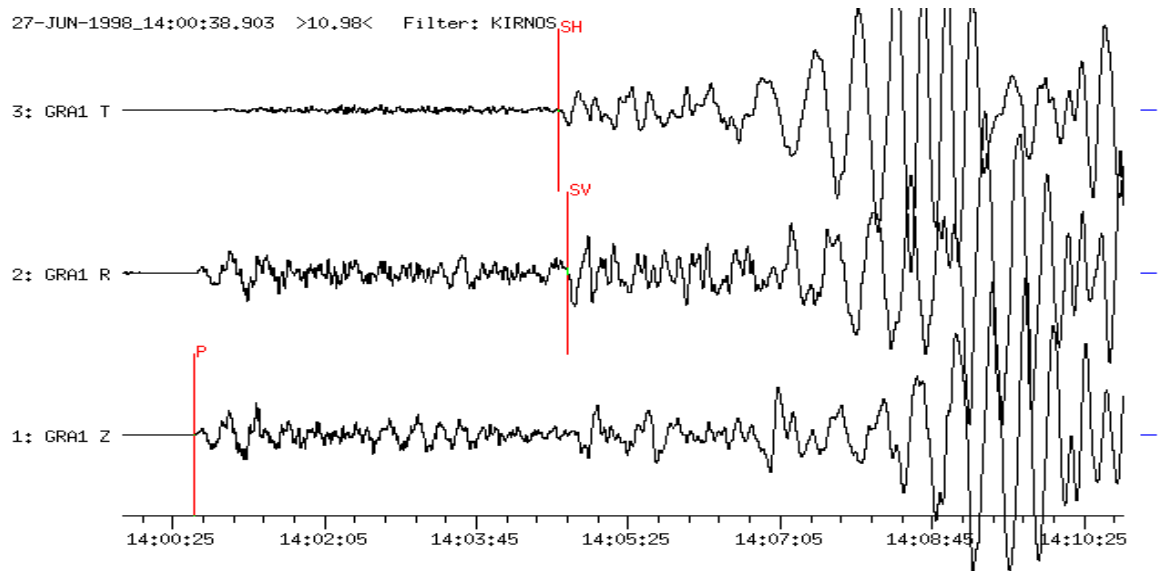
USGS QED-data: 1998-06-27 OT 13:55:49 36.95N 35.31E h = 10G Ms = 6.2  
(D = 21.6° to GRF)



**Figure 3a** Broadband seismograms with high time resolution showing the complex P-wave groups. Records were made on vertical components of 10 GRSN stations at epicentral distances between  $D = 19.7^\circ$  (GEC2) and  $24.8^\circ$  (GSH). Traces are sorted according to increasing distance. P waves on the individual traces are influenced by upper mantle discontinuities and signals are not coherent.



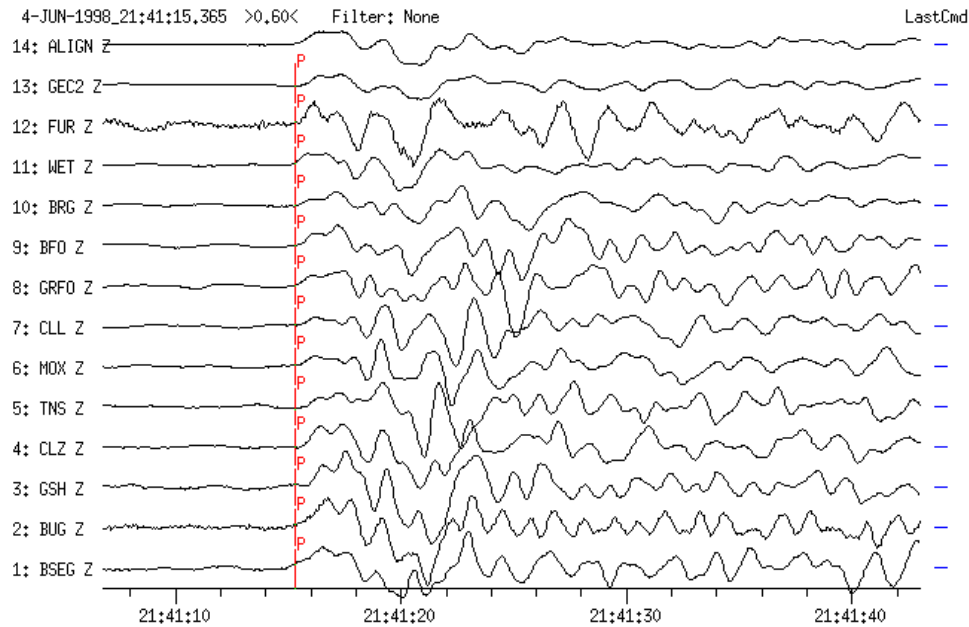
**Figure 3b** Broadband seismogram with P- and S-wave onsets recorded at 12 GRF-array stations. Traces are sorted according to increasing distance. P and S waves on the individual traces are influenced by upper mantle discontinuities and signals are also not coherent.



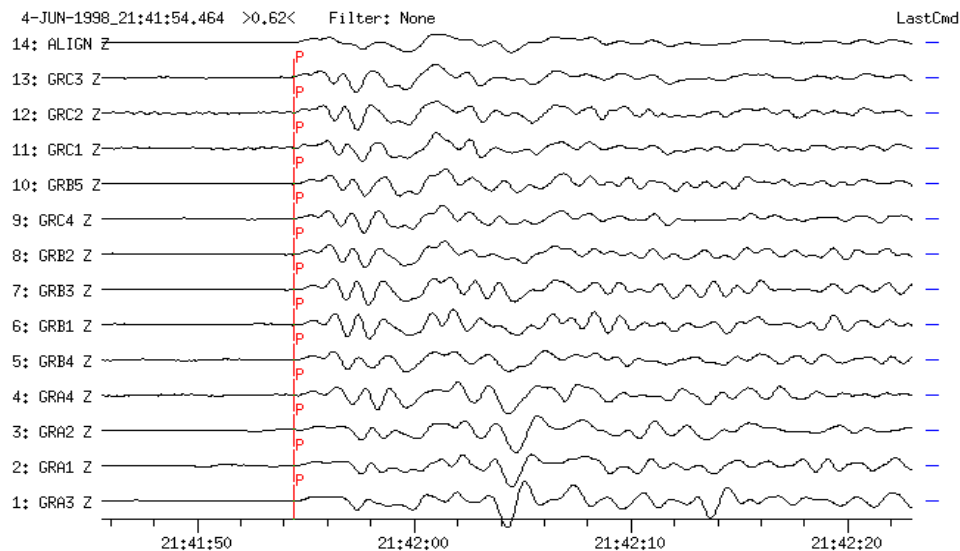
**Figure 3c** Three-component displacement-proportional KIRNOS-filtered seismogram recorded at the GRF-main-station GRA1 ( $D = 21.6^\circ$ ). The original N- and E- horizontal components have been rotated with R showing into the source direction. The time difference between the onsets SH (horizontal polarized S wave) and SV (vertical polarized S wave) is about 4 sec. The reason for this difference may be the anisotropy of upper mantle layers.

### Example 4: Earthquake in ICELAND REGION

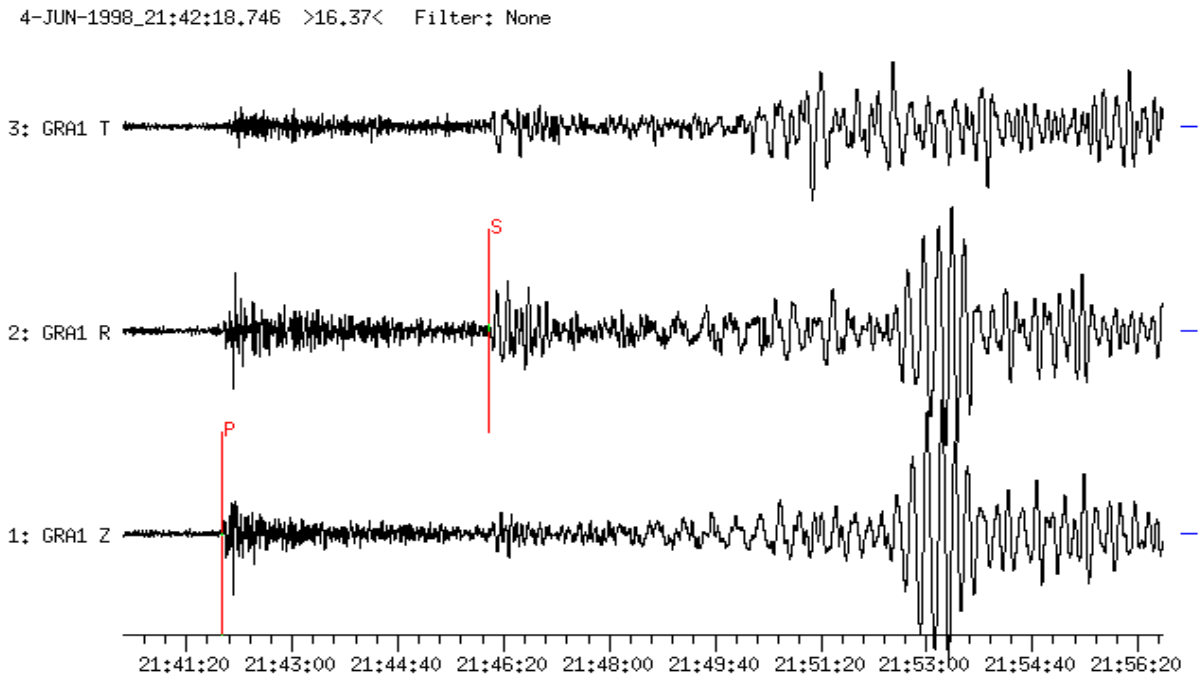
USGS QED-data: 1998-06-04 OT 21:36:54.2 64.009N 21.294W h = 10G  
mb = 5.1 Ms = 5.1 (D = 22.5° to GRF)



**Figure 4a** Broadband vertical-component seismograms recorded at 13 GRSN stations. Traces are sorted according to increasing epicentral distance (D = 18.9° to BSEG and 24.2° to GEC2), shifted in time and aligned for better signal comparison. All signals are incoherent and influenced by upper mantle discontinuities.



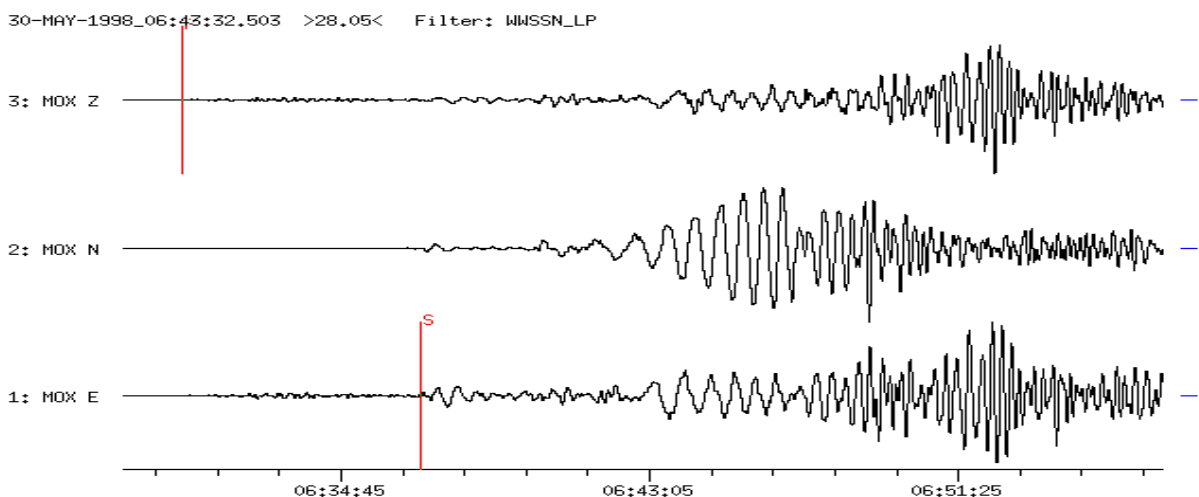
**Figure 4b** Broadband vertical-component seismograms recorded at the GRF-array stations. Traces are sorted according to increasing distance (D = 22.48° to GRA3 and 23.27° to GRC3), shifted in time and aligned for better signal comparison. Because of the smaller aperture of the array as compared to the GRSN network signals are more similar. At the nearest stations (GRA3 up to GRA4) a second onset appears about 10 s after P.



**Figure 4c** Simple three-component broadband seismogram at station GRA1 ( $D = 22.5^\circ$ ) with clear P, S and surface waves. Horizontal components have been rotated (ZRT) with R into source direction.

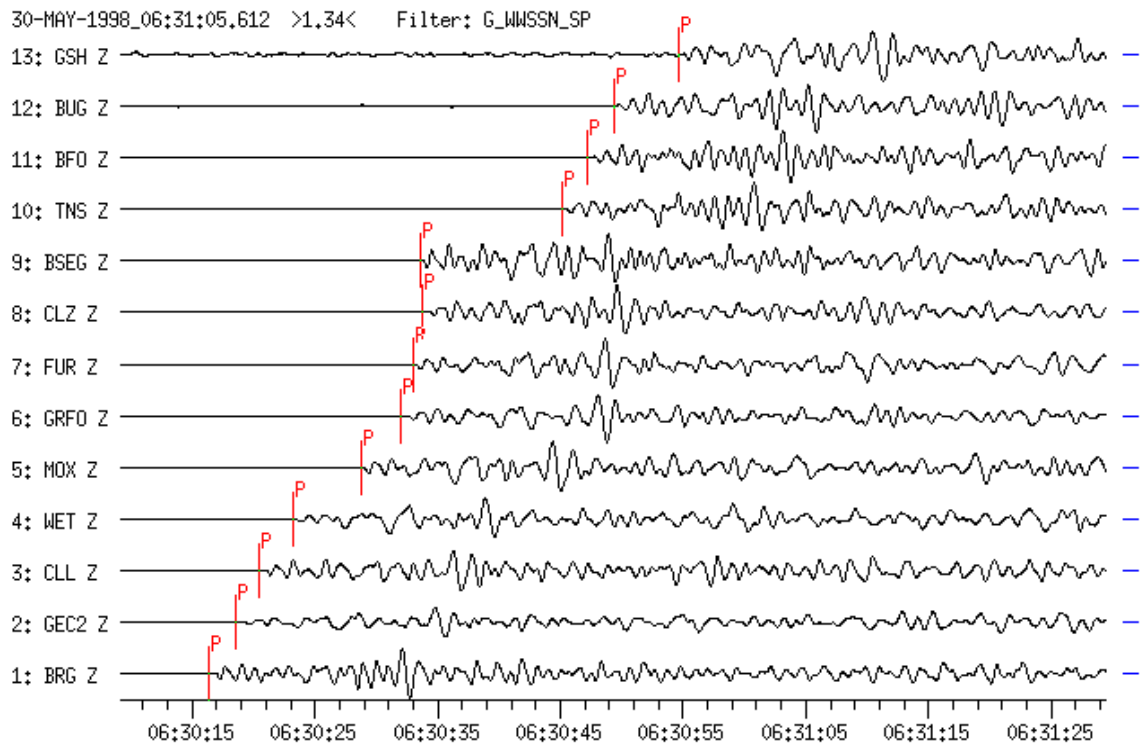
### Example 5: Earthquake at the Afghanistan -Tajikistan border region

USGS QED-data: 1998-05-30 OT 06:22:28.7 37.050N 70.086E h = 33N  
 mb = 5.8 Ms = 6.9 ( $D = 43.5^\circ$  and BAZ =  $83.7^\circ$  from GRF)



**Figure 5a** Three-component long-period seismogram (WWSSN-LP simulation filter) recorded at station MOX ( $D = 43.2^\circ$ , BAZ =  $85^\circ$ ) with P, S and dispersed surface waves. The nuclear explosion in Pakistan (see Figure 6.2) was recorded within the coda of this strong earthquake. As compared with the earthquake no surface waves has been recorded from the nuclear explosion.

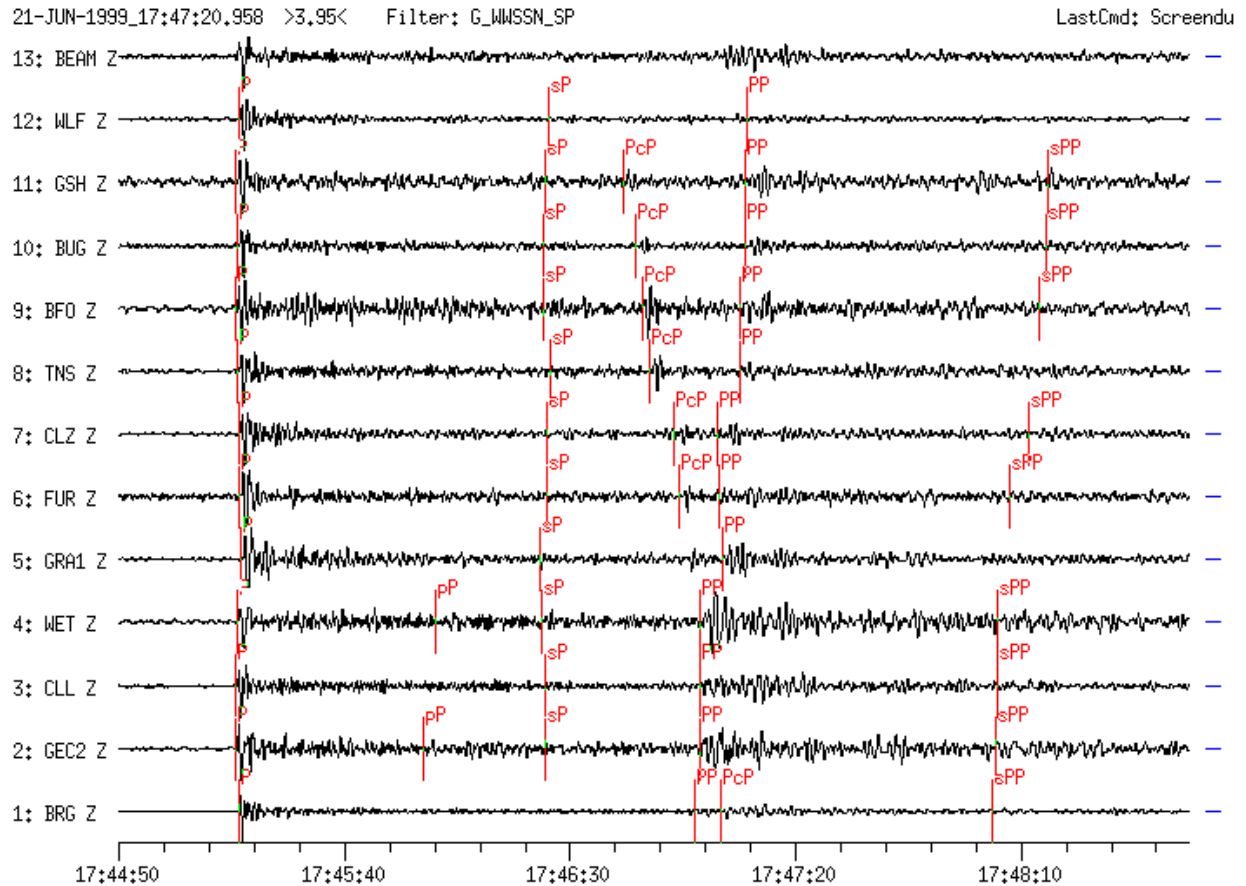




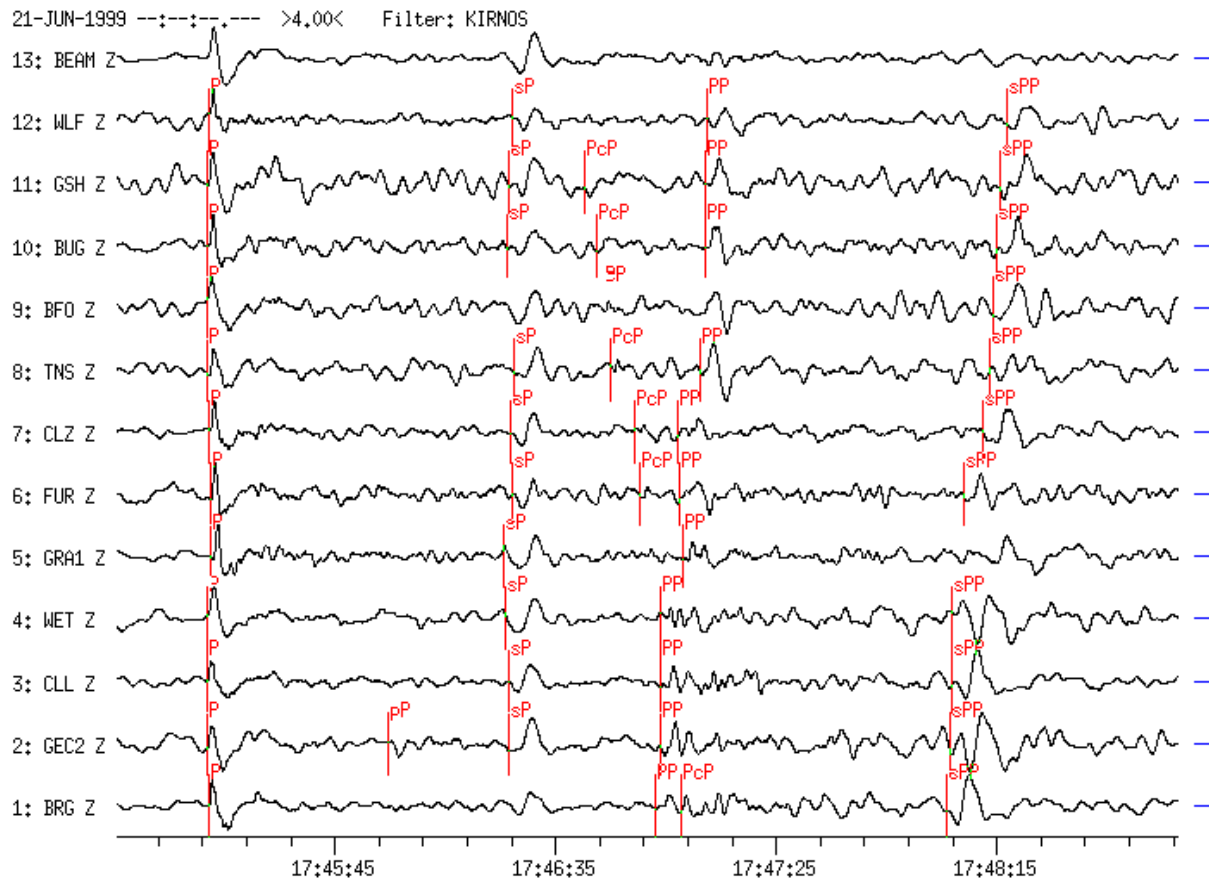
**Figure 5b** Short-period filtered P-wave onsets (WWSSN-SP simulation) recorded at 13 GRSN stations within the distance range between  $D = 41.7^\circ$  (BRG) and  $46.5^\circ$  (GSH). P-wave trains are rather complex as compared with the records of the nearby underground nuclear explosions (see DS 11.4, Figures 2 and 3).

### Example 6: Deep-focus earthquake in the HINDUKUSH REGION

USGS NEIC-data:1999-06-21 OT 17:37:29 36.40N 70.63E h = 249km  
mb = 5.7 (D = 44.3° to GRF, BAZ = 84.1 deg)



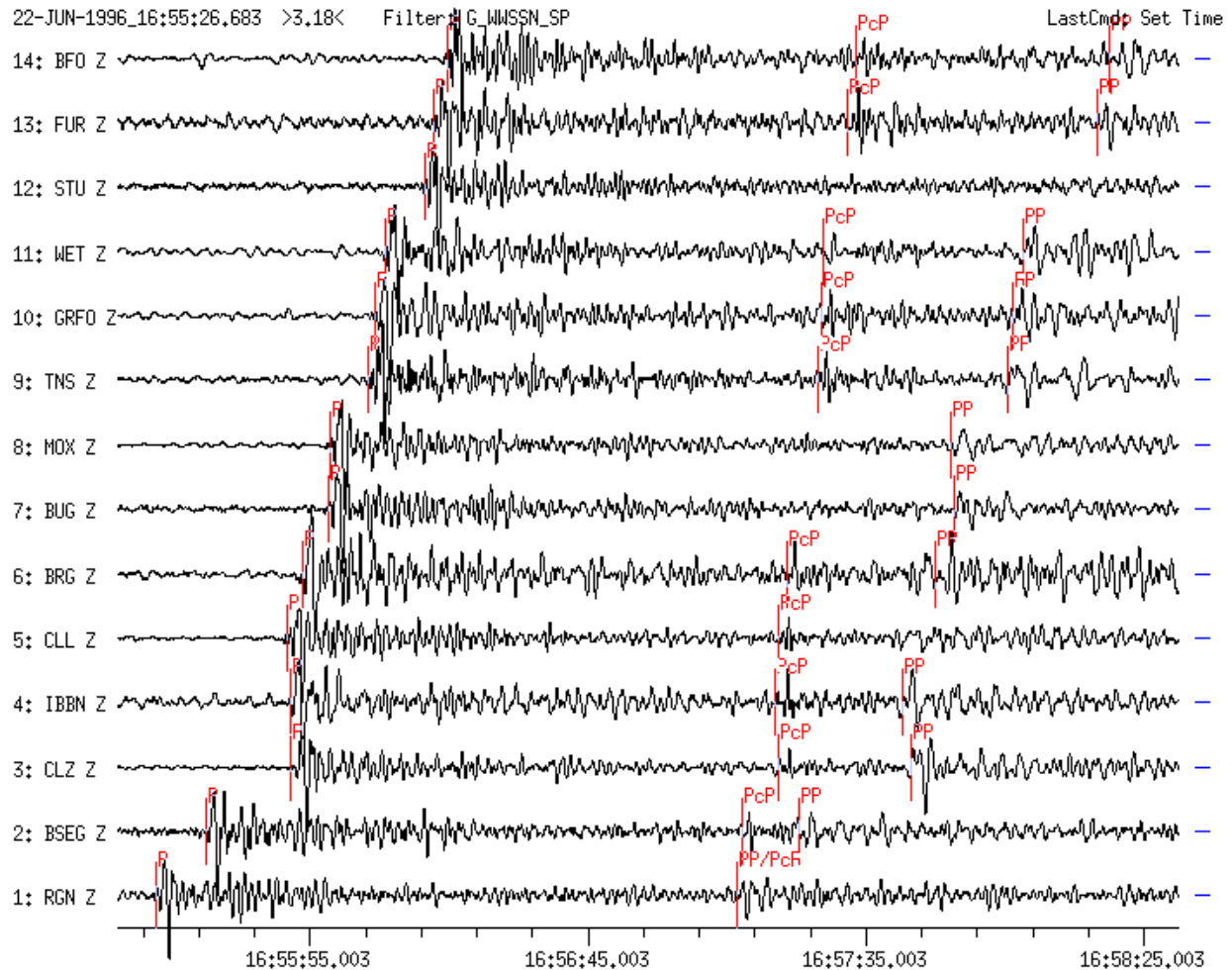
**Figure 6a** Short-period filtered seismogram (WWSSN-SP simulation) recorded at 12 GRSN, GRF(GRA1), GERESS and GEOFON stations. Traces are sorted according to distance (D = 42.4° to BRG and 47.6° to WLF), shifted and aligned for P onsets. Note that the travel-time curves for PcP and PP intersect in this distance range. Depth phases pP and sP (see the marked theoretically expected arrival times) are not visible on this short-period filtered record.



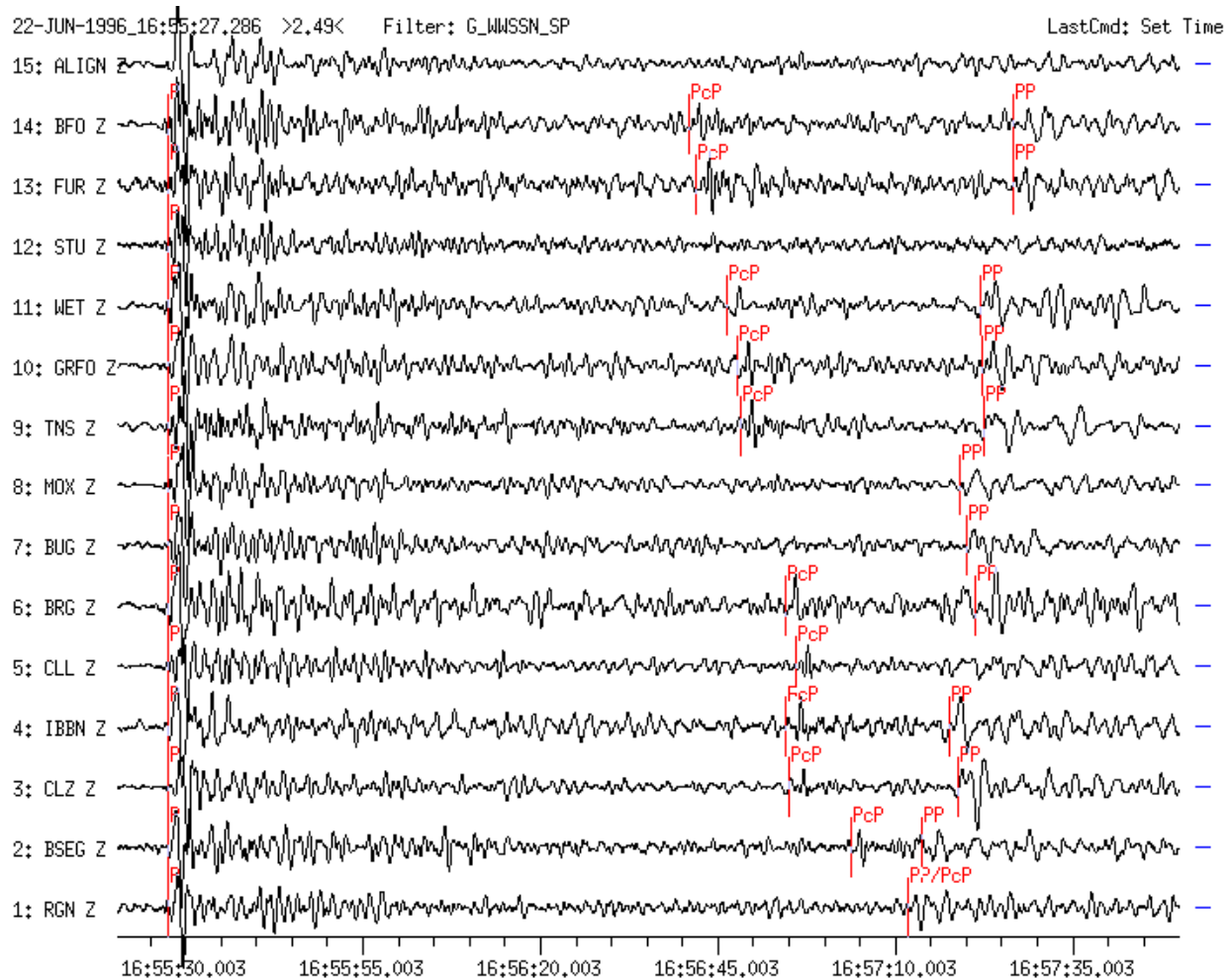
**Figure 6b** Stations and source parameter as in Figure 6a, however records of displacement-proportional Kirnos-simulation. Phases P, PP and the depth phases sP, sPP are clearly visible on these records while the phase pP is recognizable only at GEC2.

## Example 7: Earthquake in the Laptev Sea Region

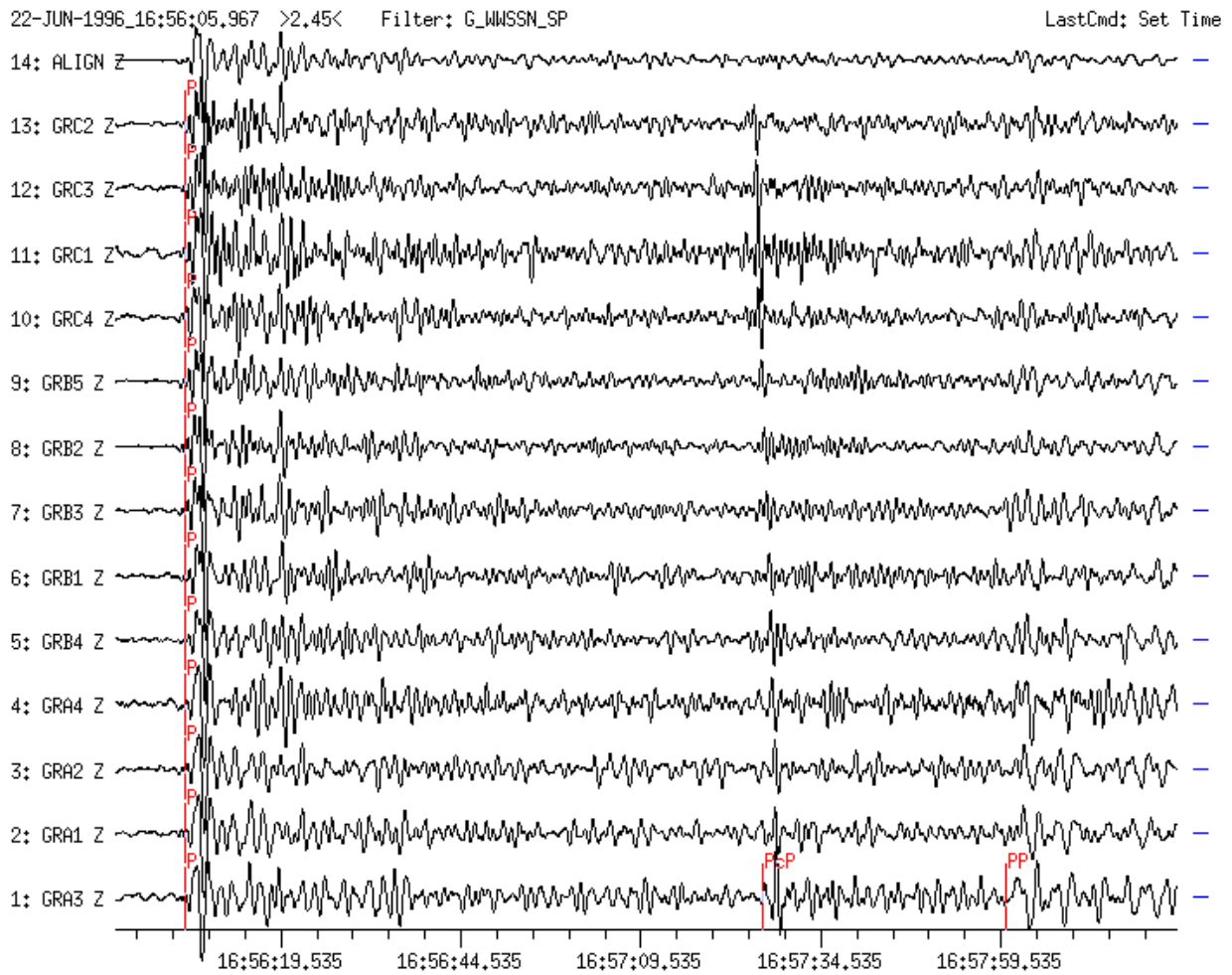
USGS NEIC-data: 1996-06-22 OT 16:47:13.1 75.812 N 134.710 E  $h = 10G$   
 $m_b = 5.6$   $M_s = 5.5$



**Figure 7a** Vertical (Z) component short-period seismograms (WWSSN-SP simulation filter) recorded at 14 GRSN stations. Traces are sorted according to increasing epicentral distance which ranges between  $D = 44.5^\circ$  for RGN (with  $BAZ = 17.5^\circ$ ) and  $51.4^\circ$  for BFO (with  $BAZ = 14.7^\circ$ ). P, PcP and PP are clearly visible. Note the decreasing travel-time difference (PcP - P) with increasing epicentral distance and the related small slowness values ( $sl < 4$  s/deg) for the core-reflected wave PcP. At a distance of about  $45^\circ$  (see record of station RGN) the travel-time curves of PcP and PP intersect. Generally, PcP is well recorded on short-period filtered records, however no PcP onset is recognizable above the noise level in the records of stations STU, MOX and BUG.



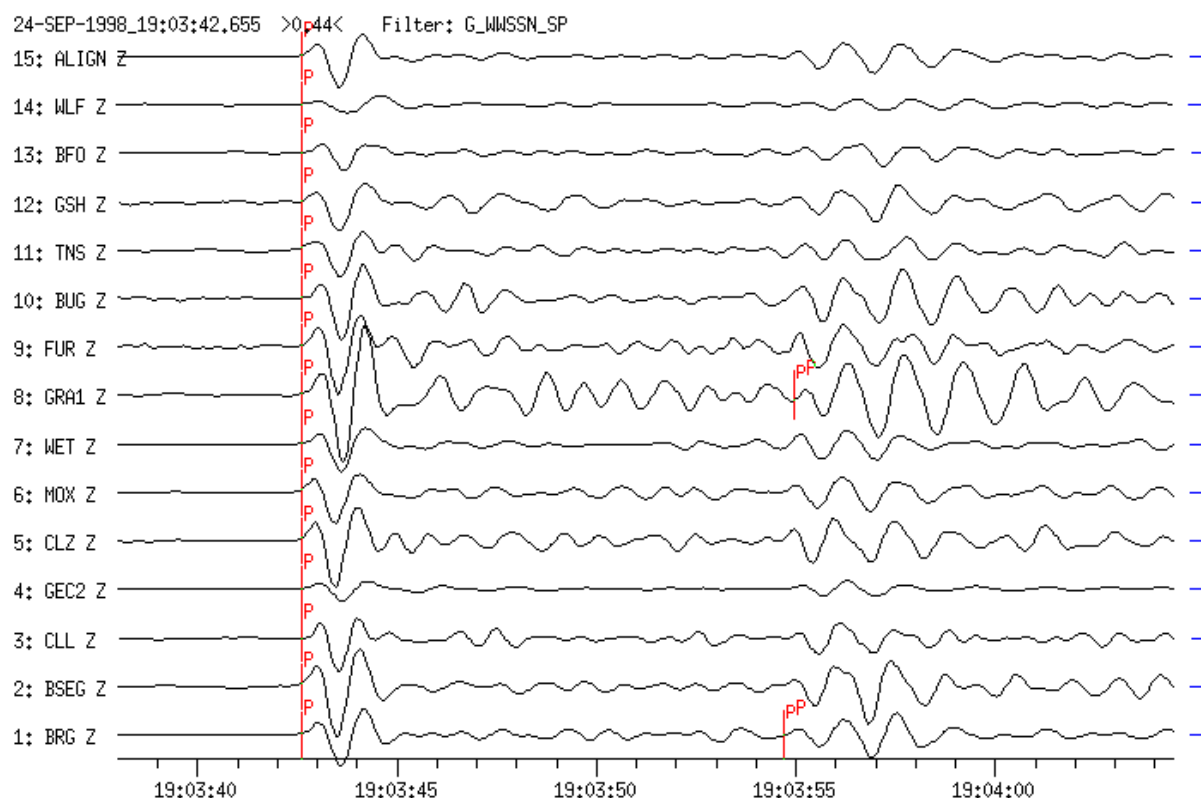
**Figure 7b** The same records as in Figure 7a, however, traces have been time-shifted and aligned for the P onsets. This figure shows more clearly the decreasing travel-time difference ( $P_{cP} - P$ ) with increasing distance from the epicenter.



**Figure 7c** Short-period filtered seismograms recorded at the GRF-array. Traces are sorted according to increasing distance ( $D = 49.48^\circ$  to GRA3 and  $50.33^\circ$  to GRC2), shifted in time and aligned for P onsets. Phases P, PcP and PP have been marked. Because of the smaller spacing of the array-stations as compared with the GRSN stations, the decrease of the travel-time difference PcP-P is less obvious.

# **Example 8: Record of an earthquake in Mongolia**

USGS-QED-data: 1998-09-24 OT 18:53:40.2 46.274 N 106.237 E h = 33km  
mb = 5.3 Ms = 5.4, (D = 59.4° and BAZ = 54° from GRF(GRA1))



**Figure 8a** This is an example for an event with coherent short-period P waves within the GRSN (WWSSN-SP simulation filter). Additionally, the record of the GEOFON-station WLF in Luxembourg is shown on trace No.14. The traces are sorted according to increasing distance (D = 57.3° to BRG and 62.1° to WLF), shifted and aligned for P onsets. The most remarkable feature is the strong variability of the P-wave signal-amplitudes within this regional network. The table below gives the measured amplitudes together with the calculated magnitudes. Body-wave magnitudes mb vary between 5.4 (GEC2) and 6.2 (GRA1). The depth phase pP was used to estimate a better source depth (h = 44 km) than the value given in the QED (h = 33 km).

**Table 8a** Result of the seismogram analysis shown in Figure 8a. The table was printed from the GRSN databank and contains in the uppermost line source parameters like date, an event identification number (ev\_id) and the name of the analyst (KLI for Klinge). The following lines give the analyzed stations, onset times, phases, polarities, components, periods T, maximum amplitudes A, body- and/or surface-wave magnitudes mb/MS, epicentral distances D, beam-slowness b\_slo, beam\_azimuths b\_az. The lowermost part of the table contains source parameters like analysis center (SZGRF), origin time OT, latitude, longitude, average magnitude values for mb and MS, depth of the source and source region. Note the significant differences between the magnitude estimates mb from records of different network stations.

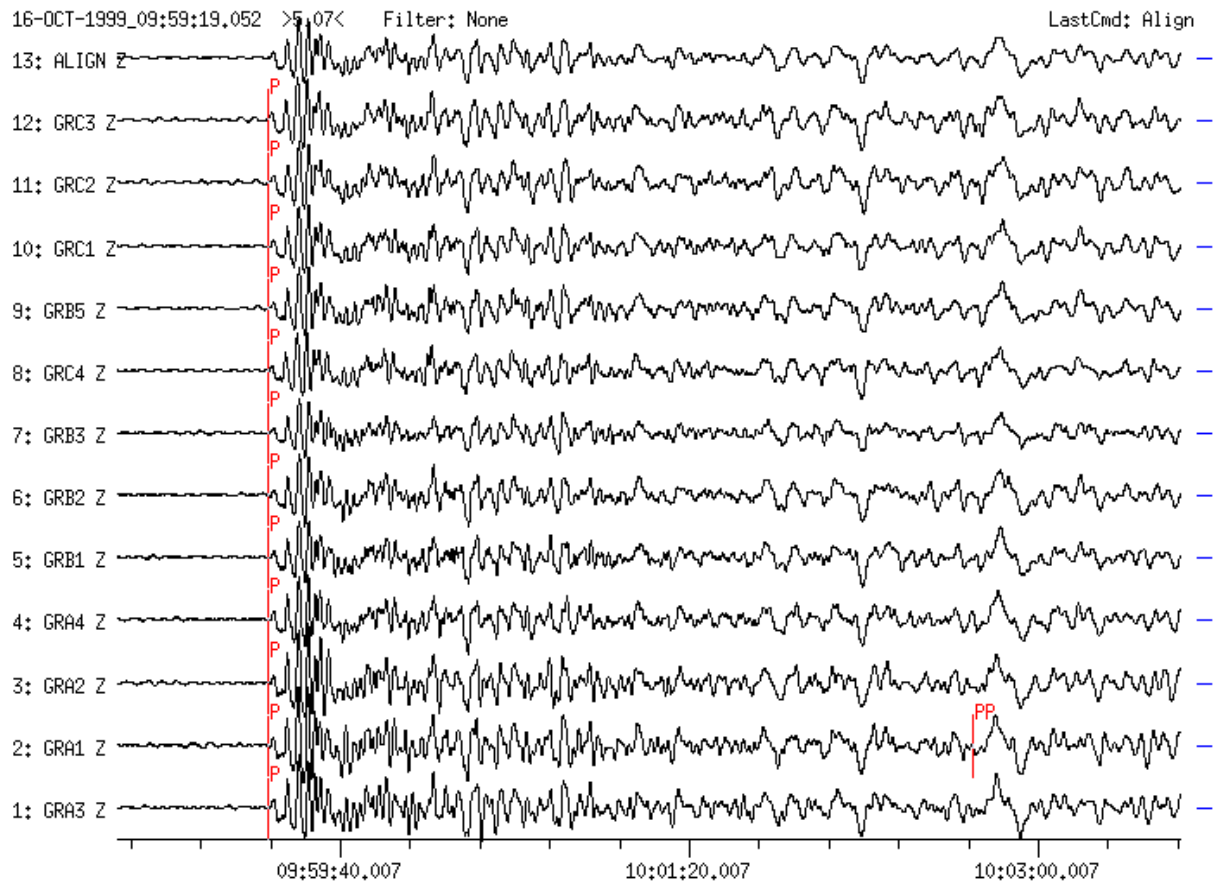
| 1998-09-24          |               |        |          |        |           |           |        |  |  |  |  |  |  |
|---------------------|---------------|--------|----------|--------|-----------|-----------|--------|--|--|--|--|--|--|
| ev_id 980924007 KLI |               |        |          |        |           |           |        |  |  |  |  |  |  |
| BRG                 | 19:03:27.2    | e P    | Z        | T 1.2  | A 135.5   | mb 5.9    | D 58.3 |  |  |  |  |  |  |
| ALIGN               | 19:03:27.2    | e P    | Z        | T 1.1  | A 124.1   | mb 5.8    |        |  |  |  |  |  |  |
| BSEG                | 19:03:28.2    | e P    | Z        | T 1.1  | A 198.3   | mb 6.0    | D 58.4 |  |  |  |  |  |  |
| CLL                 | 19:03:28.6    | e P    | Z        | T 0.9  | A 98.9    | mb 5.8    | D 58.5 |  |  |  |  |  |  |
| GEC2                | 19:03:36.3    | i P    | c Z      | T 1.2  | A 46.9    | mb 5.4    | D 59.6 |  |  |  |  |  |  |
| CLZ                 | 19:03:36.6    | e P    | Z        | T 1.1  | A 177.7   | mb 6.0    | D 59.6 |  |  |  |  |  |  |
| MOX                 | 19:03:36.9    | e P    | Z        | T 1.2  | A 122.4   | mb 5.8    | D 59.6 |  |  |  |  |  |  |
| WET                 | 19:03:38.5    | e P    | Z        | T 1.2  | A 107.2   | mb 5.7    | D 59.9 |  |  |  |  |  |  |
| BRG                 | 19:03:39.3    | e pP   | Z        |        |           |           |        |  |  |  |  |  |  |
| GRA1                | 19:03:42.6    | i P    | c Z      | T 1.1  | A 286.5   | mb 6.2    | D 60.4 |  |  |  |  |  |  |
|                     |               |        |          |        | b_slo 6.8 | b_az 54   |        |  |  |  |  |  |  |
| BUG                 | 19:03:48.5    | e P    | Z        | T 1.1  | A 162.4   | mb 5.8    | D 61.4 |  |  |  |  |  |  |
| FUR                 | 19:03:48.6    | e P    | Z        | T 1.1  | A 174.3   | mb 5.8    | D 61.3 |  |  |  |  |  |  |
| TNS                 | 19:03:49.9    | e P    | Z        | T 1.1  | A 103.0   | mb 6.0    | D 61.5 |  |  |  |  |  |  |
| GSH                 | 19:03:54.5    | e P    | Z        | T 1.3  | A 132.3   | mb 6.0    | D 62.3 |  |  |  |  |  |  |
| GRA1                | 19:03:55.0    | e pP   | Z        |        |           |           |        |  |  |  |  |  |  |
| BFO                 | 19:03:57.4    | e P    | Z        | T 1.1  | A 55.2    | mb 5.6    | D 62.8 |  |  |  |  |  |  |
| WLF                 | 19:04:00.1    | e P    | Z        | T 1.7  | A 80.3    | mb 5.6    | D 63.0 |  |  |  |  |  |  |
| GRA1                | 19:11:52.2    | e S    | E        |        |           |           | D 60.4 |  |  |  |  |  |  |
| GEC2                | 19:30:36.0    | e L    | Z        | T 19.9 | A 3895.8  | MS 5.5    |        |  |  |  |  |  |  |
| GRA1                | 19:31:03.6    | e L    | Z        | T 20.6 | A 3398.7  | MS 5.5    |        |  |  |  |  |  |  |
| SZGRF               | OT 18:53:39.3 | 45.30N | 106.84E  |        | mb_av 5.8 | MS_av 5.5 |        |  |  |  |  |  |  |
| DEP                 | 44km          | ▲      | MONGOLIA |        |           |           |        |  |  |  |  |  |  |



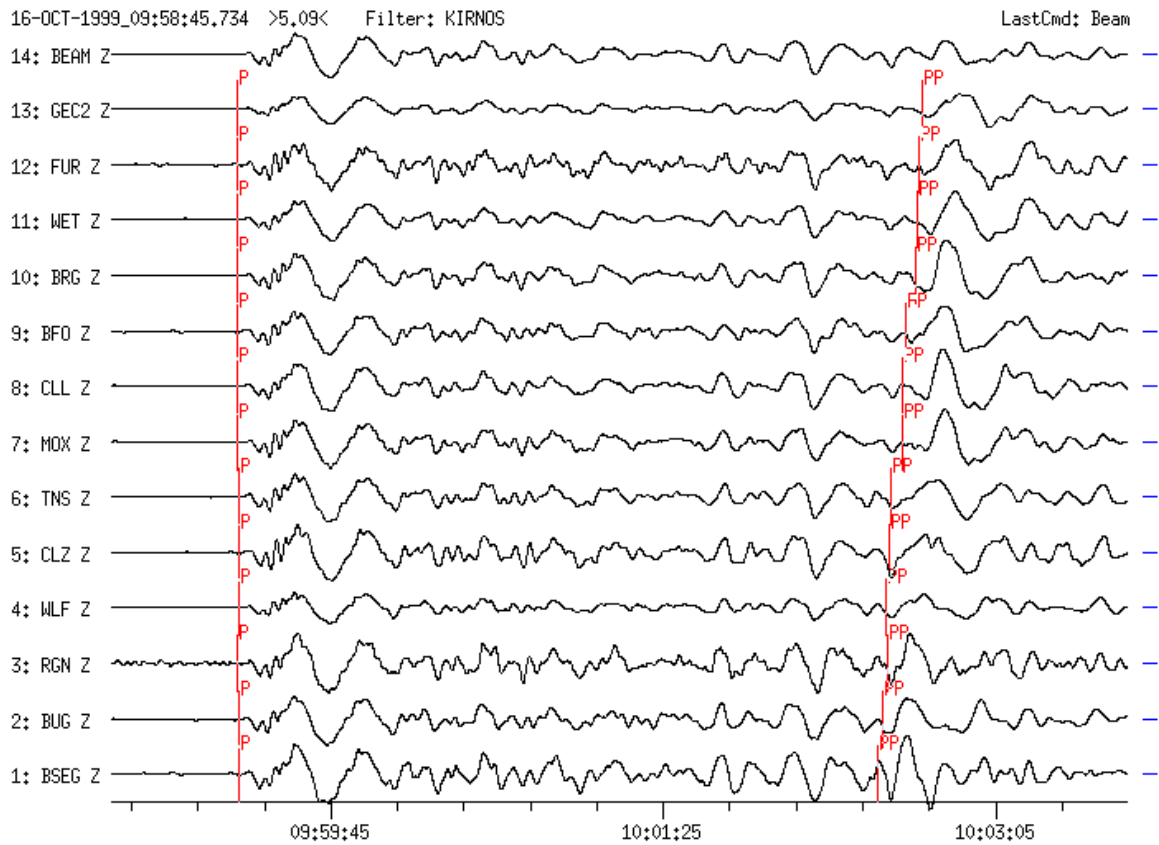
### Example 9: Earthquake in California

SZGRF-data:1999-10-16 OT 09:46:55 34.9N 115.9 W mb = 6.6 Ms = 7.9

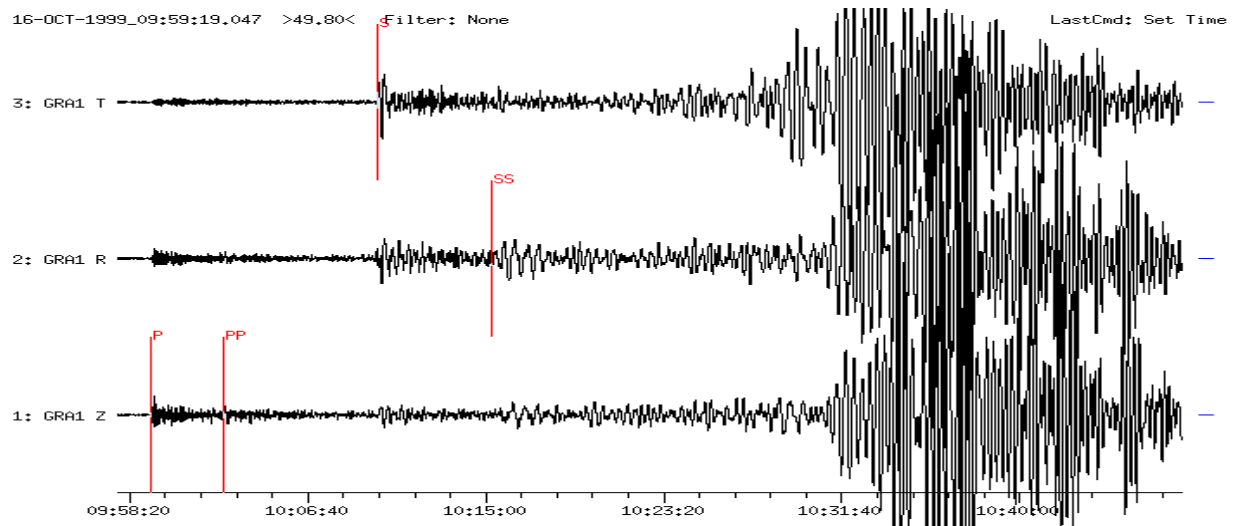
The event happened east of Los Angeles (D = 83.6° BAZ = 319° from GRF) .



**Figure 9a** Broadband vertical-component seismograms recorded at 12 GRF-array stations. Traces are sorted according to increasing distance (D = 83.6° to GRA3 and 84.4° to GRC3), shifted in time and aligned for P onsets. The coherent phases P and PP are marked, however some more onsets appear ahead of PP.



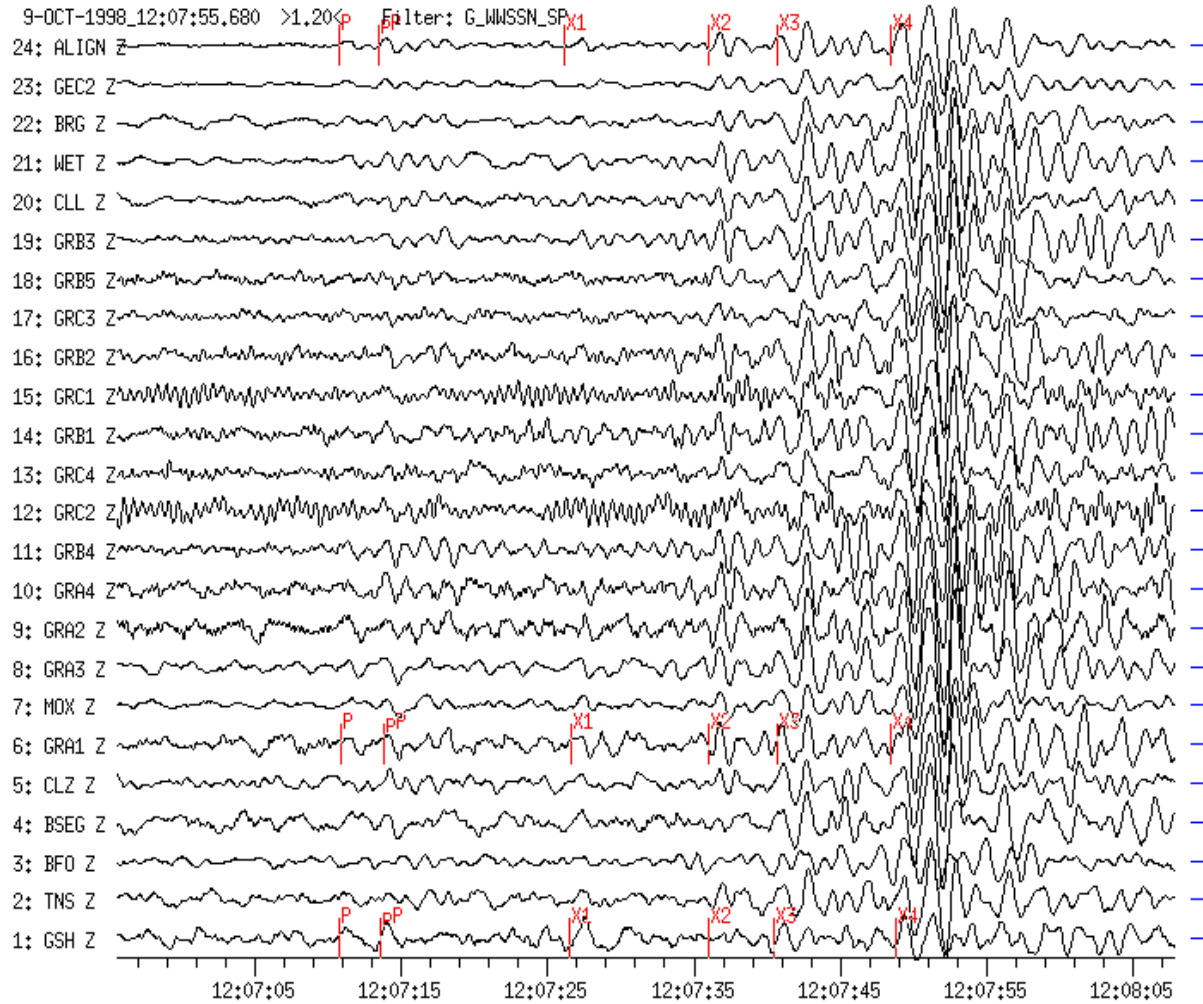
**Figure 9b** Broadband vertical-component seismograms recorded at 13 GRSN-stations. Traces are filtered (Kirnos-simulation), sorted according to increasing distance ( $80.1^\circ$  to BSEG and  $85.3^\circ$  to GEC2), shifted in time and aligned for P onsets. The used displacement proportional broad-band filter displays clearly the coherent part of the wave train from the network. As in Figure 9a, coherent phases appear ahead of PP that have slowness values as P.



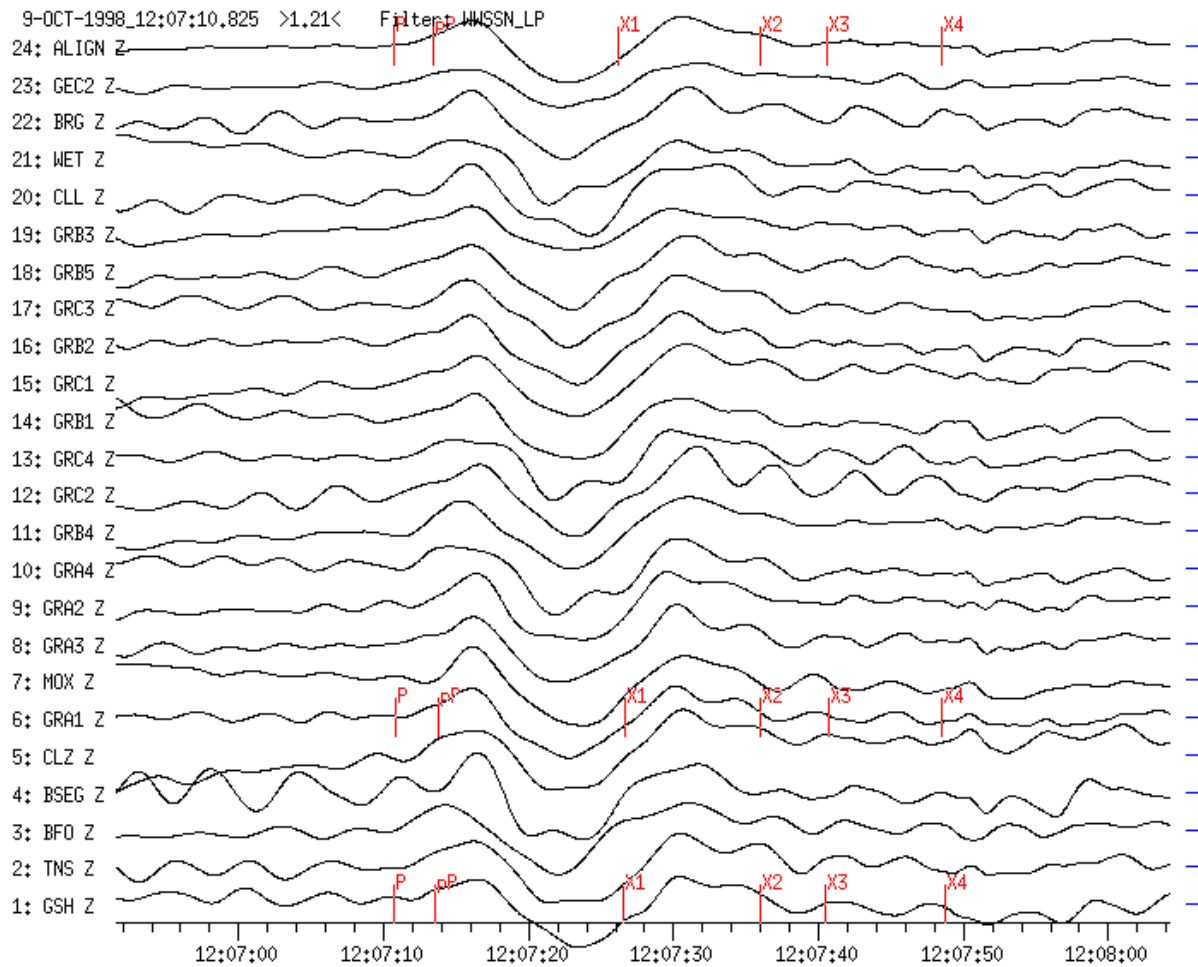
**Figure 9c** Broadband three-component seismogram recorded at GRA1 ( $D = 83.6^\circ$ ). The horizontal components are rotated into the R and T direction. Phases P, PP, S, SS and surface waves are displayed. Rayleigh waves recorded on radial and vertical components appear later than Love waves recorded on the transversal component only.

# **Example 10: Earthquake near the coast of NICARAGUA**

USGS NEIC-data: 1998-10-09 OT 11:54:29.0 11.337N 86.429W h = 10km  
mb = 5.6 Ms = 5.6 (D = 86.3° and BAZ = 282° from GRF(GRA1))



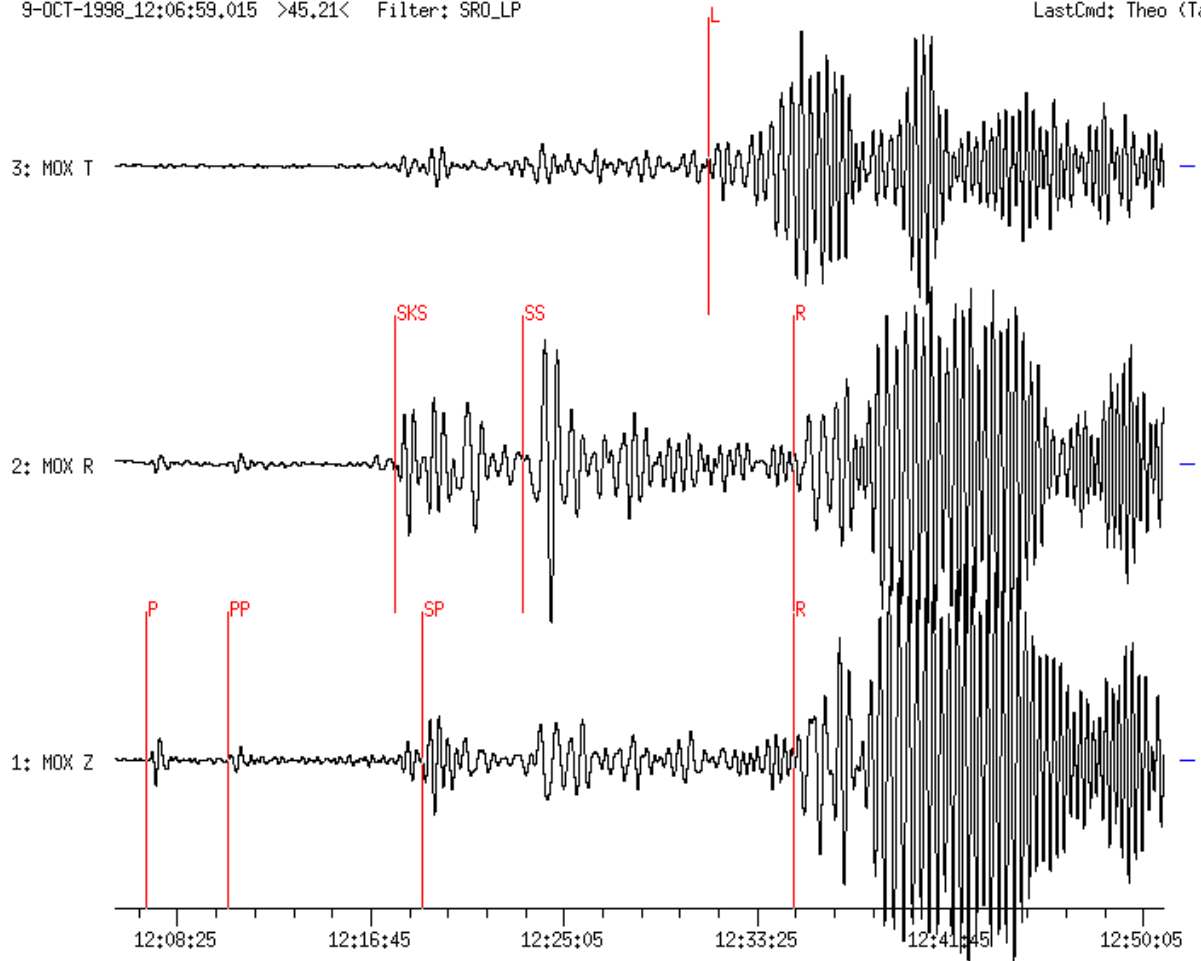
**Figure 10a** Short-period seismograms (WWSSN-SP simulation filter) from 23 broadband stations of the GRF-array, the GRSN network and the GERESS array (GEC2). Traces are time-shifted, aligned for P and sorted according to increasing distance (D = 83.1° to GSH and 88.1° to GEC2). Note the large number of onsets within the first 40 sec of the P-wave group (P, pP, X1, X2, X3, X4). The reason for these multiple onsets may be a multiple rupture process or (in some cases) reflections from the nearby subduction zone.



**Figure 10b** Long-period seismograms (WWSSN-LP simulation filter) of the same stations as in Figure 10a, depicted with high time-resolution. Traces are aligned and sorted according to increasing distance. Note the very long-period P-wave onset ( $T \approx 14$  s) in the time window between P and X2 where the P-wave amplitudes are relatively small in the short-period filtered records whereas at X4, which is by far the largest onset in Figure 10a, no long-period wave onset is to be seen. This seems to speak of an initially “slow” earthquake rupture which then escalated into a faster rupture or the break of a “harder” asperity which generated more short-period energy.

9-OCT-1998\_12:06:59.015 >45.21< Filter: SRO\_LP

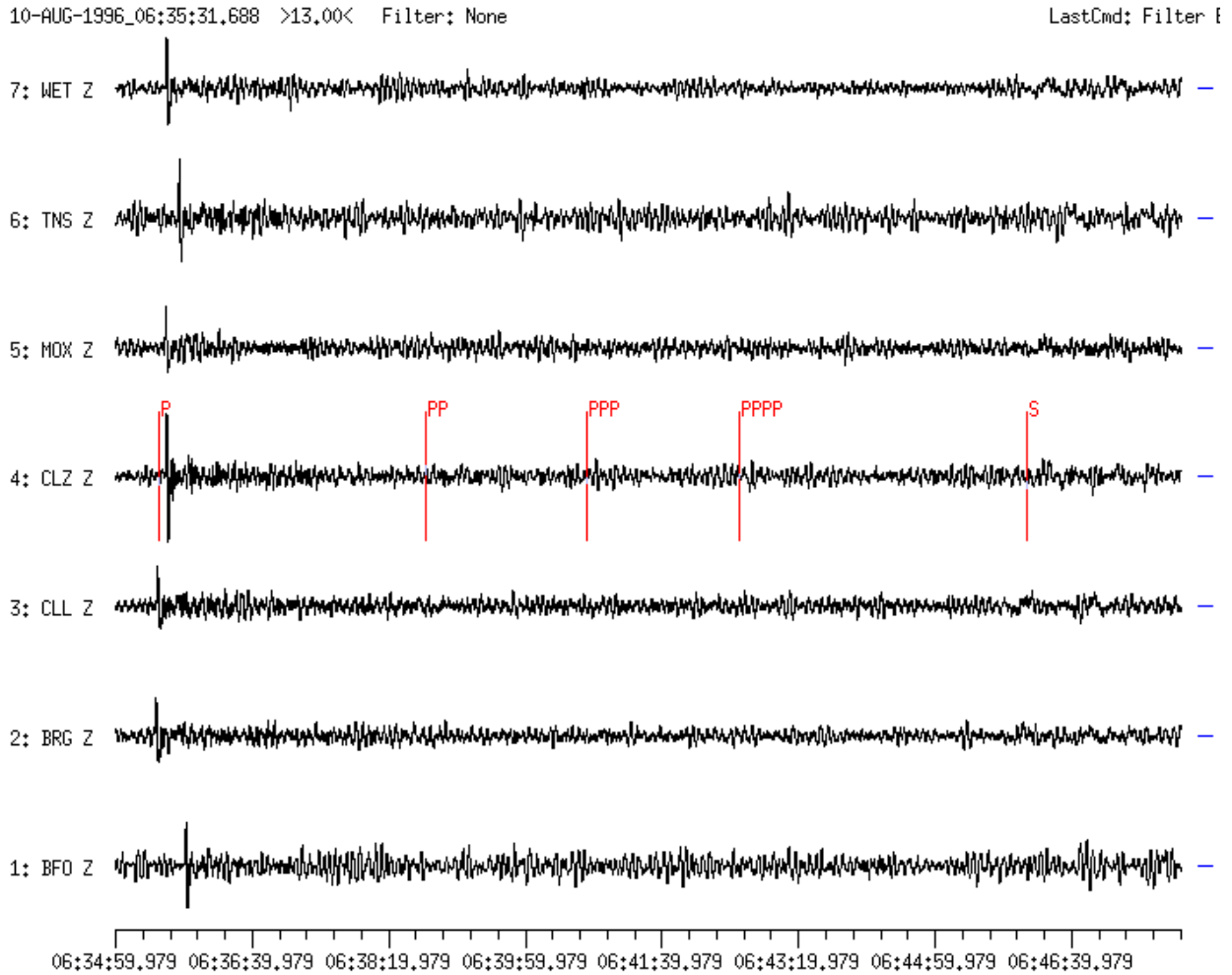
LastCmd: Theo (Table



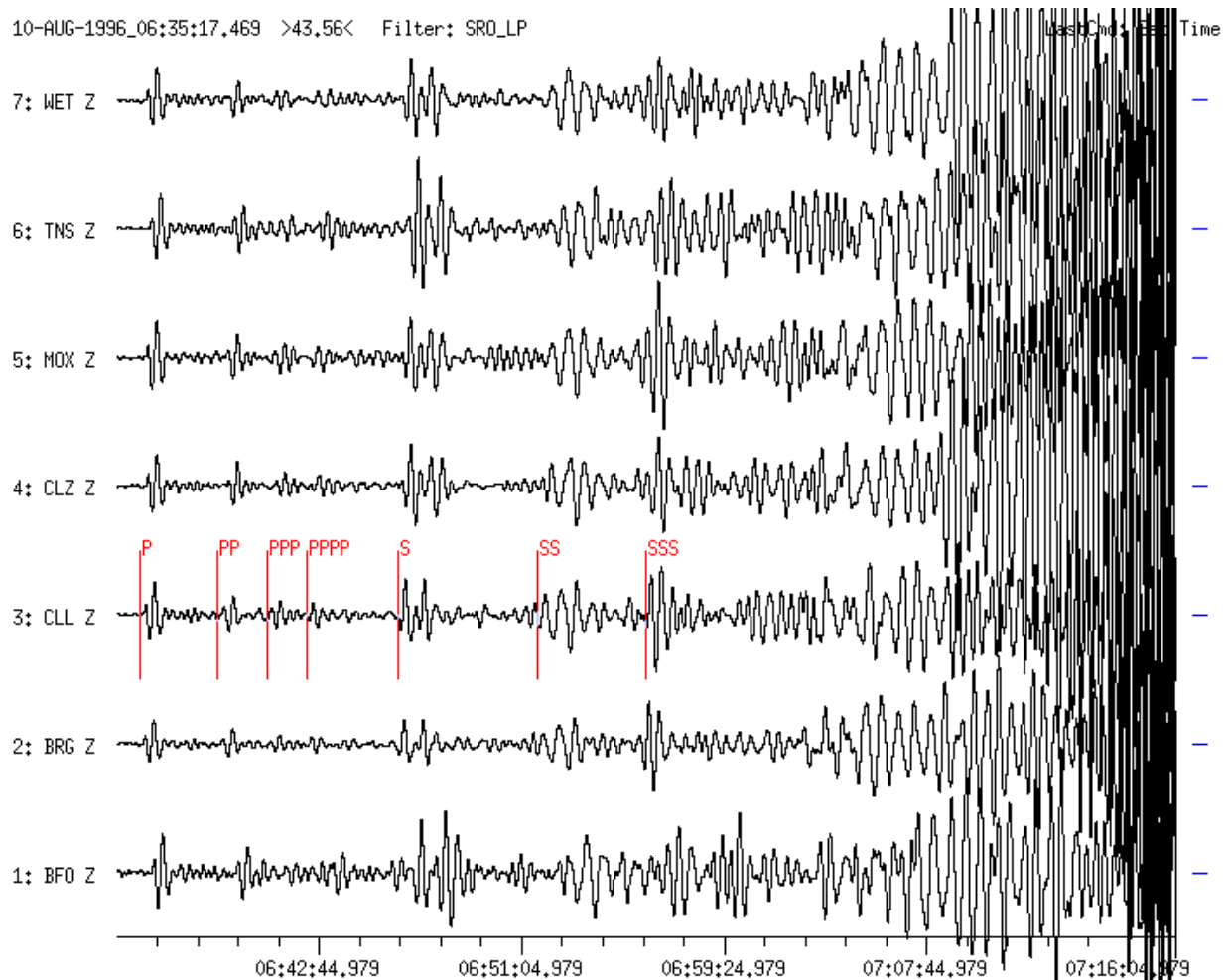
**Figure 10c** Time-compressed long-period filtered three-component seismogram (SRO-LP simulation filter) of the Nicaragua earthquake recorded at station MOX ( $D = 86.4^\circ$ ,  $BAZ = 283^\circ$ ). Horizontal components have been rotated into the R (radial) and T (transverse) direction. The seismogram shows long-period phases P, PP, SKS, SP, SS and surface waves L (or LQ for Love wave) and R (or LR for Rayleigh wave). Note the remarkably simple waveforms of P and PP as compared with the complicated structure in the short-period P-wave group in Figure 10a.

### Example 11: Earthquake in Taiwan Region

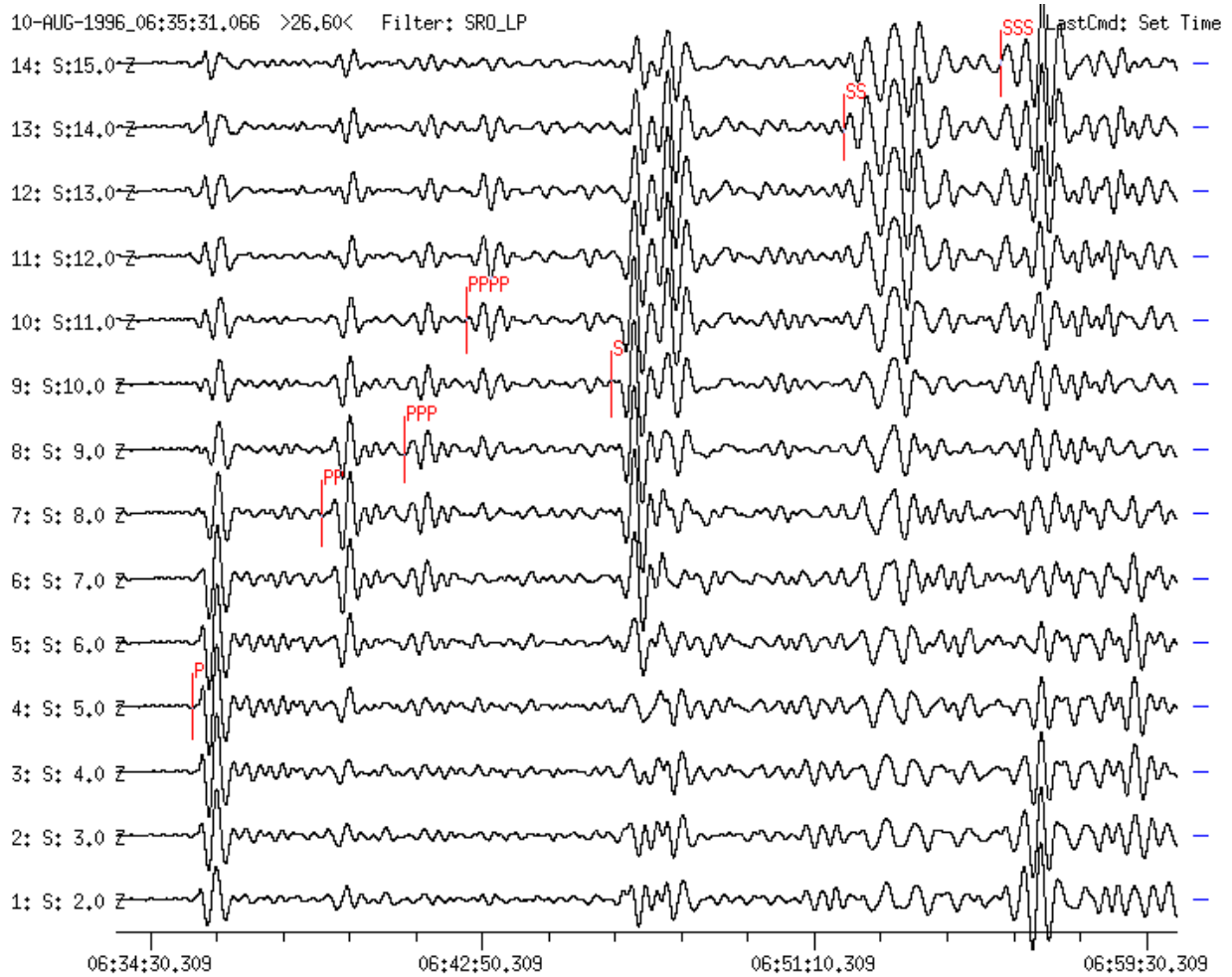
USGS NEIC-data: 1996-08-10 OT 06:23:08 24.032 N 122.550 E h = 46G  
mb = 5.2 Ms = 5.2 ( D = 87.1° and BAZ = 56.6° from BFO)



**Figure 11a** Broadband vertical-component seismograms of the Taiwan earthquake recorded at 7 GRSN-stations within the distance range  $D = 82.9^\circ$  and  $87.1^\circ$ . Only the P-wave onset is recognizable on the records. Secondary onsets such as surface reflections of the P and S wave (theoretical onset times marked on the CLZ trace) are not to be seen above the noise level.



**Figure 11b** The same event as in Figure 11a after long-period filtering (SRO-LP simulation filter). Note the pronounced onsets of PP, PPP, PPPP as well as S, SS and SSS which are marked at the record trace of station CLL ( $D = 82.9^\circ$ ). Phase identification is eased by using absolute (see overlay to Fig. 2.48) or differential theoretical travel-time curves (see Figure 4 in Exercise EX 11.2). This is essential when interpreting single station recordings. Modern seismogram analysis program can automatically mark the theoretically expected onset times of the various phases when the epicentral distance to the station and the source depth are known or assumed. When a network of stations or array is available, phase identification is supported by *vespagram* analysis as shown in the following figure.

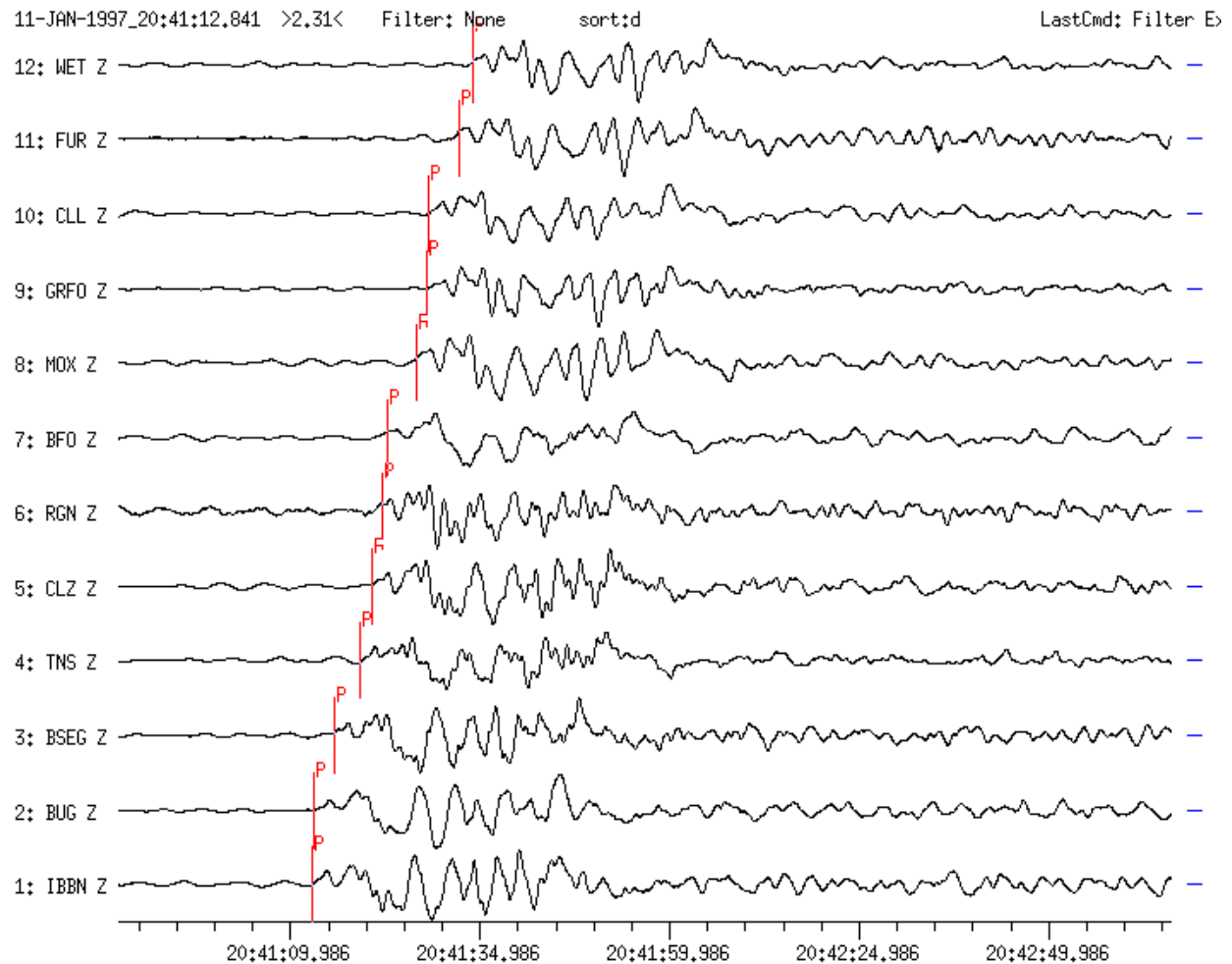


**Figure 11c** For better phase identification a [vespagram analysis](#) (see 9.7.7 in Chapter 9) of the long-period filtered traces was made. Identified phases have been marked on their respective slowness trace S (e.g., P on trace 4 with slowness  $S = 5.0 \text{ s/}^\circ$  and PP on trace 7 with  $S = 8.0 \text{ s/}^\circ$ ). In the vespagram traces the respective phases have the largest amplitudes at their proper slowness value.

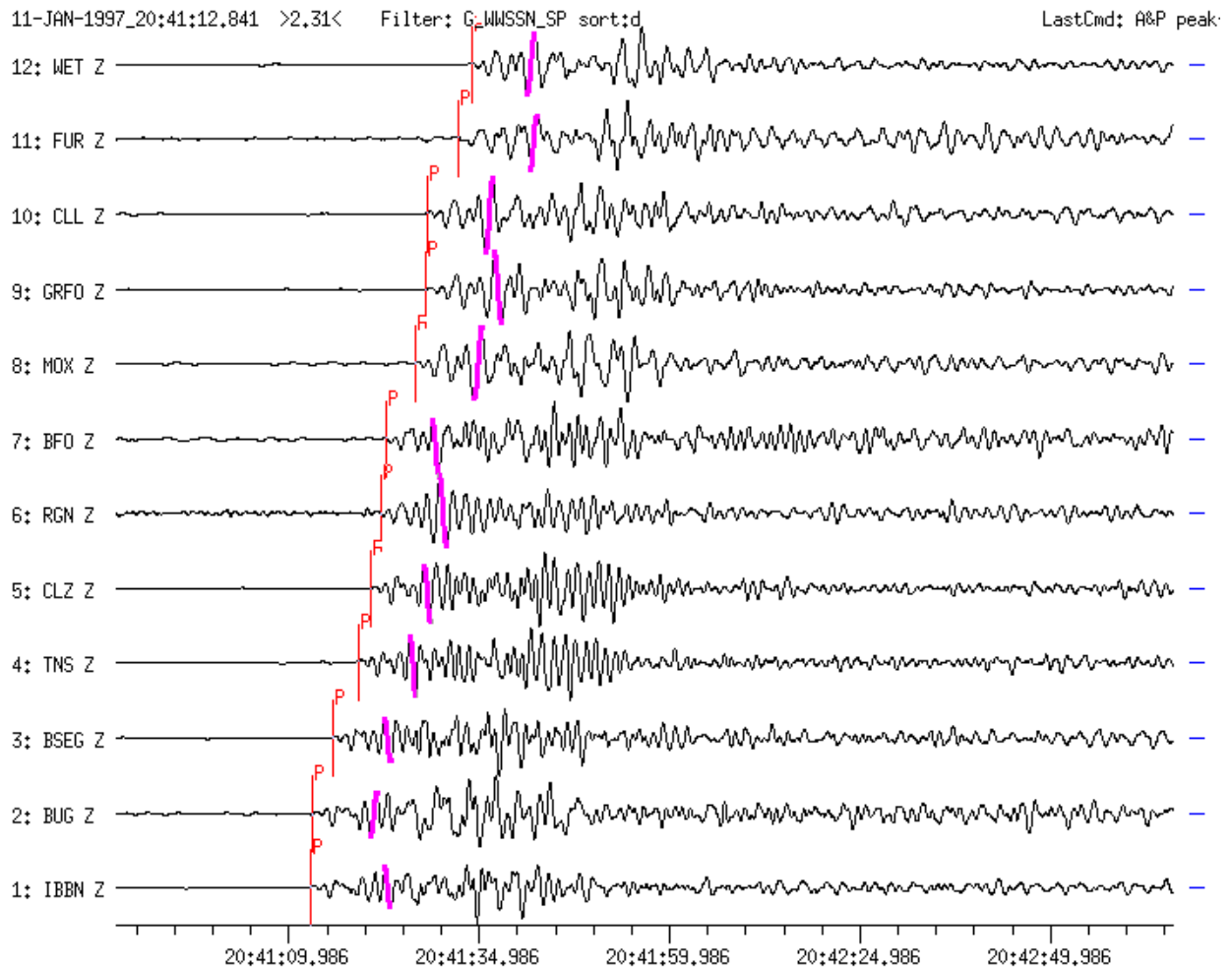


**Example 12:** Earthquake in the region of Michoacan, Mexico

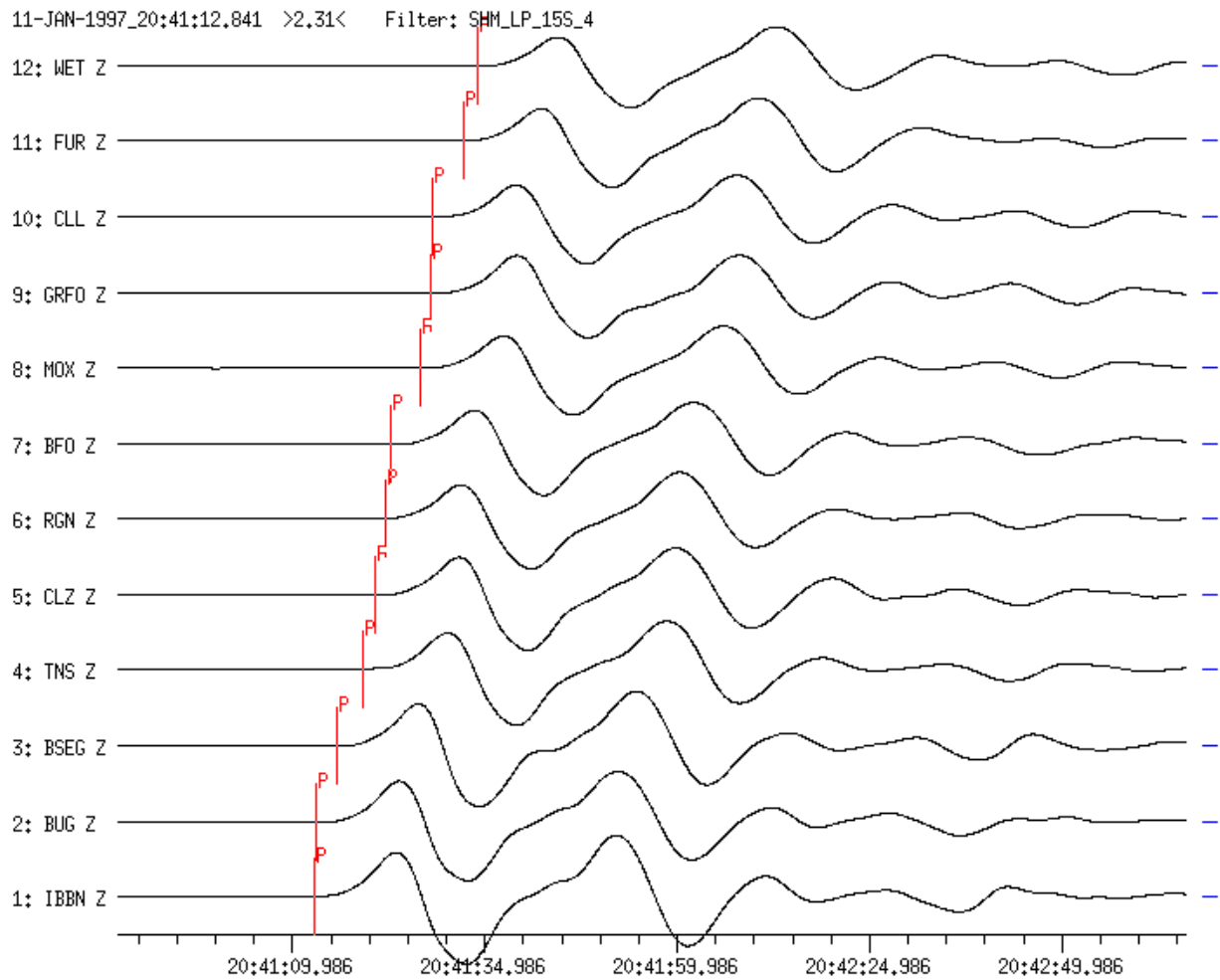
USGS NEIC-data: 1997-01-11 OT 20:28:26.0 18.193 N 102.800 W h = 33G  
 mb = 6.5 Ms = 6.9 (D = 90.9° and BAZ = 299.7° from GRFO site)



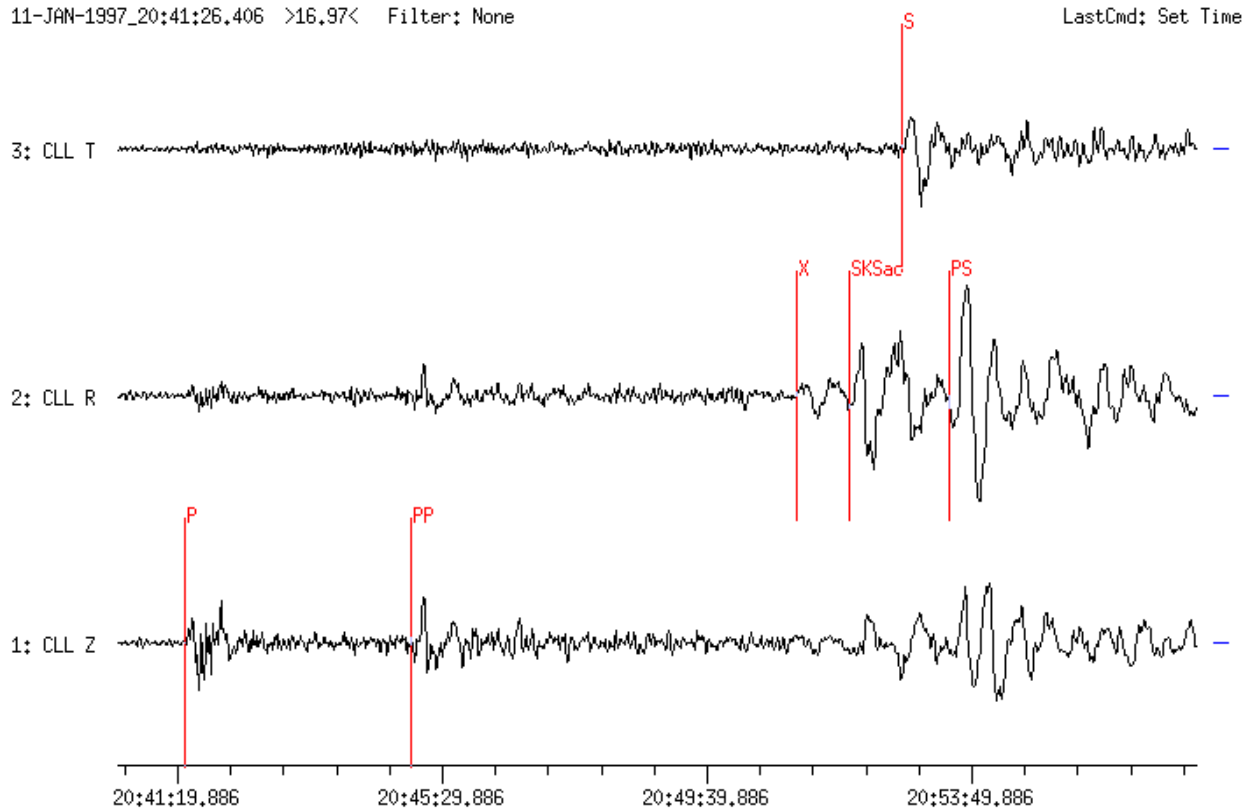
**Figure 12a** Broadband vertical-component seismograms recorded at 12 GRSN-stations are shown. Traces are sorted according to increasing distance of the stations (D = 87.7° to IBBN and 92.1° to WET). P-wave onsets are marked.



**Figure 12b** Records at the same stations as in Figure 12a, however after application of a short-period WWSSN-LP simulation filter. This is necessary for amplitude and period measurements required for standardized mb body-wave magnitude determinations. Maximum double-amplitudes ( $2A$ ) and half-periods ( $T/2$ ) are marked for the first wave group. Later onsets may belong to secondary or depth phases.

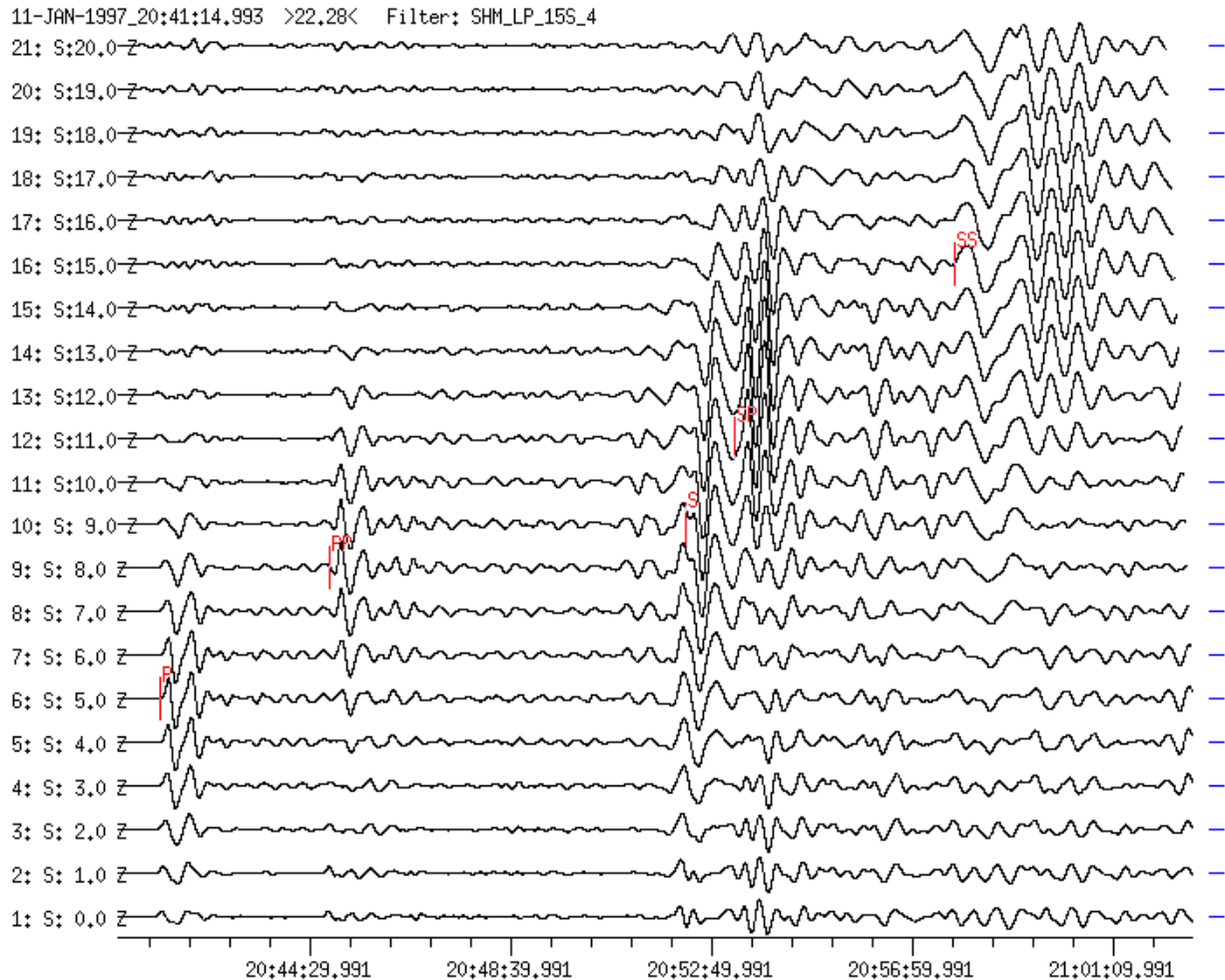


**Figure 12c** Records at the same stations as in Figure 12a, however after application of a 4<sup>th</sup>-order Butterworth low-pass filter ( $f_c = 15$  s). They show the long-period energy content of the earthquake with P-wave periods of about 20s! All long-period network traces are coherent. In this case the network can be used as an array.

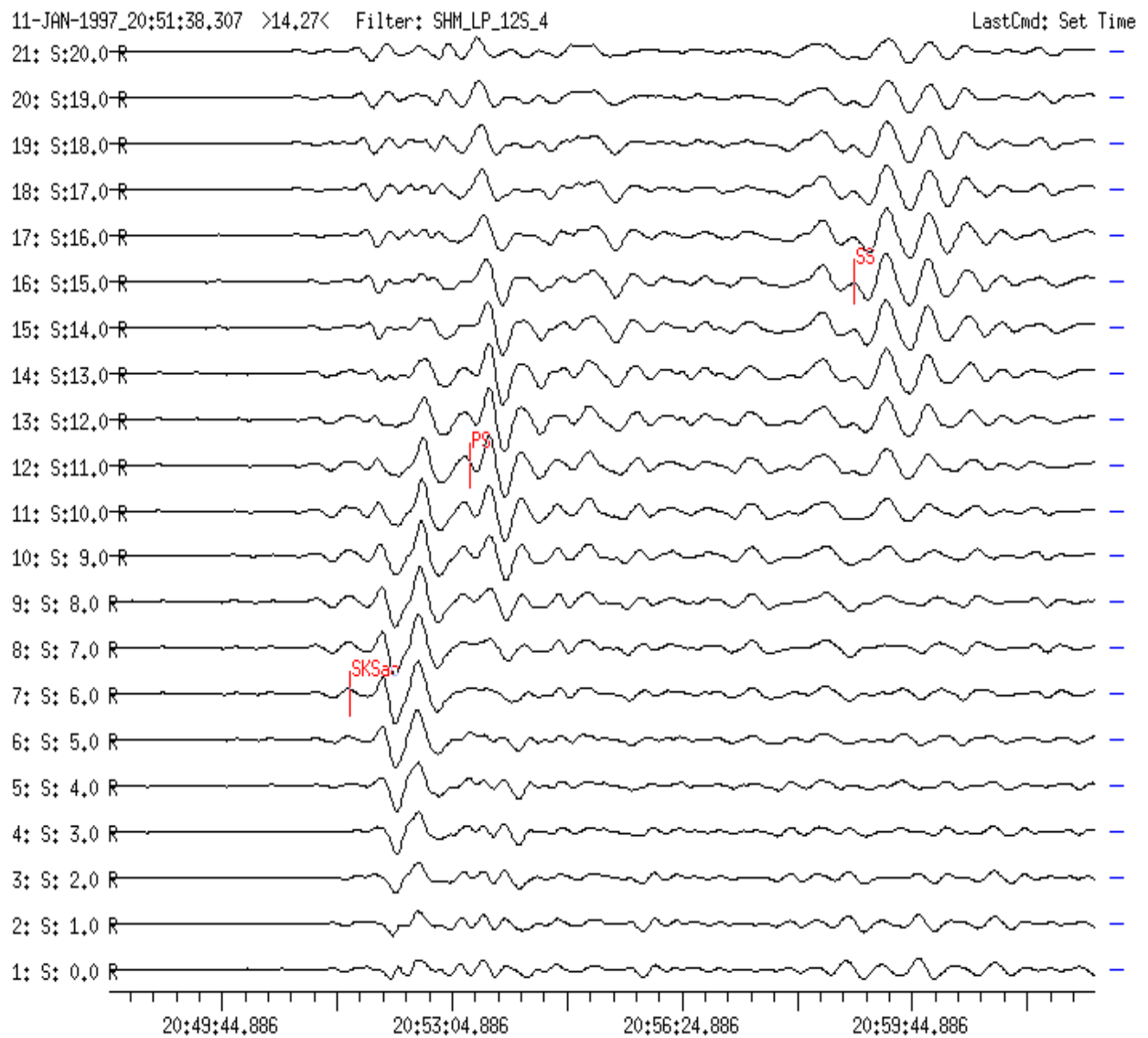


**Figure 12d** Broadband three-component seismogram recorded at station CLL ( $D = 91.0^\circ$ ,  $BAZ = 301^\circ$ ). Note that at teleseismic distances P waves and their multiples show up most clearly in the Z component because of the small incidence angle, decreasing with distance, and the oscillation of P waves in ray direction. The horizontal components are rotated, with R into the source direction. The radial component shows most clearly the direction of wave propagation (or toward the source; i.e., with a  $180^\circ$  ambiguity which can be resolved by taking the direction of P-wave first-motion – up or down – into account; see EX 11.2). Accordingly, in the R component all waves show up clearly which are polarized in the Z-R plane. These are, besides the primary and multiple P waves also all waves which have been converted during their propagation at discontinuities from P into S and vice versa such as PS, PPS, SP, SSP etc. but also SKP, PKS, SKS, SKKS etc. The latter had to travel at least one segment along their travel path through the Earth's outer core (K) which is liquid and transmits no S waves. S waves, however, which have been generated by mode conversion from P waves, can oscillate only in the same Z-R plane as P itself. And when a mantle S waves hits the core-mantle boundary, only their vertically polarized SV energy can partially be converted into a P wave and penetrate into the outer core while any SH energy will be totally reflected back into the mantle. This also explains, why multiple S-wave reflections such as SS, SSS, SSSS contain an increasing part of SH energy. The primary S waves, as originating from the earthquake rupture process, may contain both SV and SH energy in variable proportions, depending on the rupture orientation in space. The SH part shows up in records of the transversal (T) component. When comparing energy arrivals in the S-wave time window in the T and R components one can discriminate between S and SKS (see figure above). Note that SKS and PS are recorded also in Z, however with smaller amplitudes because of their oscillating perpendicular to the ray orientation. An unknown phase, denoted X, appears ahead of SKS in R. Surface waves are not visible because they arrive outside of the displayed time window.

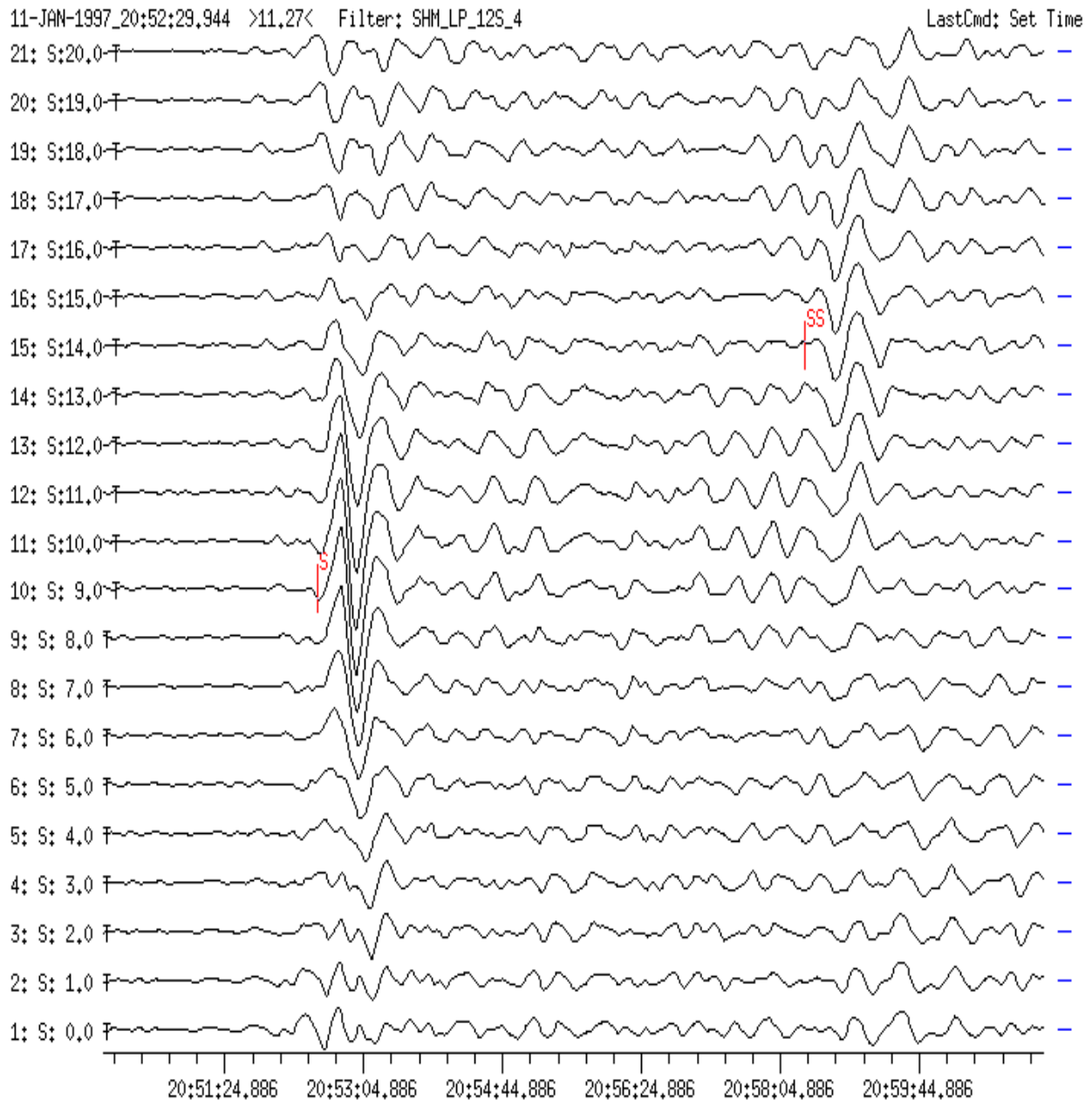
For single station analysis phases can be determined with the help of (differential) travel-time curves (see Figure 2 in Exercise EX 11.2). For a station network vespagram analysis (see also Chapter 9, section 9.7.7) proves to be a much better analysis method because of the additional slowness determination. The following three figures (12e - 12g) show vespagrams of the vertical (Z), radial (R) and transverse (T) components recorded at 12 GRSN-stations. To get a better signal coherency, all traces were filtered with a 4<sup>th</sup>- order long-period low-pass filter ( $f_c = 15$  s).



**Figure 12e** The identified phases from the vertical-component vespagram are marked on the respective slowness traces where they have their largest amplitudes (e.g. P on trace 6: belonging to a slowness  $S = 5$  s/°, PP on trace 9 with  $S = 8$  s/°, SP on trace 12 with  $S = 11$  s/° and SS on trace 16 with  $S = 15$  s/°).



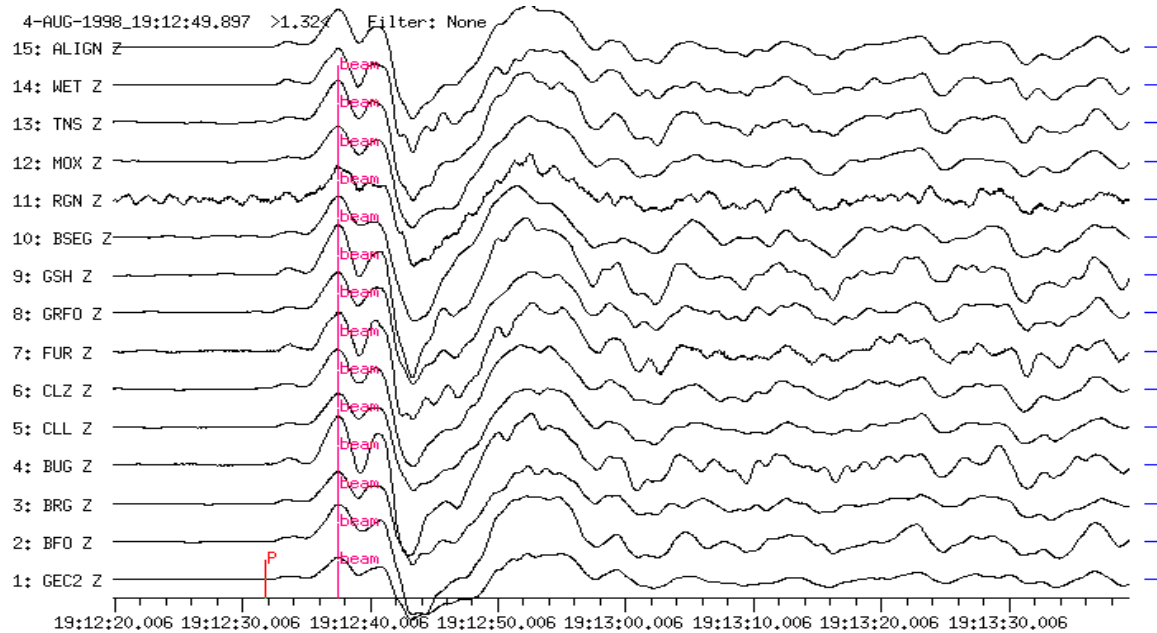
**Figure 12f** The identified phases from the radial-component vespagram are marked on slowness the respective slowness traces (SKSs on trace 7 with a slowness  $S = 6 \text{ s}/^\circ$ , PS on trace 12 with  $S = 11 \text{ s}/^\circ$  and SS on trace 16 with  $S = 15 \text{ s}/^\circ$ ).



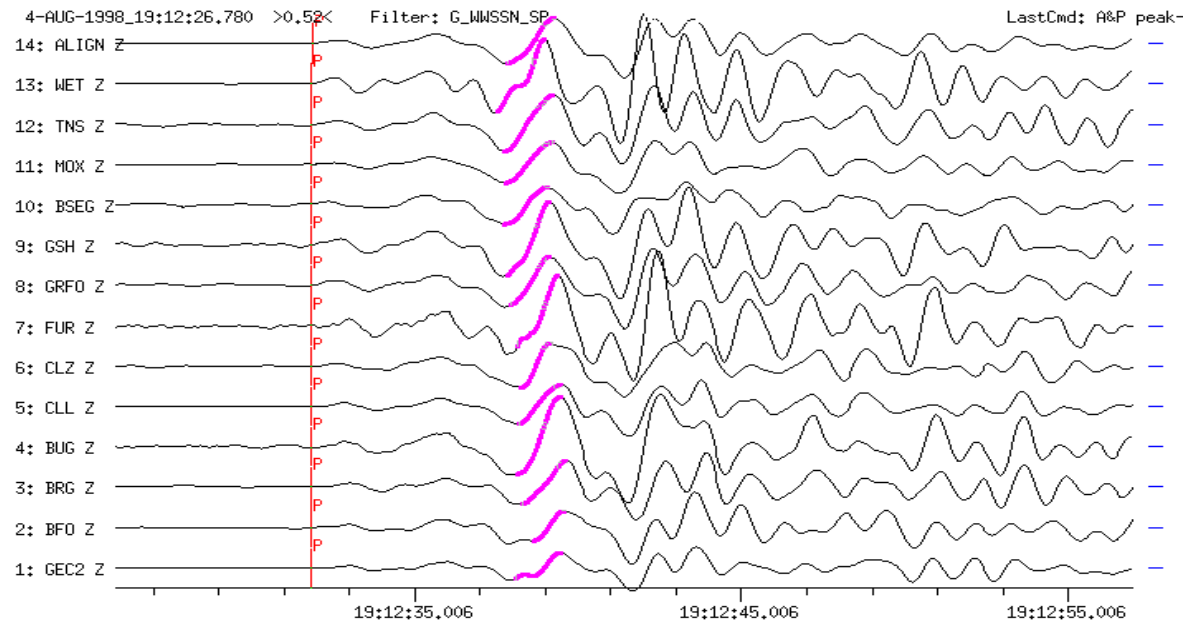
**Figure 12g** The identified phases from the transverse-component vespagram are marked on their respective slowness traces (S on trace 10 corresponding to a slowness  $S = 9 \text{ s}/^\circ$ , SS on trace 15 with  $S = 14 \text{ s}/^\circ$ ). Note, by comparing the identifications in Figs. 12e to 12g that the associated slowness values might somewhat differ because the maximum amplitudes vary only slightly when changing the slowness for about  $\pm 1$  or  $2 \text{ s}/^\circ$ .

### Example 13: Earthquake near the coast of Ecuador

USGS NEIC-data: 1998-08-04 OT 18:59:18.2 0.551S 80.411W h = 19G  
mb = 6.2 Ms = 7.1 (D = 91.5° and BAZ = 270.9° from GRF(GRA1) )

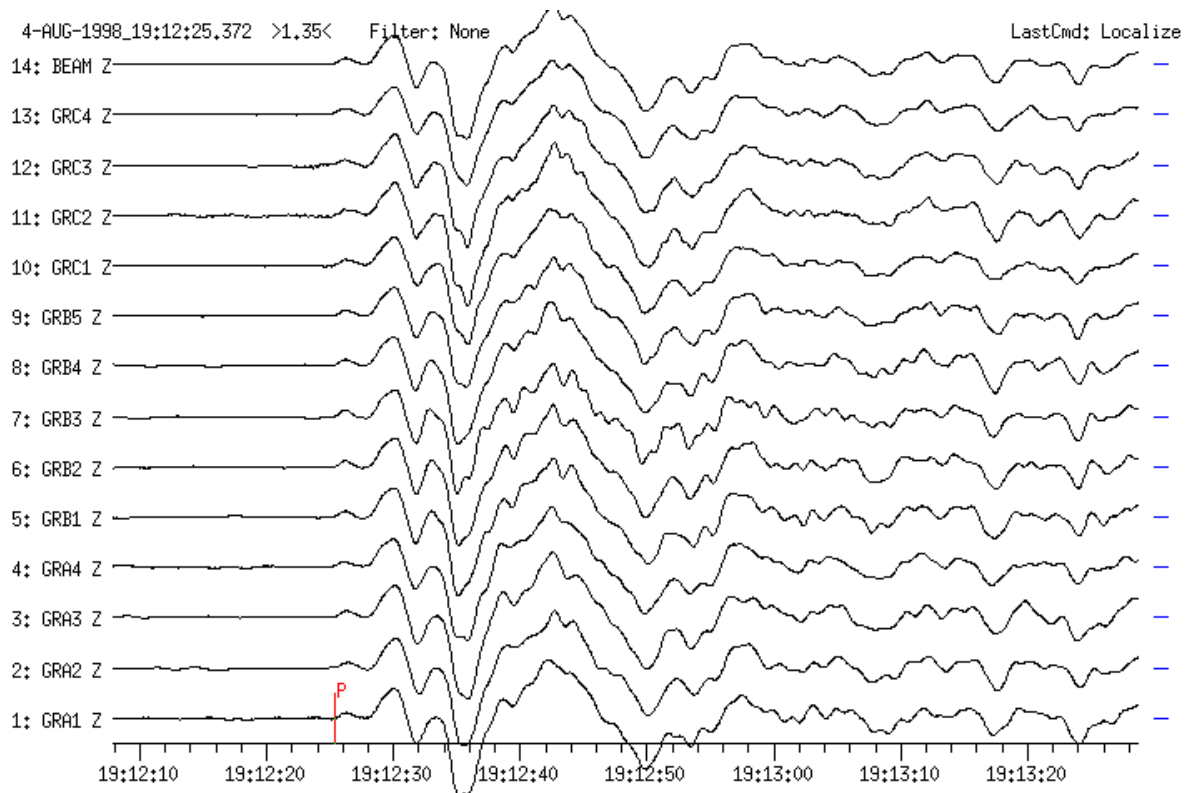


**Figure 13a** Broadband vertical-component seismograms recorded at 14 GRSN stations (D = 88.4° to 93.2°). Traces are time-shifted and aligned. The different waveforms observed at different network stations are coherent in the long-period range.

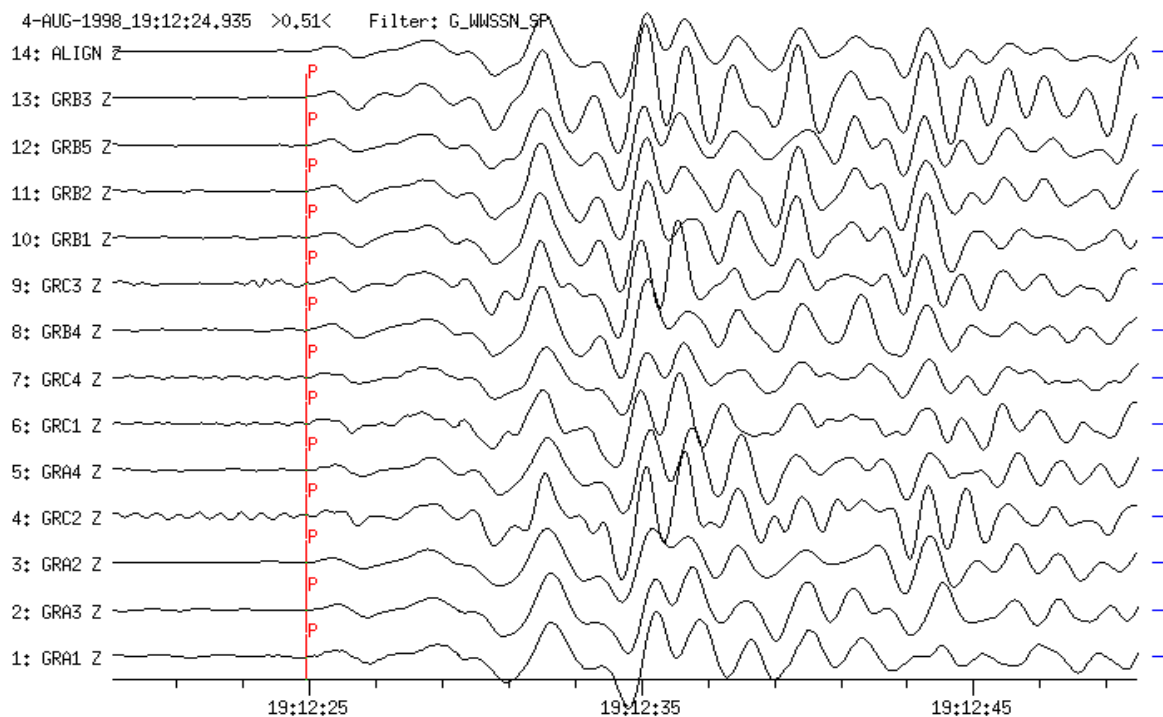


**Figure 13b:** After short-period filtering (WWSSN-SP simulation) the coherency of the different station-waveforms is bad. The maximum double P-amplitudes used for mb-estimation are emphasized. Because of the variability of station amplitudes within the network individual mb-values vary between 6.0 for BFO and GEC2 and 6.7 for other stations.

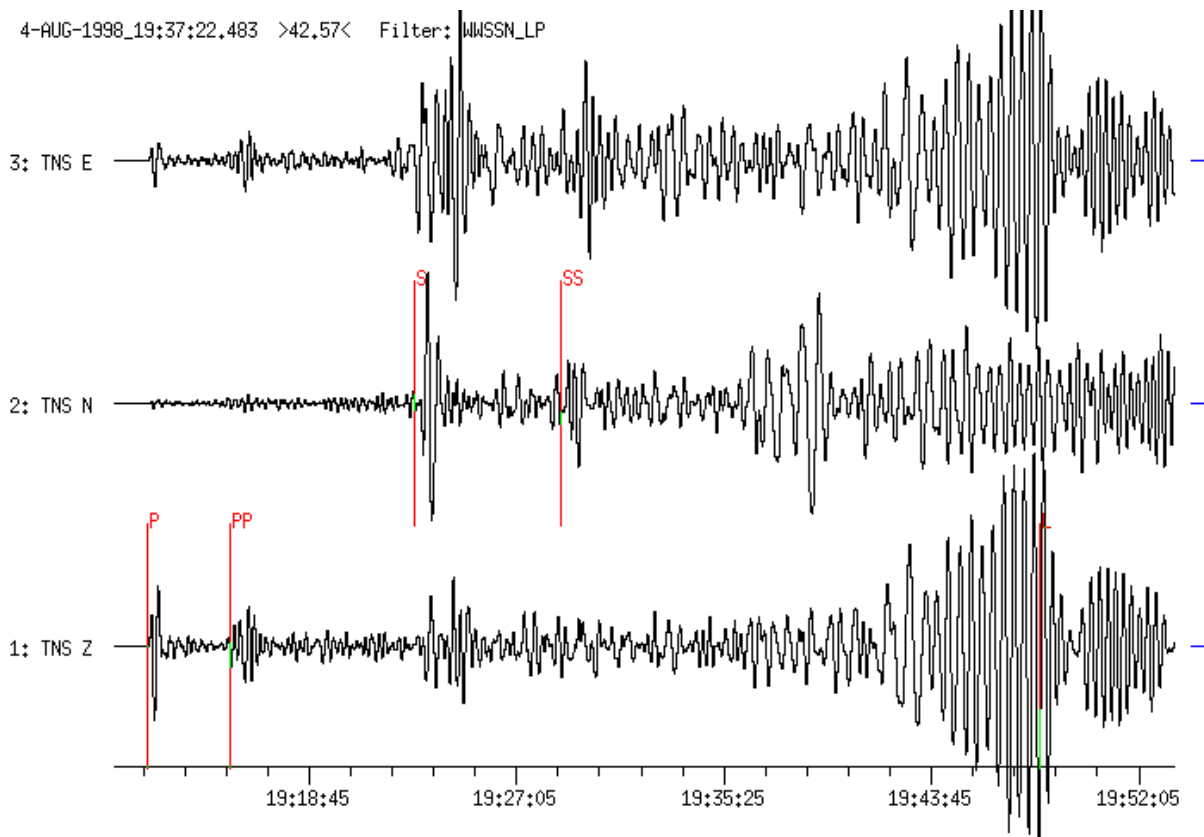




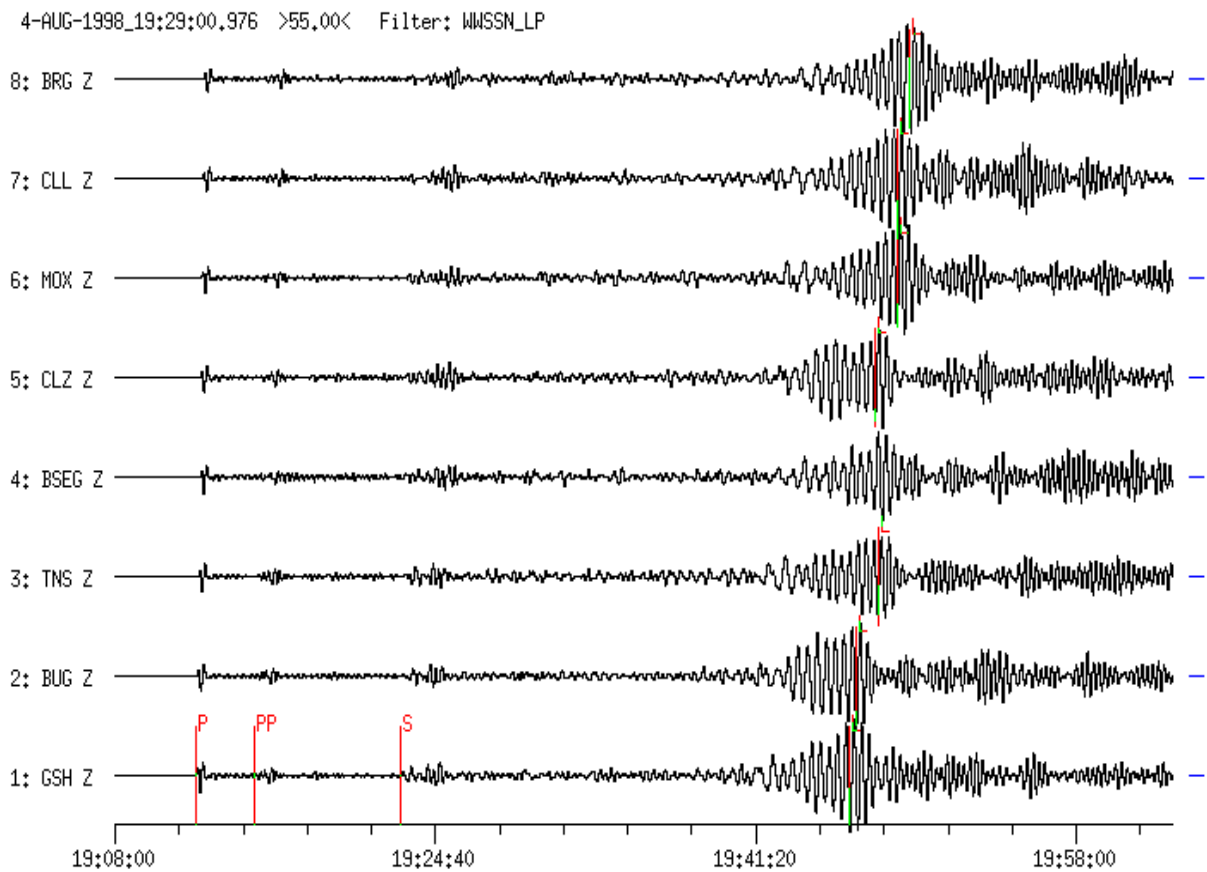
**Figure 13c** Broadband records of the vertical-components at 13 GRF-array stations ( $D = 91.5^\circ$  to  $91.7^\circ$ ). Because of the much smaller aperture of the array as compared to the GRSN the signal coherence is good for all array-traces.



**Figure 13d** Even for the short-period filtered records of the GRF-array stations (WWSSN-SP simulation) the signal coherence is still good.



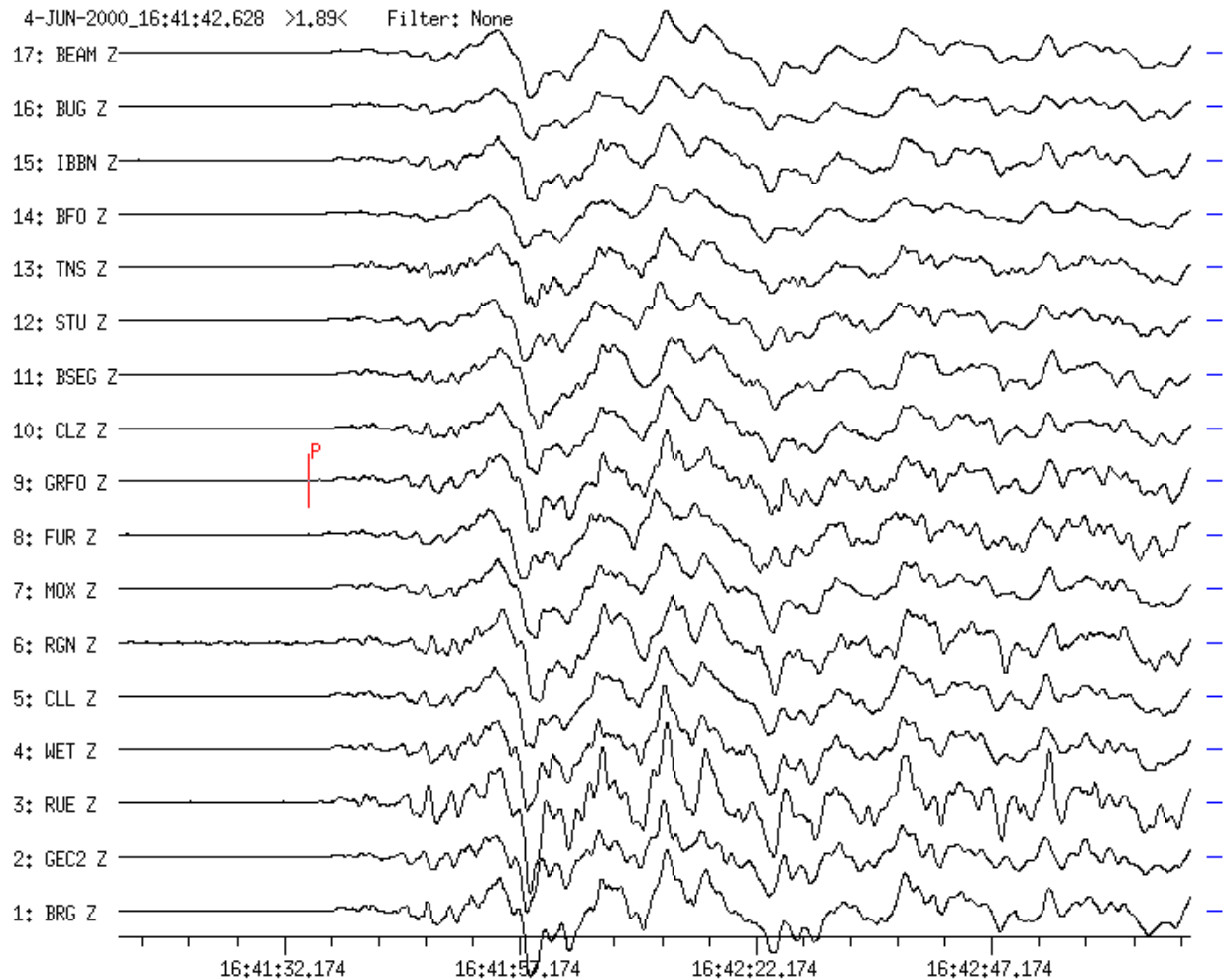
**Figure 13e** Three-component long-period filtered (WWSSN-LP simulation) seismogram recorded at station TNS ( $D = 89.7^\circ$ ,  $BAZ = 269^\circ$ ). Note the much larger P-wave amplitude in the E-W component as compared to the N-S component because of the source location in the west. Also note the strong secondary phases PP, S and SS. The maximum surface-wave amplitude on the vertical component is marked (L). The estimated magnitude is  $M_s = 7.2$ . This is very close to the average value determined by the USGS NEIC from data of the global seismic network ( $M_s = 7.1$ ).



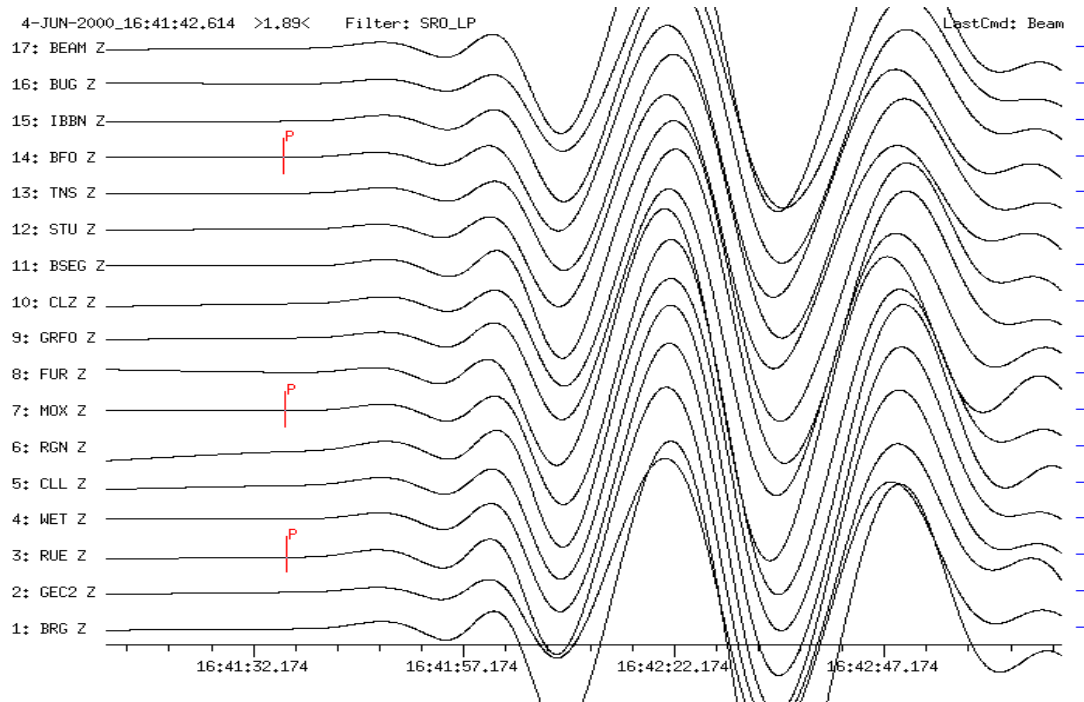
**Figure 13f** Long-period filtered (WWSSN-LP simulation) vertical-component records of the Ecuador earthquake at 8 GRSN stations within the distance range  $D = 88.4^\circ$  (GSH) to  $93.2^\circ$  (BRG). The variability of the maximum surface-wave amplitudes throughout the network is less than that of short-period body-wave amplitudes (compare with Figure 13b). Accordingly,  $M_s$  estimates from individual stations are more reliable than respective  $m_b$  estimates.

**Example 14:** Earthquake in Southern Sumatra Region

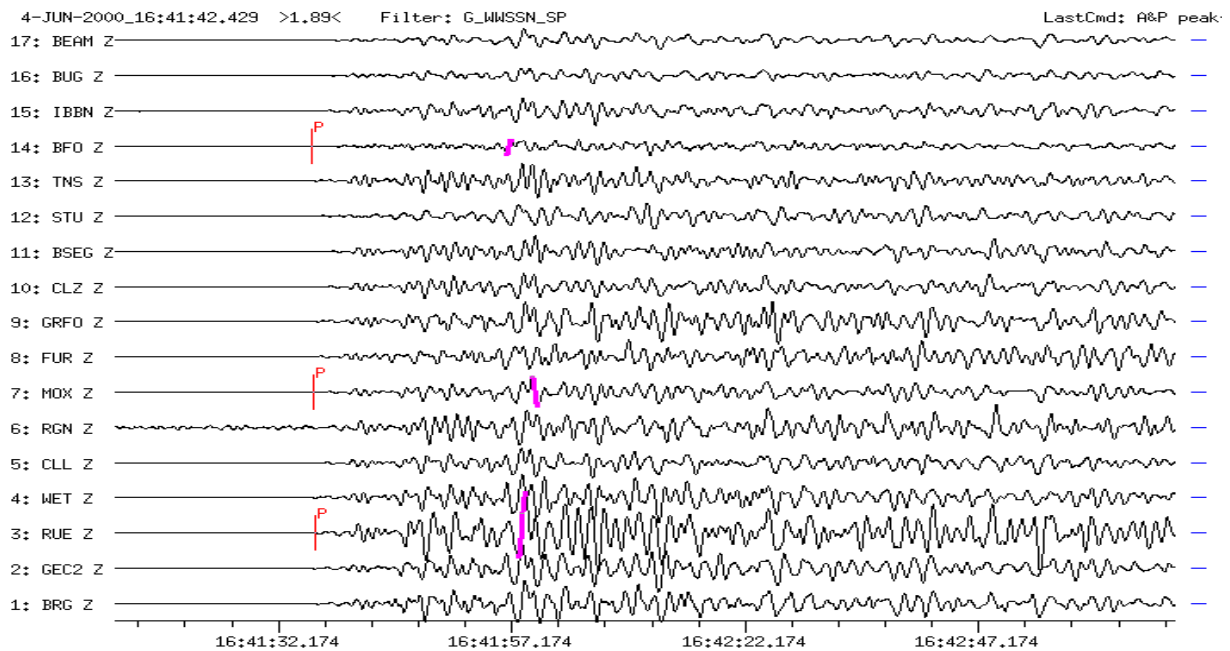
USGS NEIC-data: 2000-06-04 OT 16:28:25.8 4.773 S 102.050 E h = 33G  
 mb = 6.8 Ms = 8.0 (D = 94.1° and BAZ = 92.5° from GRF)



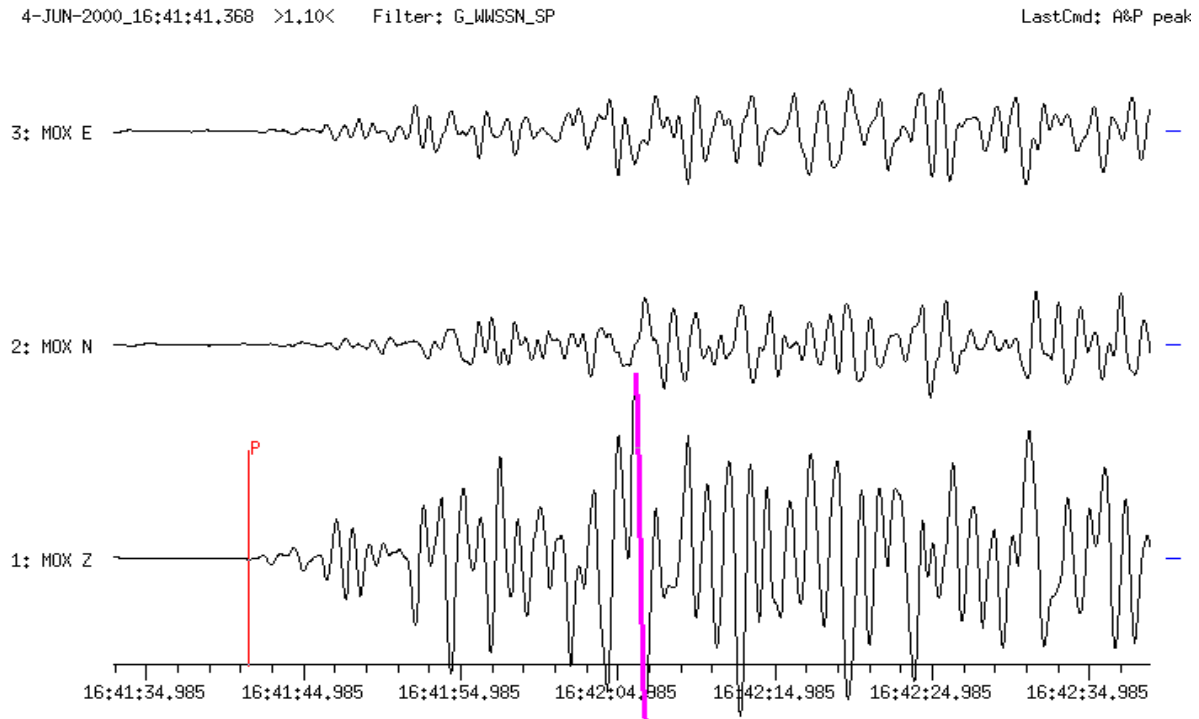
**Figure 14a** Broadband vertical-component seismograms recorded at 16 GRSN-stations within the distance range  $D = 92.5^\circ$  to  $96.7^\circ$ . Traces are time-shifted and aligned for better comparison of the individual records. Note the good coherency of all traces. For these broadband records the network works like an array. The weak P-wave first arrival is marked on the record of station GRFO. Maximum P-wave amplitudes were recorded about half a minute later (multiple rupture event with successively larger energy release?).



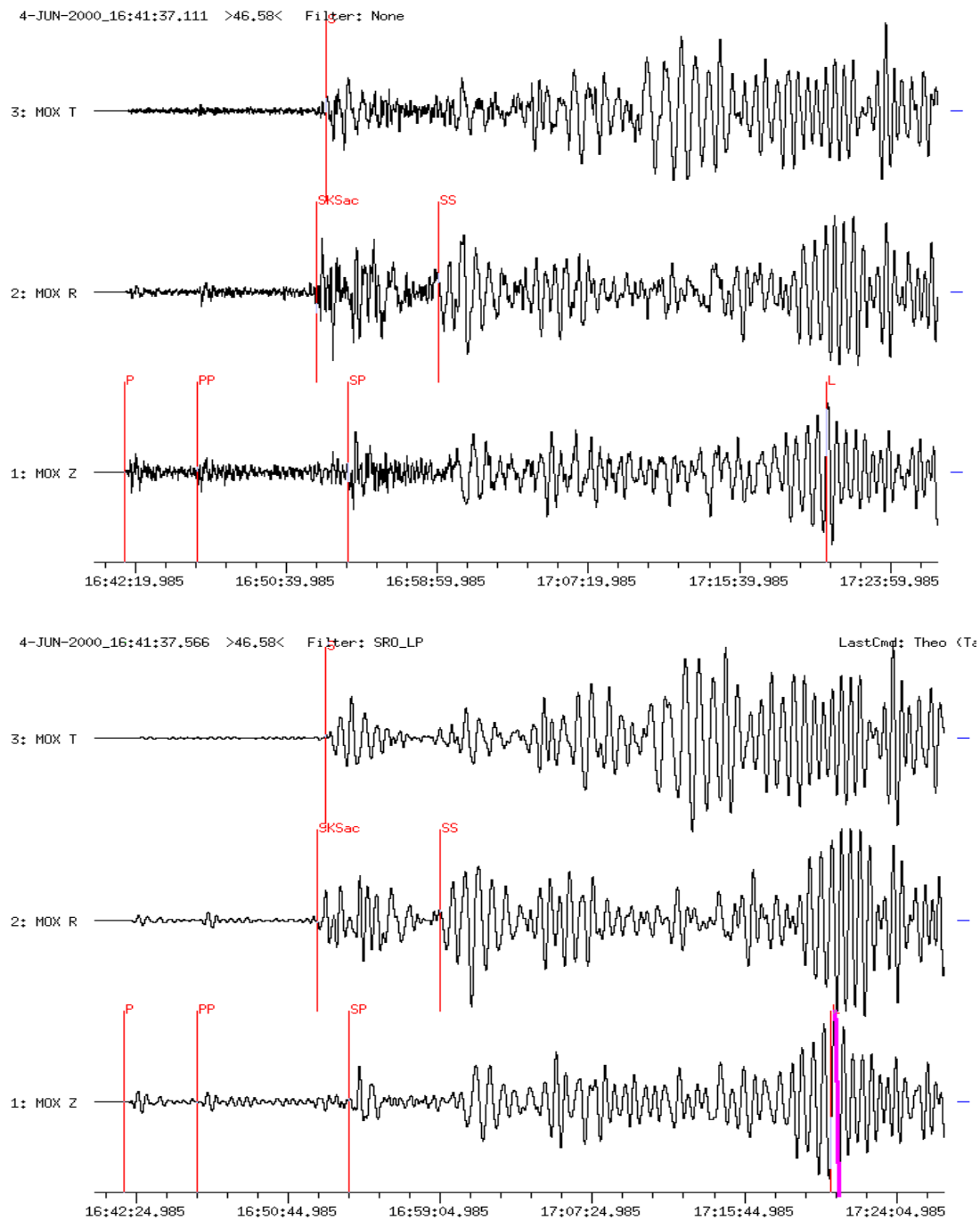
**Figure 14b** Seismograms of the same earthquake recorded at the same stations as in Figure 14a after applying a long-period filter (SRO-LP simulation). Note the very high signal coherence of the long-period P-waves with periods  $T \approx 25$  s. Also the amplitudes are comparable at all stations.



**Figure 14c** For mb estimates short-period records (WWSSN-SP simulation) have to be used to measure the maximum amplitudes with periods near 1 s. The respective filtered traces for the same earthquake and stations as in Figure 14a are shown. For three stations the P wave arrivals and their maximum double-amplitudes  $2A$  have been marked. The latter differ by a factor of five for this event. This corresponds to a difference of 0.7 magnitude units!



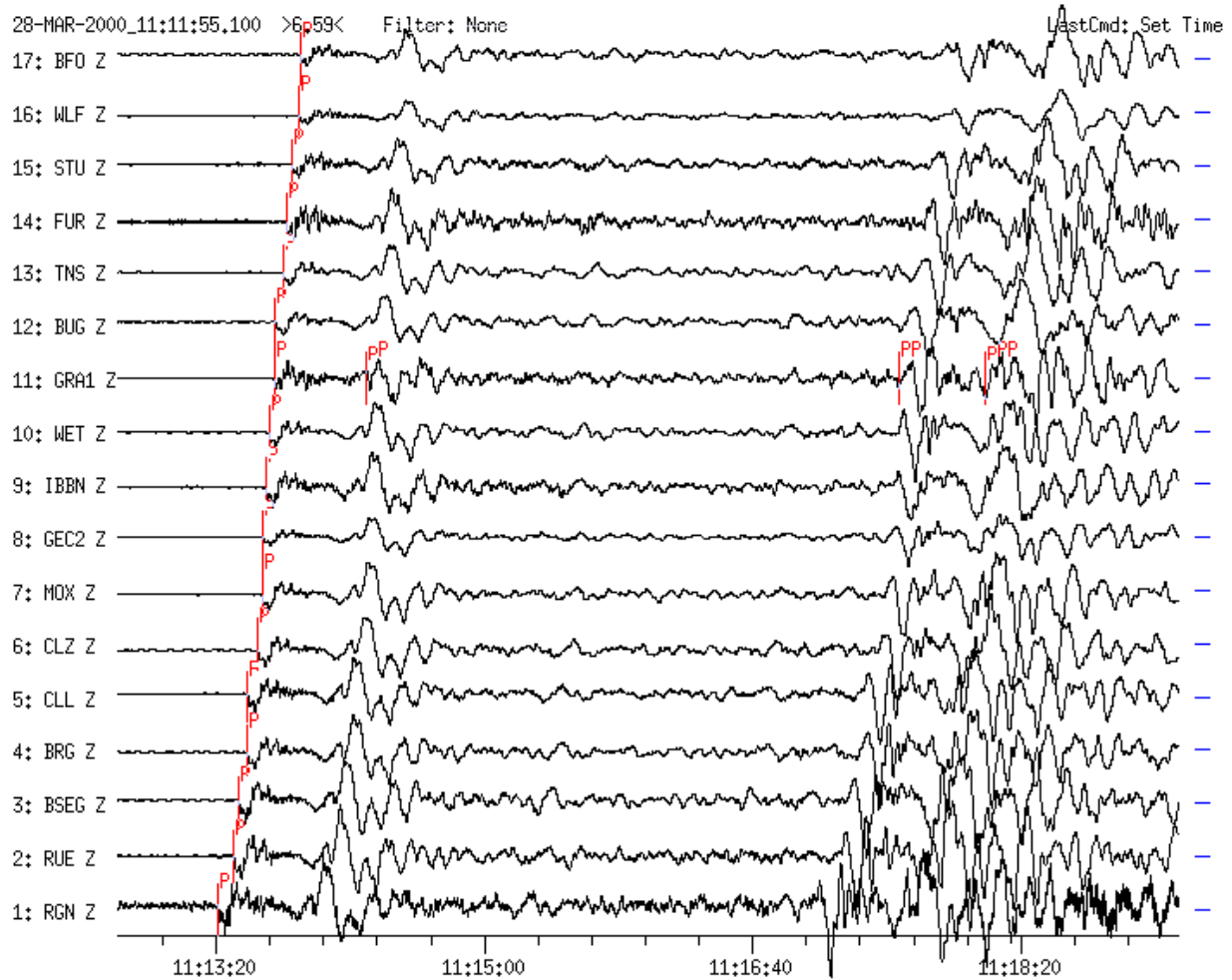
**Figure 14d** Three-component single station record of the Sumatra earthquake at station MOX ( $D = 93.9^\circ$ ,  $BAZ = 93^\circ$ ). The short-period filtered P-wave group is shown and the maximum peak-to-peak amplitude  $2A$  is marked as in the Z component. In this display with high time resolution the complicated rupture process of this event, consisting of a series of sub-events with successively increasing energy release within the first 25 s is very obvious. This is a very important information also for more detailed research work on source processes. Therefore, also for routine data analysis a detailed reporting of the onset times, amplitudes and periods of these sub-events is strongly recommended. An example for such an analysis is given in Tab. 3.1 of Chapter 3. In the given case, the amplitude of the P-wave maximum around 16:42:06 is about 40 times larger than that of the first arrival. This corresponds to a difference of 1.6 magnitude units between the initial and the largest sub-event of this earthquake. Accordingly, instructions for mb measurements within the first five half-cycles (NOAA in the 1960s) or within the first 5 s (IMS nowadays) might dramatically underestimate not only the energy release of earthquake in general (stronger ones in particular, because mb will always saturate at values around 6.5; see Figs. 3.5 and 3.16 and related discussions in sections 3.1.2.3, 3.2.5.2, and 3.2.7 of Chapter 3). Even worse, they will particularly underestimate the high-frequency energy release which is most relevant for potential damage assessments and for these mb is principally more suited than  $M_s$  or  $M_w$  provided, that mb is determined really from the maximum amplitudes in short-period P-wave train.



**Figure 14e** Broadband (top) and long-period filtered 3-component seismograms (SRO-LP simulation; below) of the Sumatra earthquake at station MOX ( $D = 93.9^\circ$ ). The phases P, PP, S, SKSac, SP, SS and the maximum surface wave L have been marked. For better phase identification the horizontal components N and E have been rotated into the R and T directions. The S wave is seen best on the transversal component T and P, PP, SKS and SP on Z and on the radial horizontal component R, respectively. For surface-wave magnitude estimation the maximum ground displacement with a period between 18 and 22 s was measured at the marked position L from the SRO-LP filtered vertical component. Additionally, amplitudes from the horizontal components can be used for a horizontal surface-wave magnitude value.

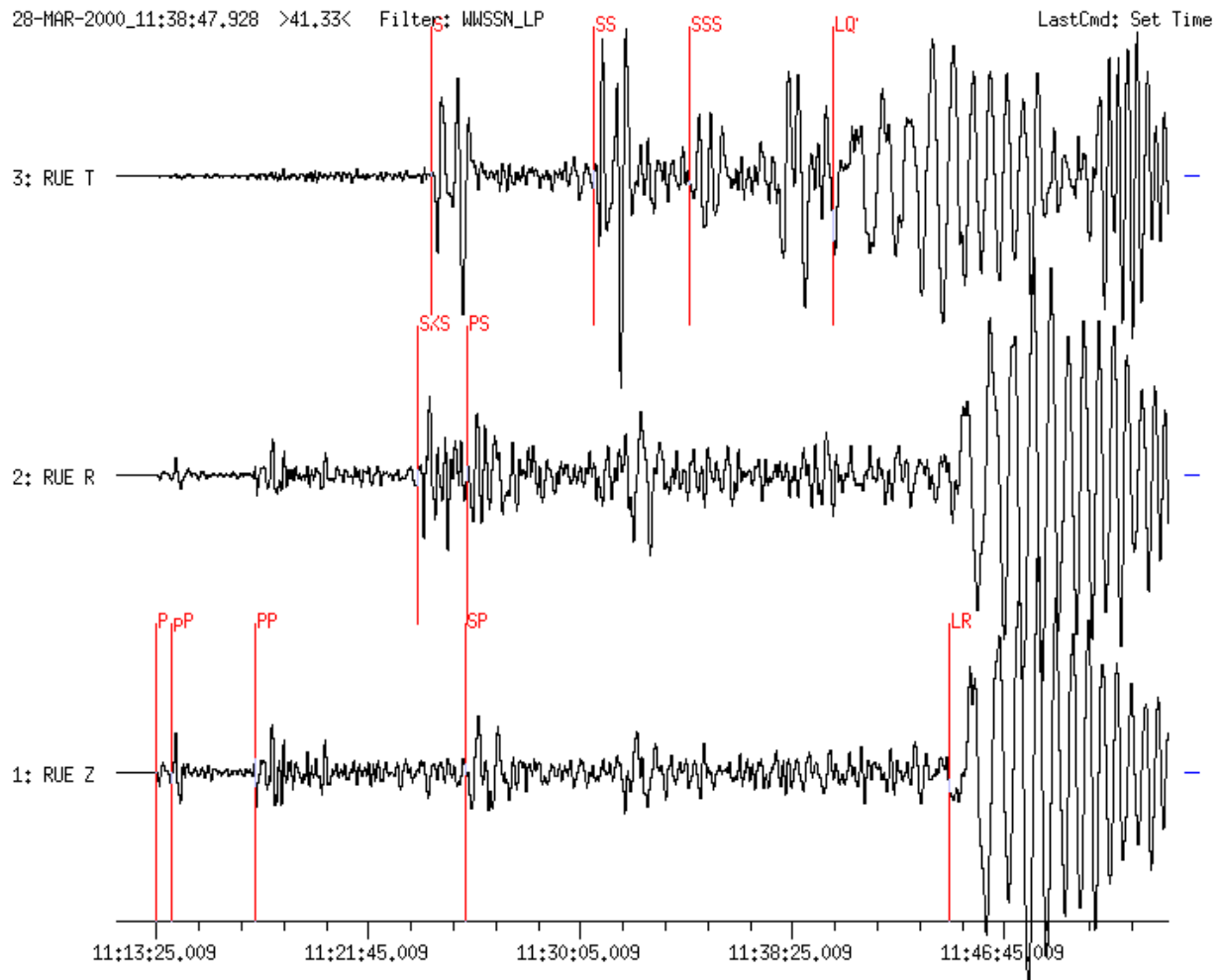
### Example 15: VOLCANO ISLANDS REGION

USGS NEIC-Daten: 2000-03-28 OT 11:00:21.7 22.362N 143.680E h = 119D  
mb = 6.8 (D = 96.8° BAZ = 43.5° from GRF(GRA1))



**Figure 15a** Broadband vertical-component seismograms recorded at 17 GRSN-, GRF- and GEOFON-stations. Traces are sorted according to increasing distance (D = 92.4° to RGN and 99.1° BFO). The phases P, PP and the depth phases pP and pPP are recorded very well.





**Figure 15b** Long-period filtered (WWSSN-LP simulation) three-component seismogram of the Volcano Island earthquake recorded at station RUE near Berlin ( $D = 93.7^\circ$ ,  $BAZ = 45^\circ$ ). The horizontal components are rotated, with R in source direction. Phases P, pP, SP and the onset of the Rayleigh waves LR are marked on the vertical component, SKS and PS on the radial component and S, SS, SSS as well as the Love waves onset LQ on the transversal component, respectively.

| Topic   | Additional seismogram examples at distances beyond 100°                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p><b>Klaus Klinge</b> (formerly Federal Institute for Geosciences and Natural Resources, 30655 Hannover, Germany);<br/>E-mail: <a href="mailto:klaus.klinge@googlemail.com">klaus.klinge@googlemail.com</a></p> <p><b>Siegfried Wendt</b>, Geophysical Observatory Collm, University of Leipzig, D-04779 Wermsdorf, Germany; E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a></p> <p><b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> |
| Version | October, 2002; DOI: 10.2312/GFZ.NMSOP-2_DS_11.3                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |

**Note:** Most of the examples given below show either records of the German Regional Seismic Network (GRSN; aperture about 500 x 800 km) or of the Gräfenberg broadband array (GRF; aperture 45 x 110 km; see Figs. 8.14 and 9.4 in the manual Chapters 8 and 9, respectively). The following abbreviations have been used: OT – origin time in UT (universal time), D – epicentral distance in degrees, BAZ – backazimuth in degrees, h – focal depth in kilometer.

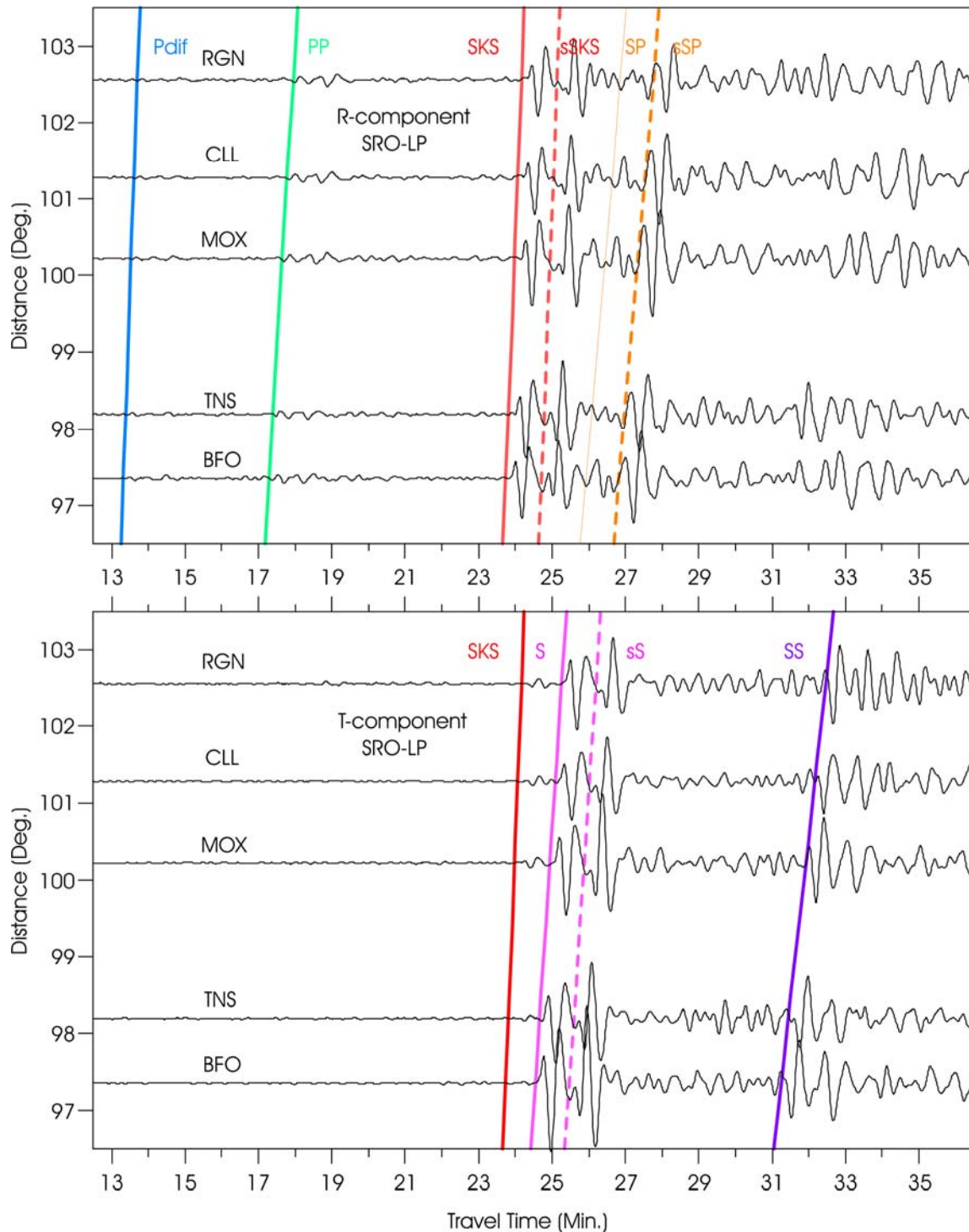
### Example 1: Earthquake in the Chile-Bolivia border region (intermediate source depth)

**USGS NEIC-data: 2002-03-08 OT 04:56:21.7 21.6S 68.2W h = 123 km Mw = 6.5;**  
**The epicentral distances of the GRSN stations range between 97° < D < 103°.**

Figures 1 a and b show enlarged sections of the recordings made of this intermediate depth earthquake at stations of the GRSN around the beginning of the shadow zone of the Earth's core (for the full records see Fig. 11.57 left). Figure 1a depicts the first 24 minutes of the long-period recordings (SRO-LP simulation) of the horizontal R and T components whereas Figure 1 b shows the related Z-component recordings of the first 8 min (lower part) and 17 min (upper part), respectively. The following diagnostic features can be recognized:

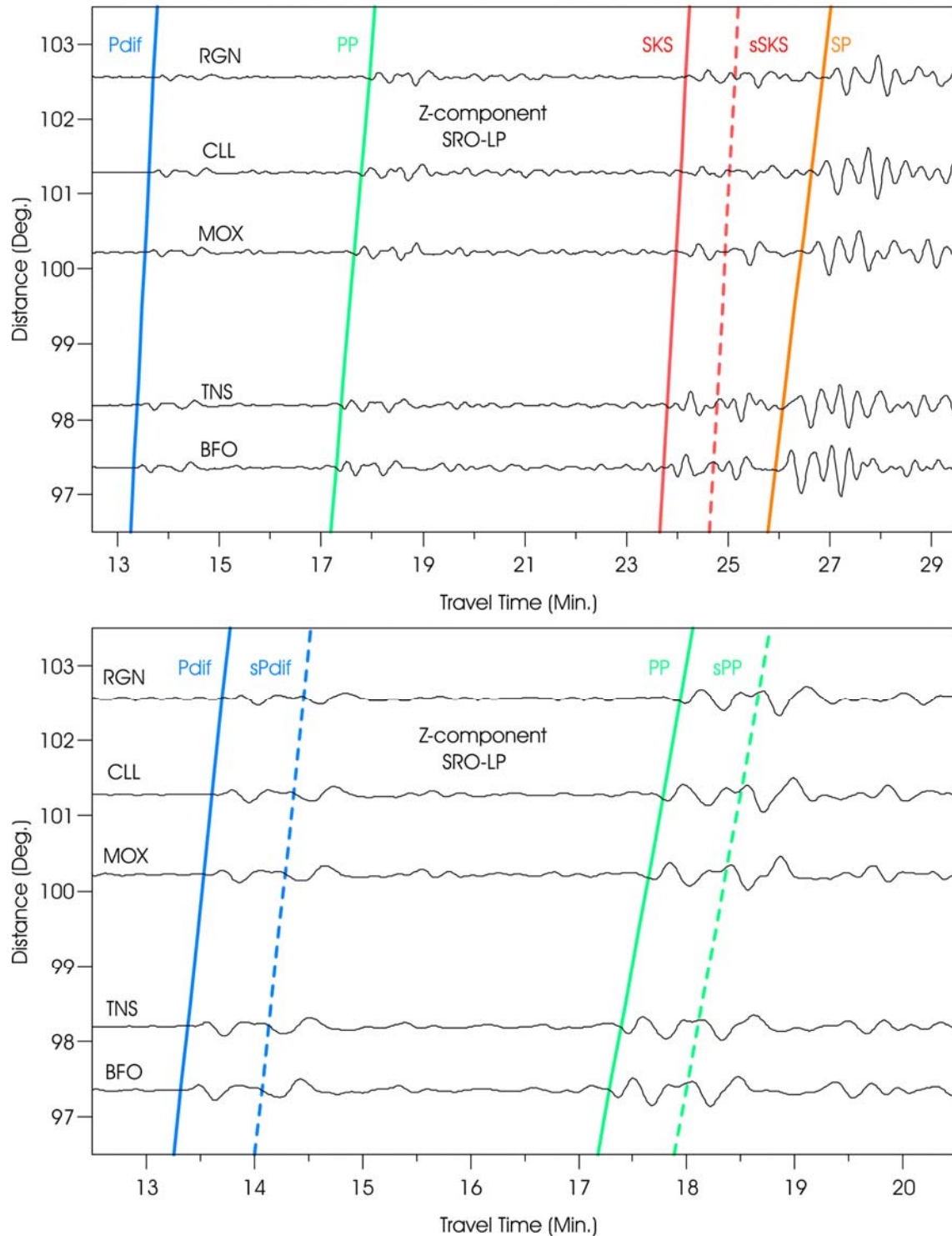
- Pdif is by far the smallest body-wave arrival, even in the vertical component, however, its SNR is still large enough for this medium-size earthquake in LP recordings;
- SKS (here also with its depth phase sSKS) forms the first strong shear-wave arrival, necessarily in the radial (R) horizontal component, whereas S (also with its depth phase sS) is a comparably strong later shear-wave arrival, here (however not always!) with large amplitudes in the transverse (T)-component records;
- SP (and its depth phase) is also (necessarily!) strongest in the R component whereas SS is frequently (not always! see Figure 2) strongest in the long-period T component;
- SKS and SP have also well developed onsets in Z-component LP records with amplitudes comparably strong or even larger than for Pdif and PP;
- Pdif and PP may also have well developed depth phases (here sPdif and sPP, respectively) in Z-component LP records;
- the missing of the depth phases pPdif and pPP in Figure 1b is due to the different P- and S-wave radiation pattern from a shear source (see Figs. 3.25 and 3.26) and the specific rupture orientation with respect to the seismic station in the considered case; this is not a general feature, rather pP, pPP etc. are often stronger;
- PP has, when compared with Pdif, a larger R component because of its larger incidence angle; and SP a larger Z component than PS.

2002-03-28 OT 04:56:21.7 h = 123 km Chile Bolivia border region



**Figure 1a** SRO-LP filtered BB records of GRSN stations. Upper traces: R component; lower traces: T component. Note that the phases SKS and SP, as well as their depth phases, have necessarily their largest horizontal amplitudes in the R component, as the longitudinal waves Pdif and PP. In contrast, S and SS may have their largest amplitudes in the T component, depending on the primary ratio of SV/SH energy radiated by the earthquake source.

2002-03-28 OT 04:56:21.7 h = 123 km Chile Bolivia border region

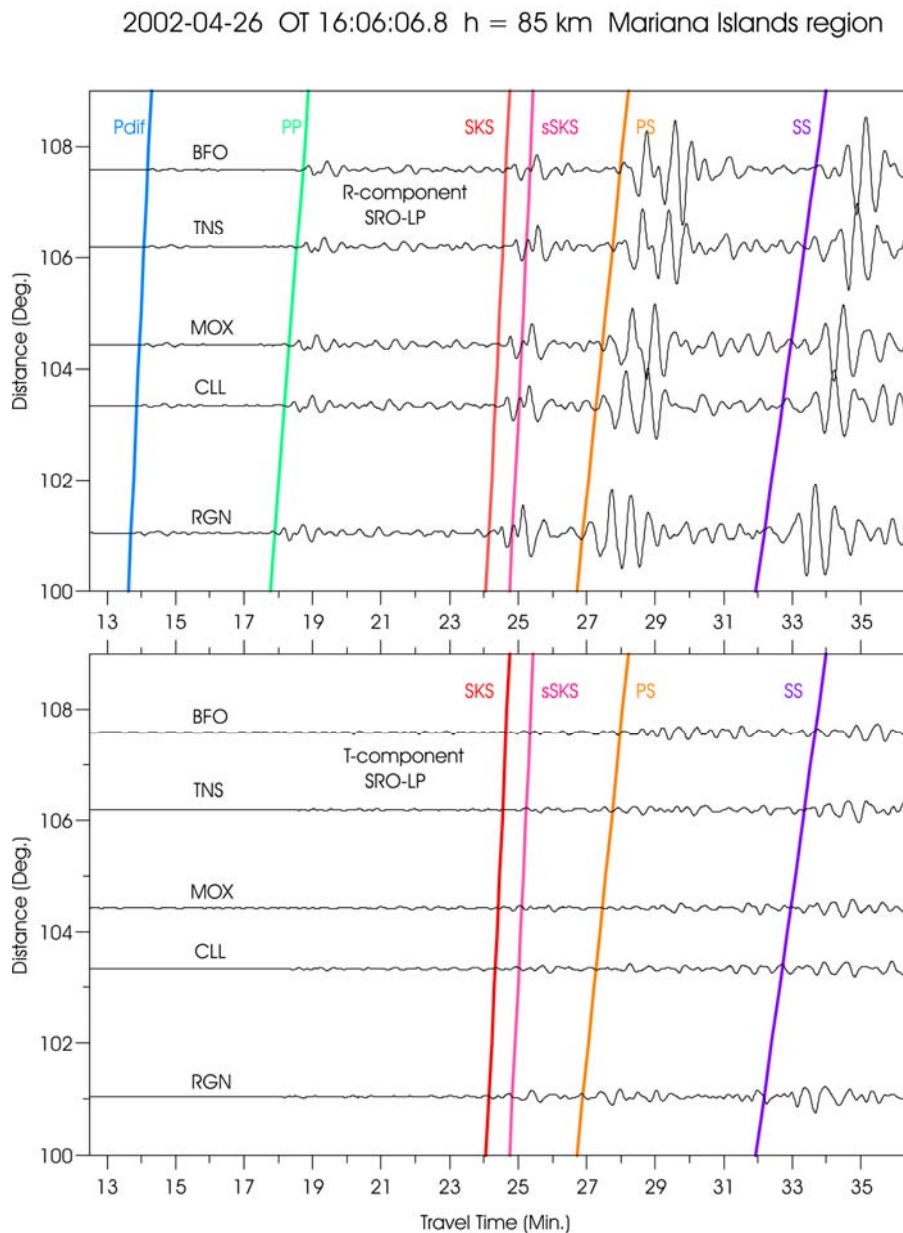


**Figure 1b** Enlarged vertical (Z)-component SRO-LP filtered BB records of GRSN stations shown in Figure 1a. Upper traces: Body-wave phases up to SP; lower traces: cut-out of the first 8 minutes of the record with longitudinal phases only. Note the well developed depth phases sPdif and sPP. The theoretically expected arrival times according to the IASP91 model (Kennett and Engdahl, 1991) have been inserted in both Figure 1a and b.

## Example 2: Earthquake in the Mariana Islands region (intermediate source depth)

USGS NEIC-data: 2002-04-26 OT 16:06:06.8 13.1N 144.6E h = 85 km  $M_w = 7.1$ ;  
The epicentral distances of the GRSN stations range between  $101^\circ < D < 108^\circ$ .

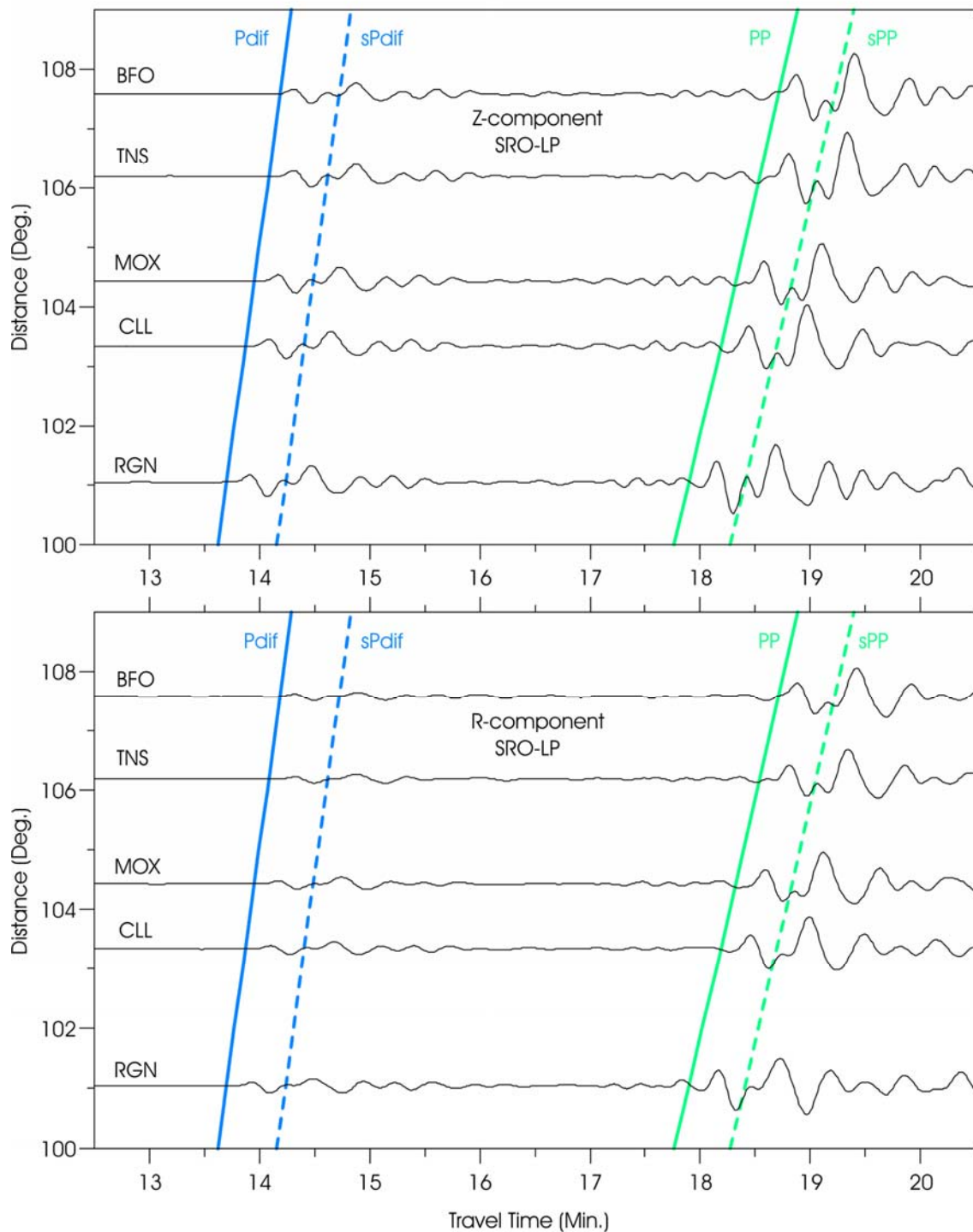
The waves from this source at nearly the same distance as for the Chile-Bolivia earthquake approach the GRSN stations from nearly the opposite backazimuth (see Fig. 11.57), however, the general wave types and waveform features in different record components are nearly the same as Figure 1, with one exception: SS is best developed in the R component and there is no S wave visible in the T component, i.e., the shear-wave energy generated by this source was almost exclusively of SV type.



**Figure 2a** SRO-LP filtered BB records of GRSN stations. Upper traces: R component; lower traces: T component. Note that in these records PS has about four times larger amplitudes than SKS and all phases are not visible or rather weak in the T component.



2002-04-26 OT 16:06:06.8 h = 85 km Mariana Islands region

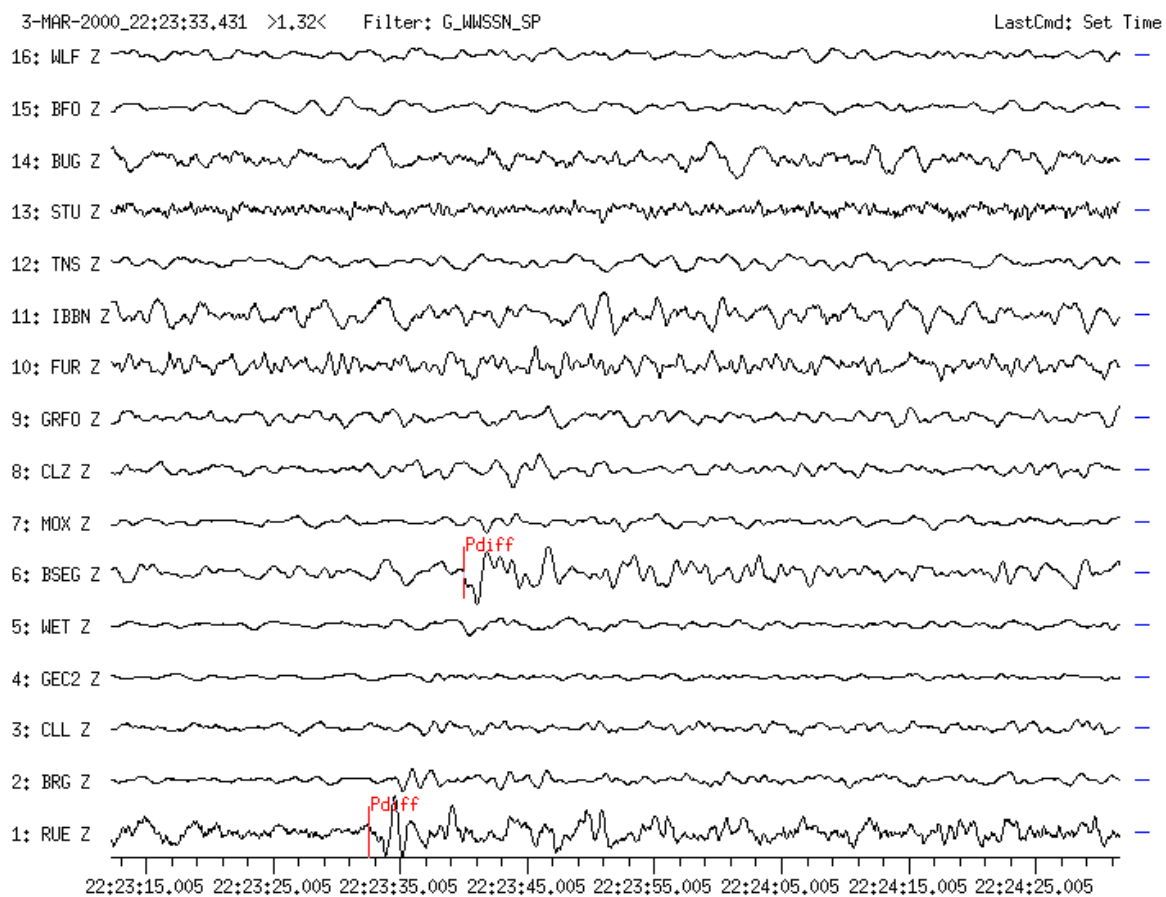


**Figure 2b** Enlarged cut-out of the SRO-LP filtered BB records of GRSN stations shown in Figure 2a. Upper traces: Z component; lower traces: R component. Note the well developed depth phases sPdif and sPP. The theoretically expected arrival times according to the IASP91 model (Kennett and Engdahl, 1991) have been inserted in both Figure 2a and b. Note that the amplitude ratio Z/R is larger than for Pdif than for PP because of the smaller (steeper) incidence angle of Pdif.

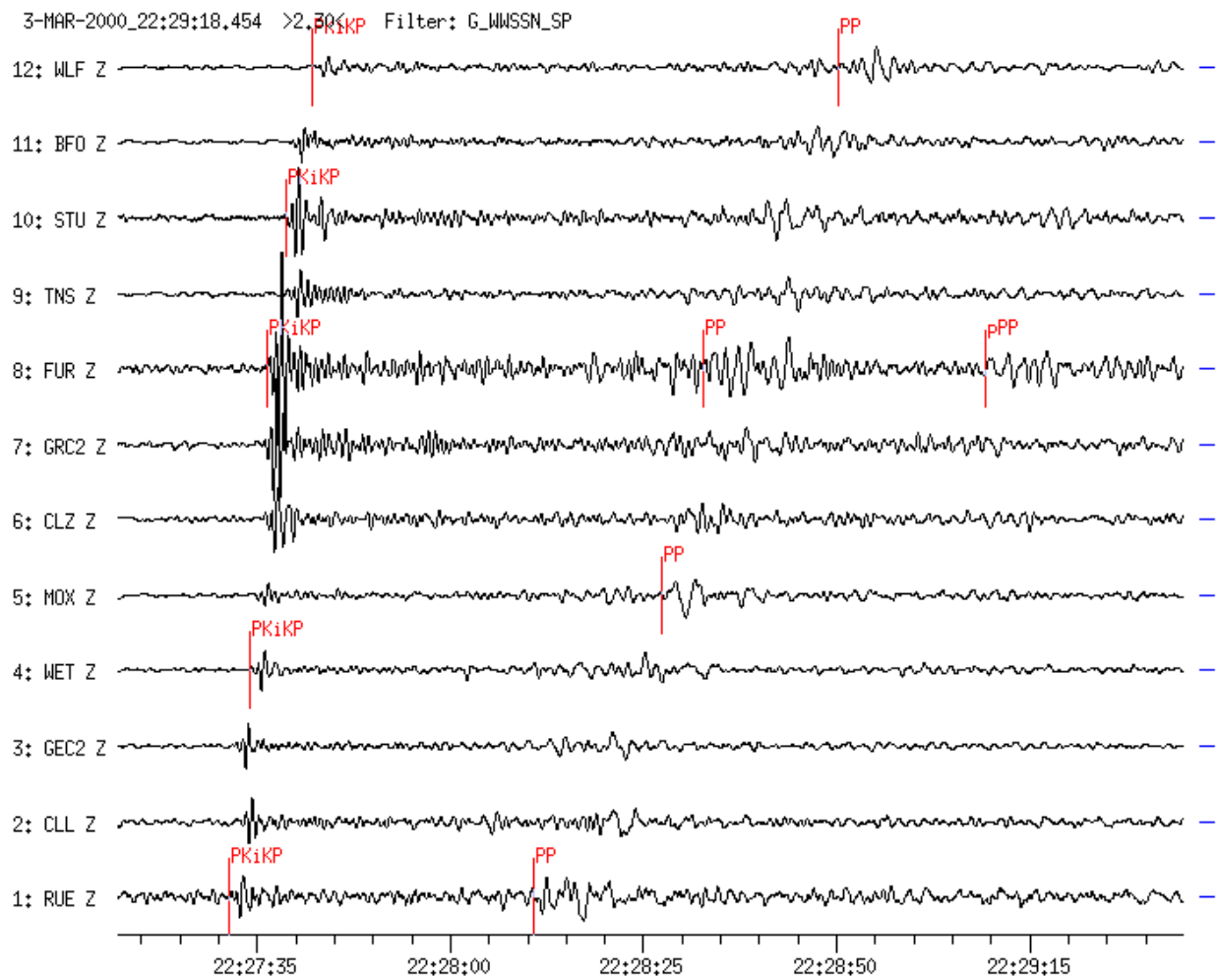
### Example 3: Earthquake in the Banda Sea region (intermediate source depth) with phases PKPdif, PKKPab and PKKPbc

USGS NEIC-data: 2000-03-03 OT 22:09:13.5 7.313S 128.642E h = 148 km mb = 6.4;  
(D = 113° and BAZ = 73.9° from GRF, h = 140 km)

Shown are various short-period filtered seismograms (WWSSN-SP simulation) recorded at GRSN-, GRF- and GEOFON-stations. All traces are sorted according to increasing epicentral distance within the range D = 110.8° (RUE) and 116.2° (WLF). Phases Pdif, PKiKP, PP, SP, PKKPbc and PKKPab are shown. Additionally, ray-paths and travel-time curves are presented.

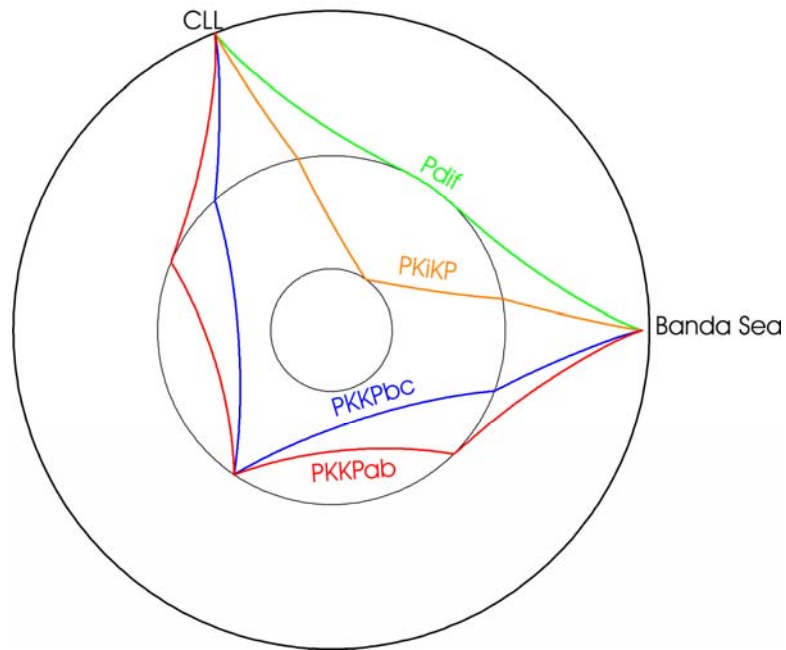


**Figure 3a** Pdif (old name Pdif) is the first arrival, however, as a diffracted wave, rather small, particularly in short-period records. Only a few GRSN stations have recorded it from this event. For more clear long-period records of Pdif see Figures 4b, 6c and 7b in the next examples.

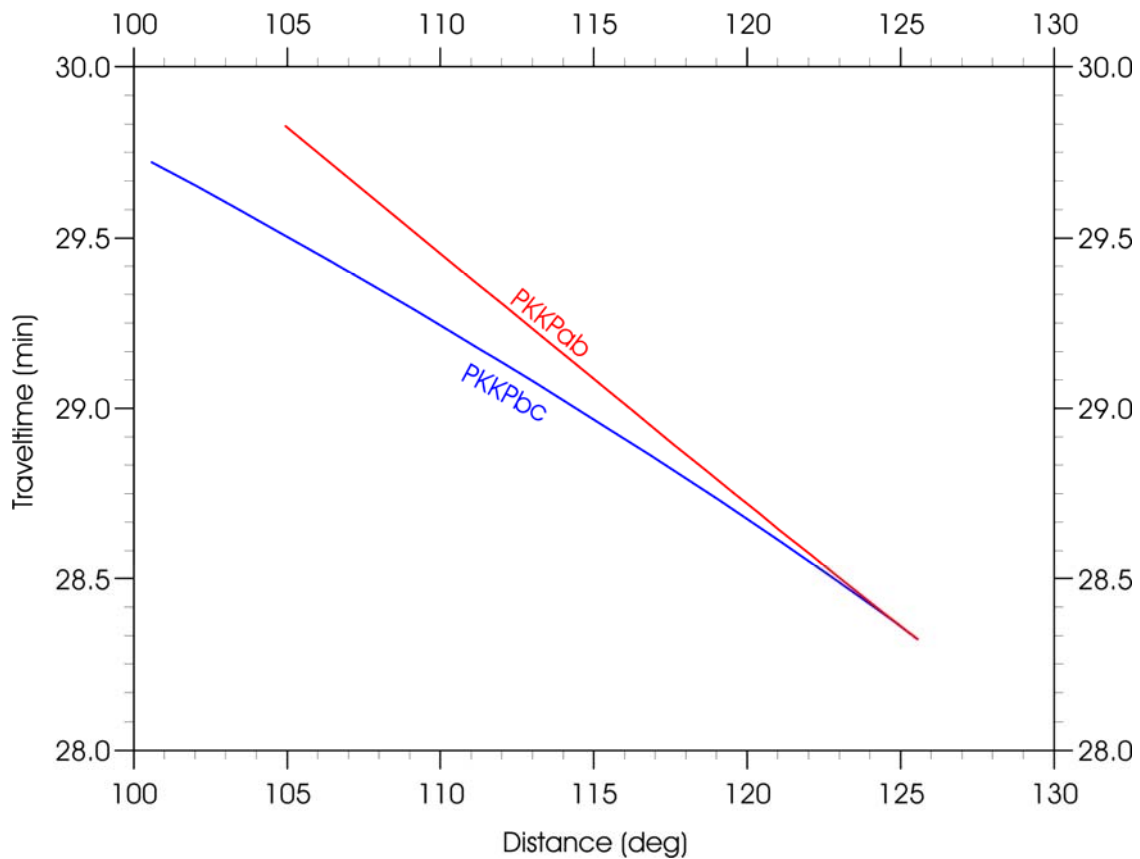


**Figure 3b** Phases PKiKP, PP and (pPP) are shown on this seismogram. PKiKP arrives about 4 min after Pdif shown in Figure 3a. Note the strong variation at PKiKP-amplitudes within the network by a factor of about 10. The onset-time of PP can not be determined exactly. The reflection point of PP is below the complex crustal structure of Tibet.

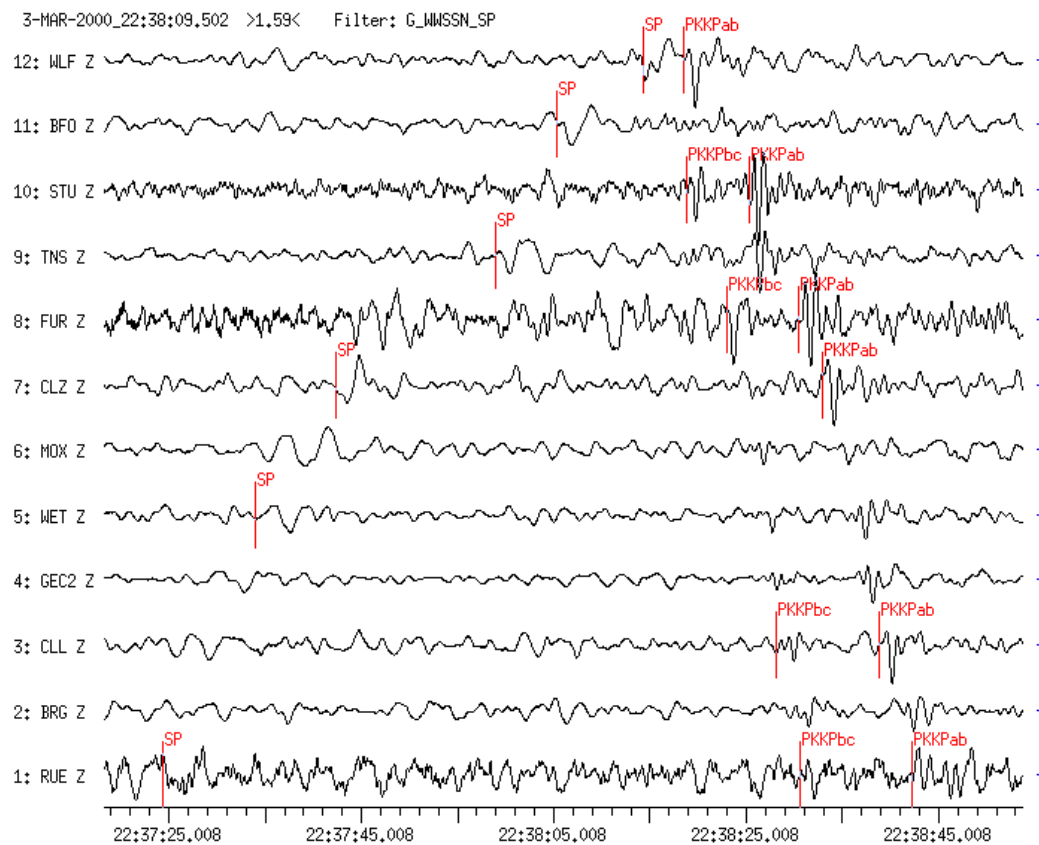




**Figure 3c** Ray paths for Pdif, PKiKP, PKKPbc and PKKPab for the considered event at a focal depth of 148 km and with an epicentral distance to station CLL of  $D = 111^\circ$ .



**Figure 3d** Travel-time curves for PKKPbc and PKKPab for a focal depth of 148 km.

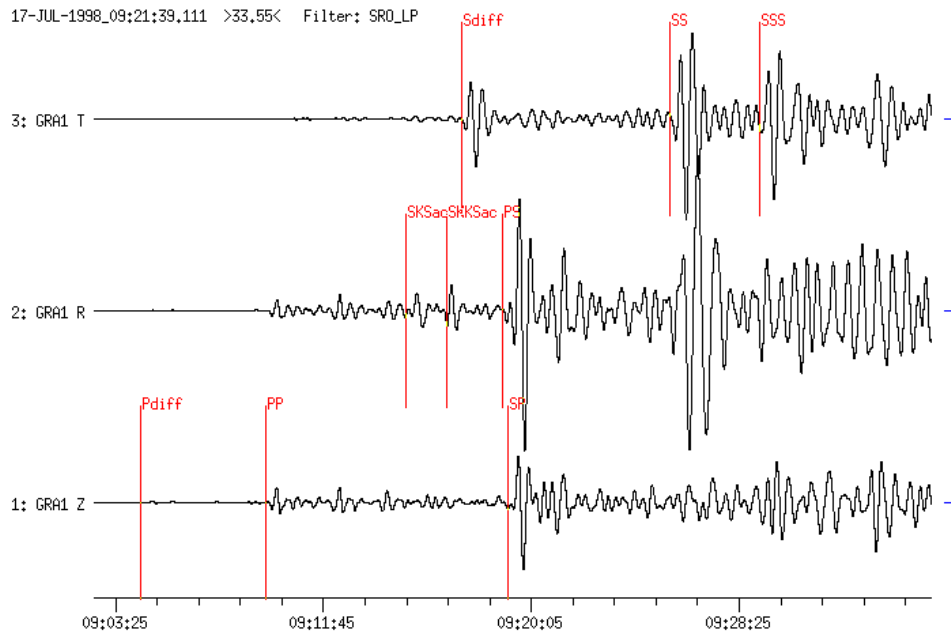


**Figure 3e** Vertical component short-period seismograms starting about 10 min after the PKiKP arrival in Figure 3b. Marked are the onset times of the phases SP, PKKPbc and PKKPab. For the branching of PKKP between about  $90^\circ$  and  $125^\circ$  see Figure 9 in EX 11.3. These late secondary phases appear in short-period records only. Their ray paths are shown in Figure 3c and their travel-time curves in Figure 3d above.

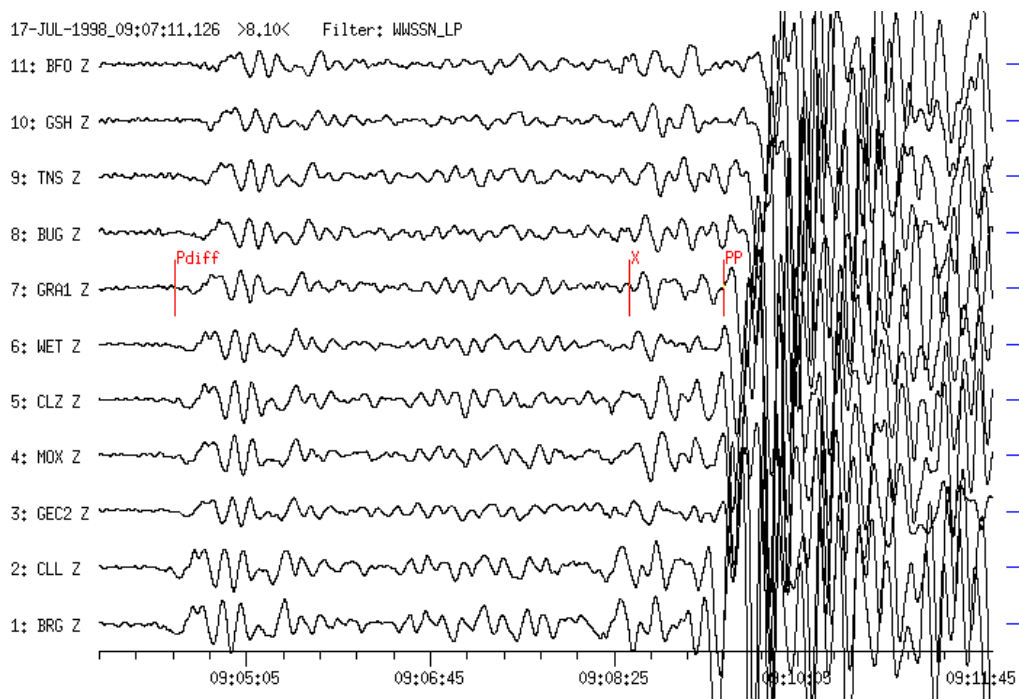
### Example 4: Earthquake in Papua New Guinea

**USGS NEIC-data: 1998-07-17 OT 08:49:15 3.08S 141.76E h = 33G Ms = 7.0;  
(D =  $117.5^\circ$  and BAZ =  $58.8^\circ$  from GRA1)**

This earthquake occurred near the coast of Papua New Guinea. Tsunami waves with a height of 10 m flooded the coast and killed about 3000 people.



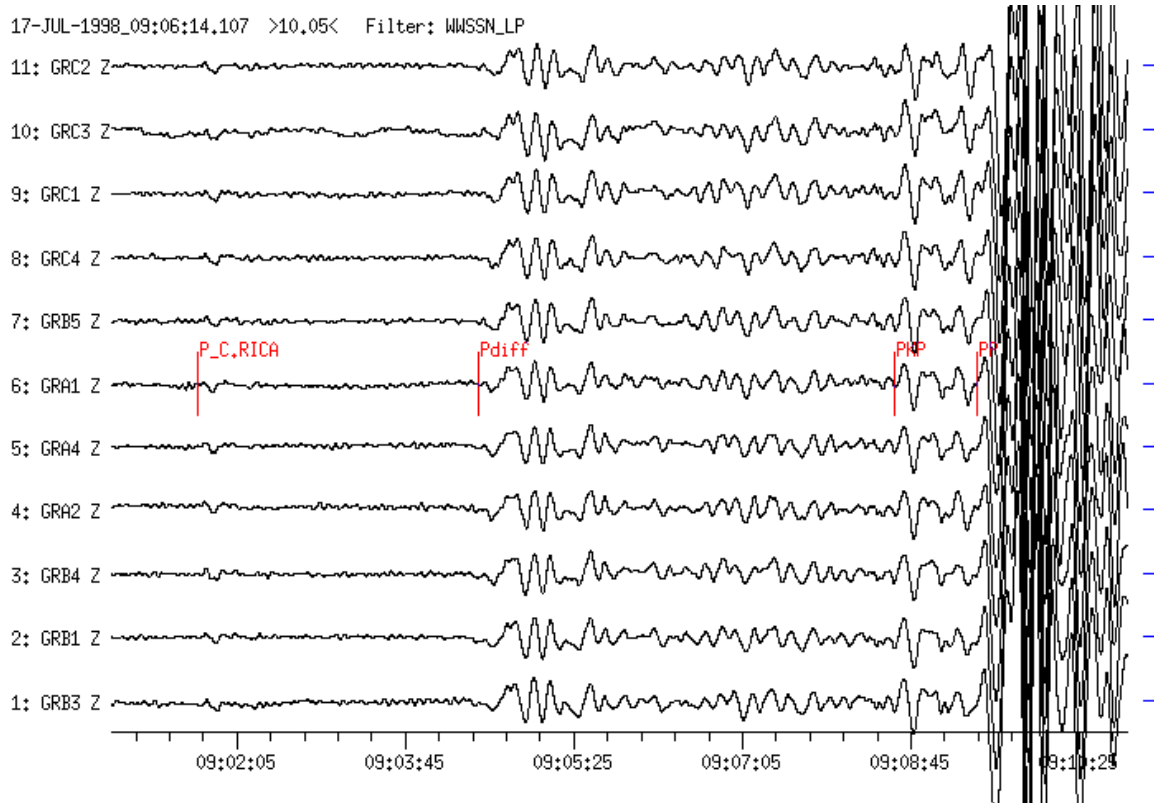
**Figure 4a** Long-period filtered three-component seismogram (SRO-LP simulation) recorded at station GRA1 ( $D = 117.5^\circ$ ). The horizontal components (RT) are rotated with R into the epicentral direction. Phases Pdif, PP and a strong SP are visible on the vertical component. While the phases SKS, SKKS and PS, polarized in the vertical propagation plane, are strong on the radial (R) component Sdif, SS and SSS are strong on the transverse (T) component.



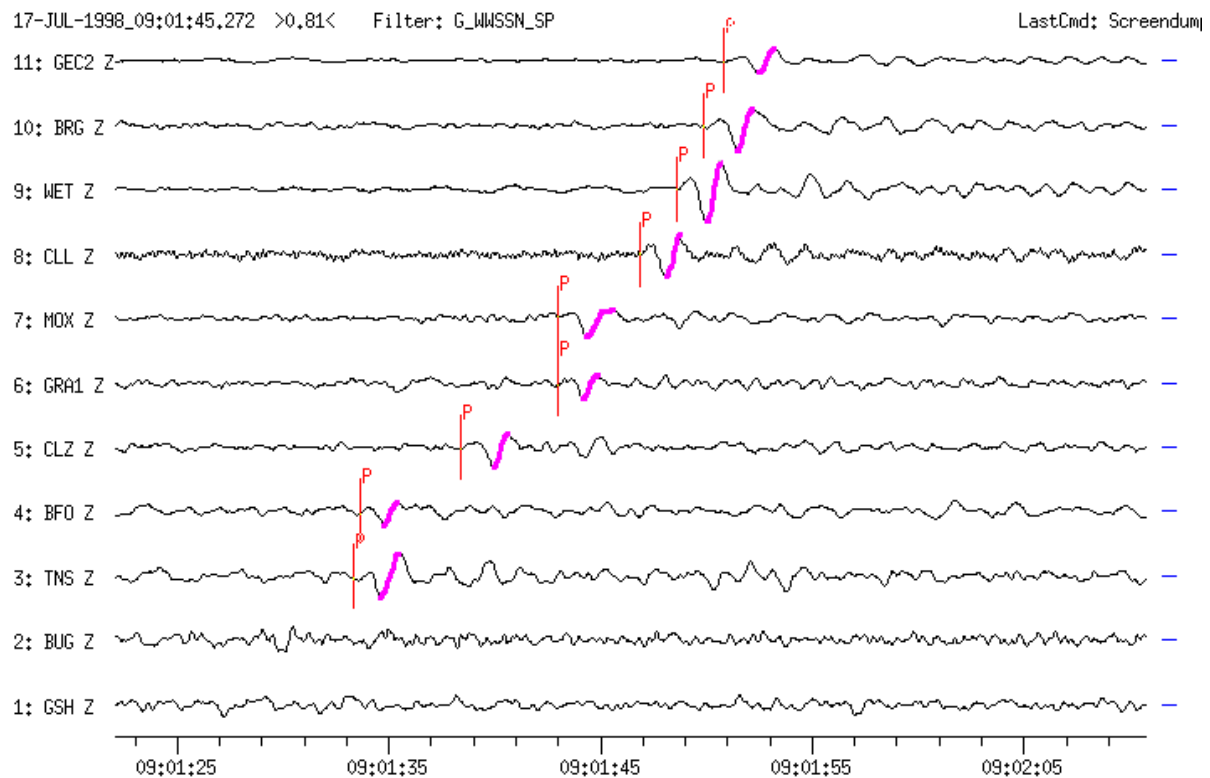
**Figure 4b** Long-period filtered vertical-component seismograms (WWSSN-LP simulation) recorded at 11 GRSN- and GRF-stations. Traces are sorted according to increasing distance ( $D = 115.4^\circ$  to BRG and  $119.9^\circ$  to BFO). Long-period onsets of Pdif and PP are very clear as well as an unidentified phase X ahead of PP. The time differences between X and PP differ from station to station. Probably X results from interfering waves PKPdf and PKiKP. An answer is given in Figure 4c.

Three minutes before the Pdif onset from the earthquake in Papua New Guinea, a small P-wave onset was recorded in the seismograms of GRSN stations from an earthquake in Costa Rica. Its source parameters were:

**USGS NEIC-data: 1998-07-17 OT 08:49:01 8.59N 83.07W h = 33G mb = 5.9;  
(D = 86.3° and BAZ = 278.8° from or GRA1)**



**Figure 4c** Long-period filtered vertical-component seismograms (WWSSN-LP simulation) recorded at 11 GRF-array stations within the distance range  $D = 86.3^\circ$  to  $86.7^\circ$ . Long-period onsets of the P wave from the Costa Rica event appear first, followed by Pdif, PKP and PP from the earthquake in Papua New Guinea. The phase X from Figure 4b could be identified from the GRF-array records as being PKPdf from the Papua New Guinea earthquake with a slowness of about  $2.0 \text{ s}^\circ$ .

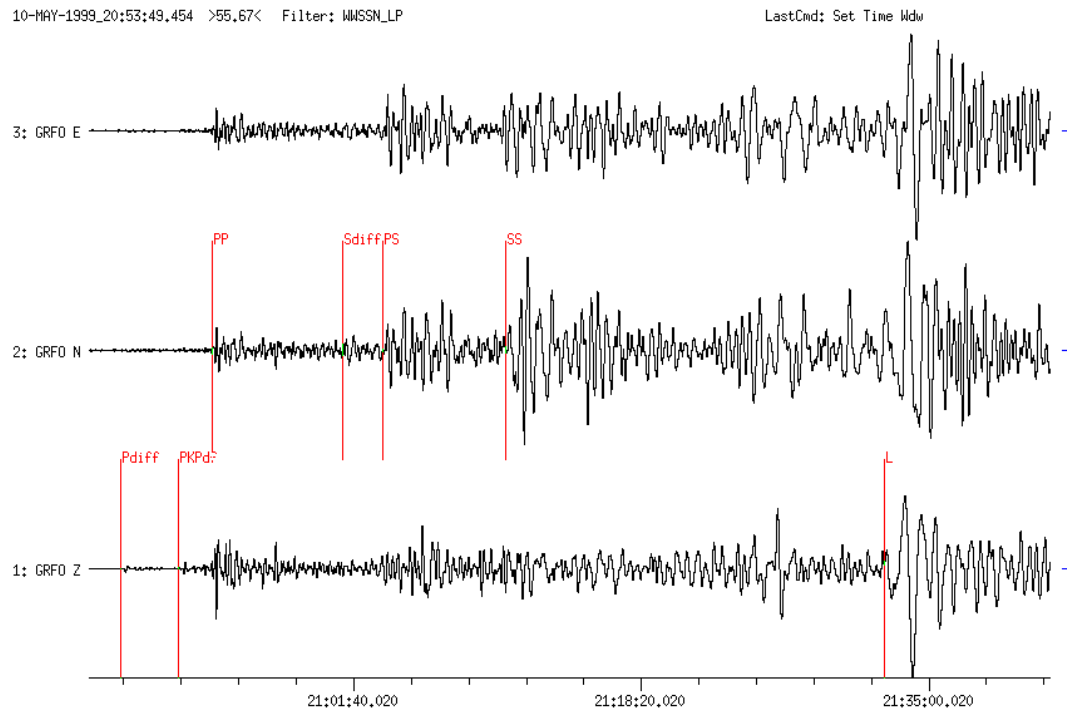


**Figure 4d** Short-period filtered seismograms (WWSSN-SP simulation) recorded from the Costa Rica earthquake at 11 GRSN-stations. Traces are sorted according to increasing epicentral distance ( $D = 83.1^\circ$  to GSH and  $88.0^\circ$  to GEC2). The station-network allows to separate both events by way of slowness and azimuth determination. The estimated body-wave magnitudes  $m_b$  vary for the GRSN-stations between values below 5.0 (BUG, GSH) and 5.4 for station WET.

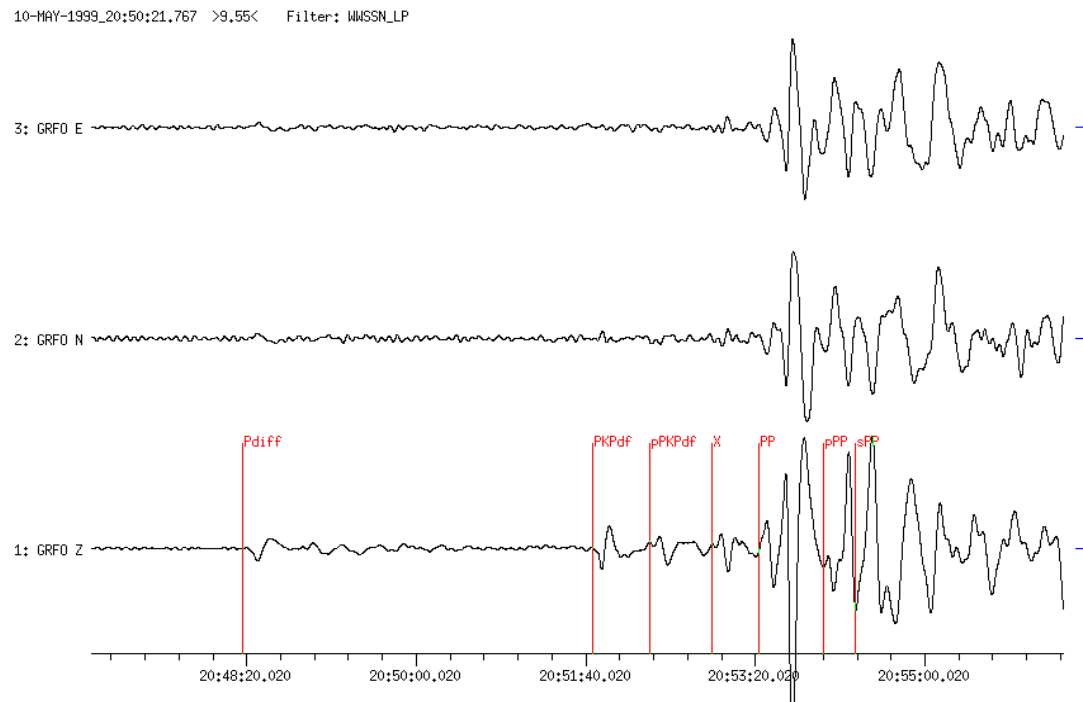
### Example 5: Earthquake in the region of New Britain, P.N.G.

**USGS NEIC-data: 1999-05-10 20:33:02.1 5.173S 150.915E  $h = 137$ D  $m_b = 6.5$ ;  
( $D = 124^\circ$  and  $BAZ = 51^\circ$  from GRFO)**

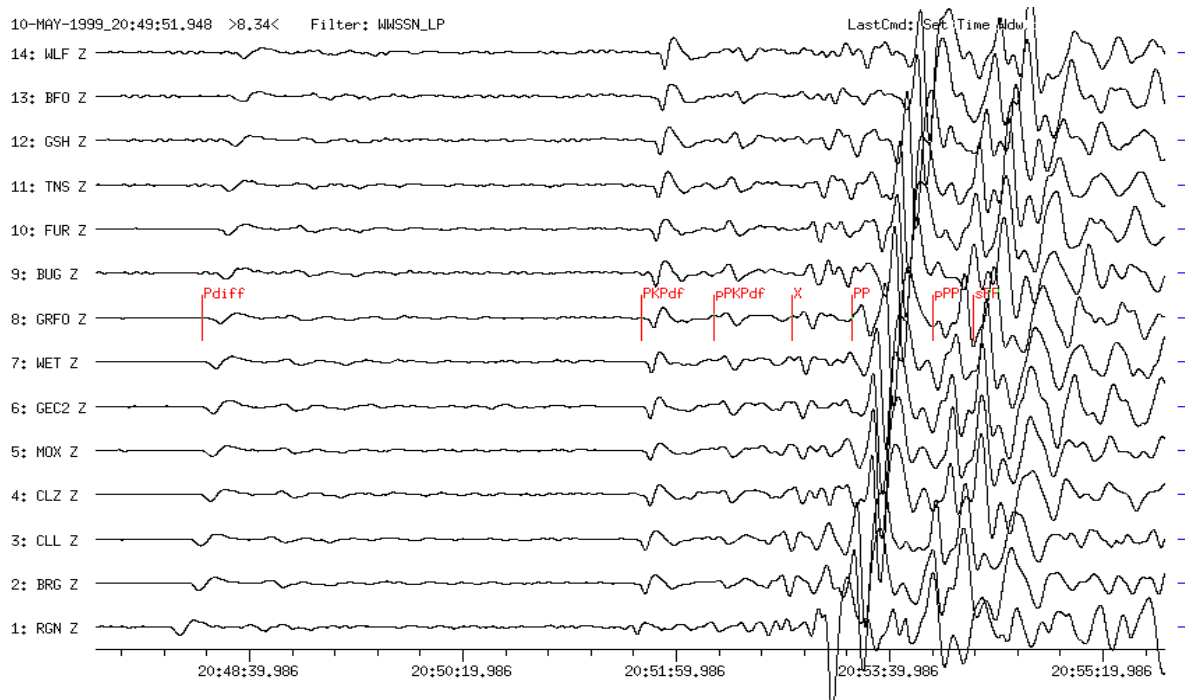
This magnitude 6.5 earthquake, recorded at GRFO and at the GRSN-network, shows the phases  $P_{dif}$ , PKP, PP,  $S_{dif}$ , PS, SS, depth phases and – because of a focal depth of 137 km – weak surface waves. A rare example is the well recorded phase  $P'P'P'P' = PKPPKPPKPPKPP$  in Figure 3e below. The ray paths for the identified phases of this event are shown in Fig. 11.60. For animation of ray propagation and seismogram formation see CD-ROM attached to Volume 2 and complementary explanations given in IS 11.3.



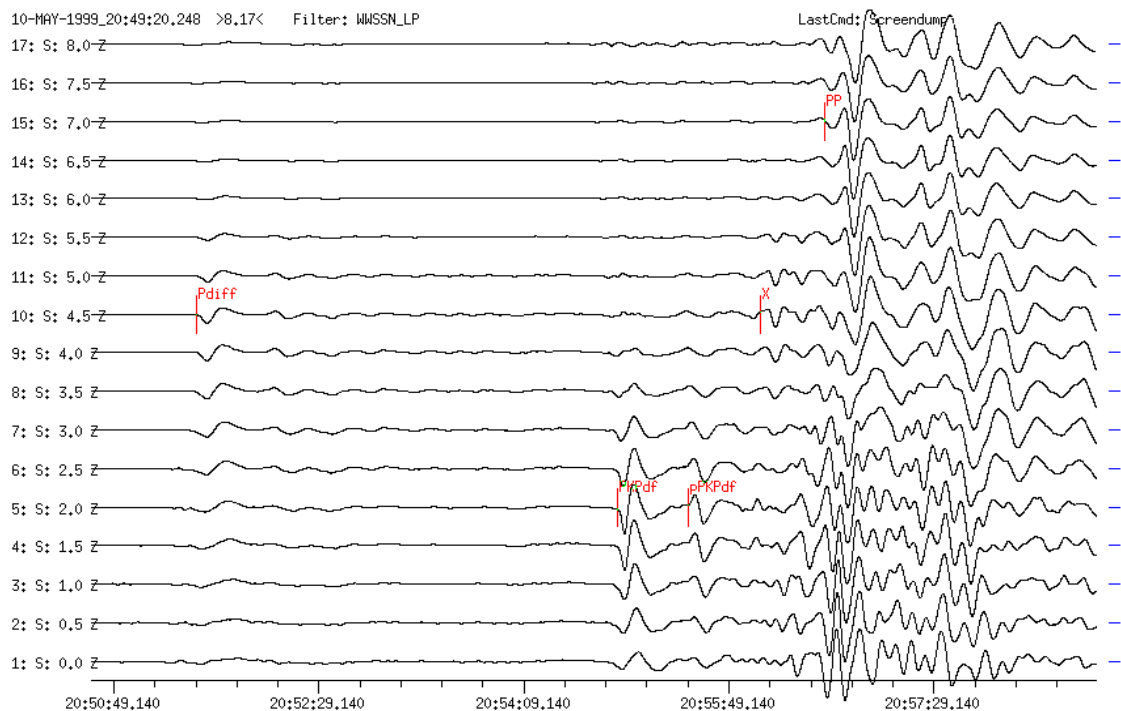
**Figure 5a** Long-period filtered three-component seismogram (WWSSN-LP simulation) recorded at station GRFO ( $D = 124^\circ$ ). The length of the record is about 1 hour. Phases Pdif, PKP, strong PP, Sdif, strong PS, strong SS and – according to the large focal depth of the earthquake ( $h = 137$  km) – relatively weak surface waves are recorded.



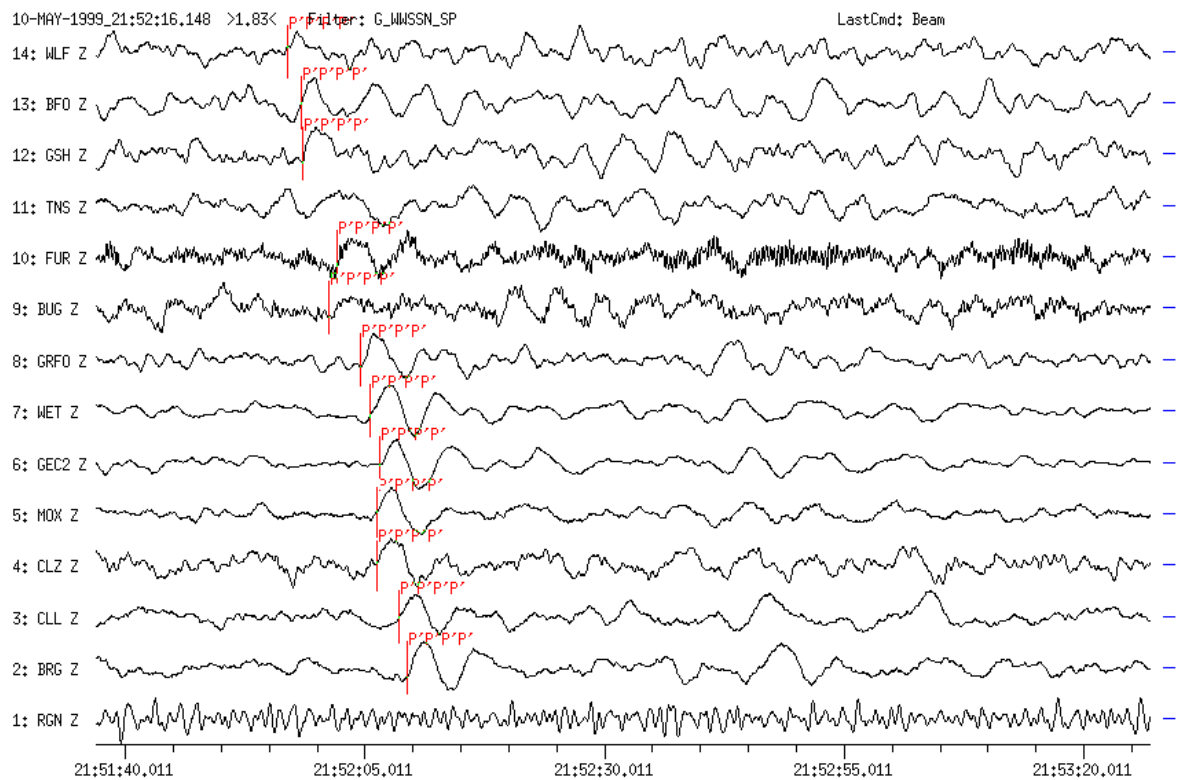
**Figure 5b** Shown are the first 10 minutes from the seismogram in Figure 5a. The higher time resolution separates well the phases Pdif, PKPdf, pPKPdf, an unidentified phase X (25s ahead of PP), PP and the depth phases pPP and sPP.



**Figure 5c** Long-period Z-component records (WWSSN-LP simulation) at 14 GRSN-, GEOFON- and GERESS-stations within the distance range  $D = 120.1^\circ$  (RGN) to  $126.6^\circ$  (WLF). This record section was used to perform the vespagram analysis (see Figure 5d).



**Figure 5d** Vespagram analysis for vertical-components was used for slowness determination and phase identification. Phases were identified according to slowness values and travel times. The seismogram analysis program SHM allows to choose slowness steps (here in increments of  $0.5 \text{ s}^\circ$ ). The analysis yields slowness values of  $4.5 \text{ s}^\circ$  for Pdif,  $2.0 \text{ s}^\circ$  for PKPdf and pPKPdf,  $7.0 \text{ s}^\circ$  for PP and for the phase X a value that would correspond to Pdif. The ray path could not be identified.



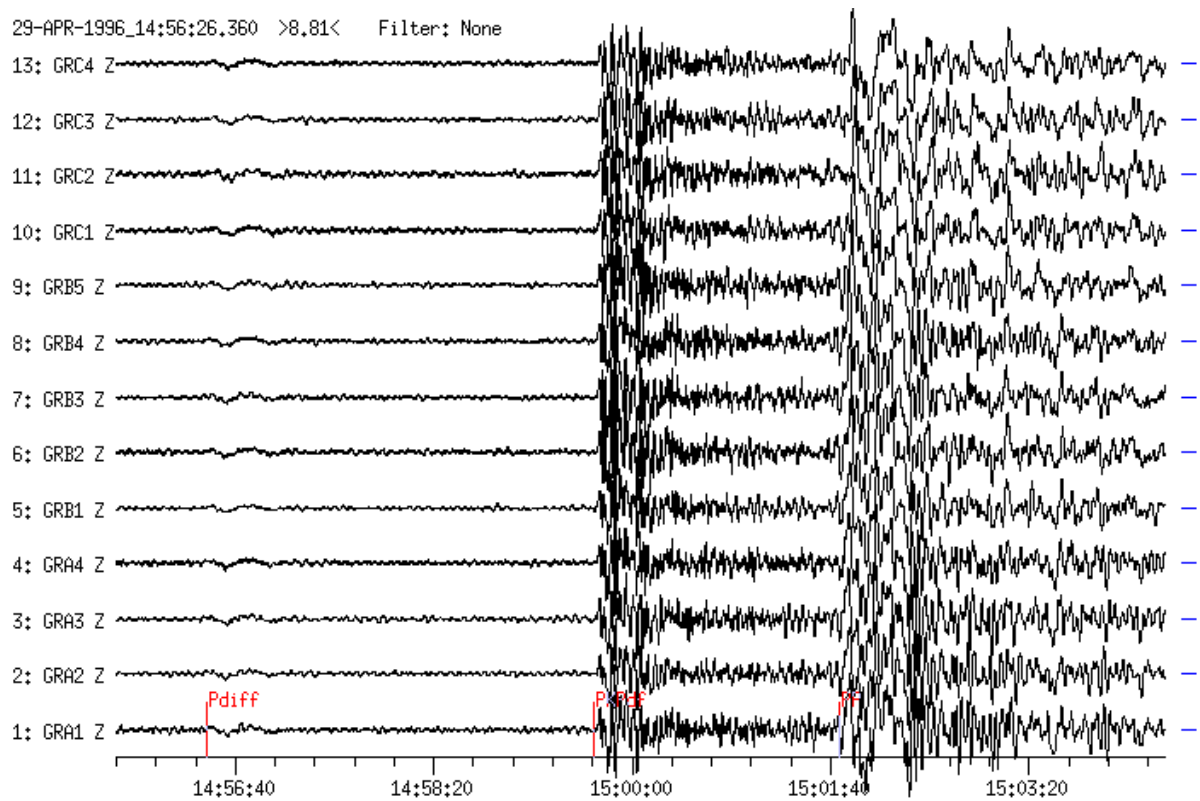
**Figure 5e** The network records show the rare phase 4P' (PKPPKPPKPPKP) with slowness  $S = 2.5 \text{ s}^\circ$ . It arrives 79 minutes after the origin time from the opposite azimuth direction (BAZ = 239°).

### Example 6: Earthquake in the region of Solomon Islands

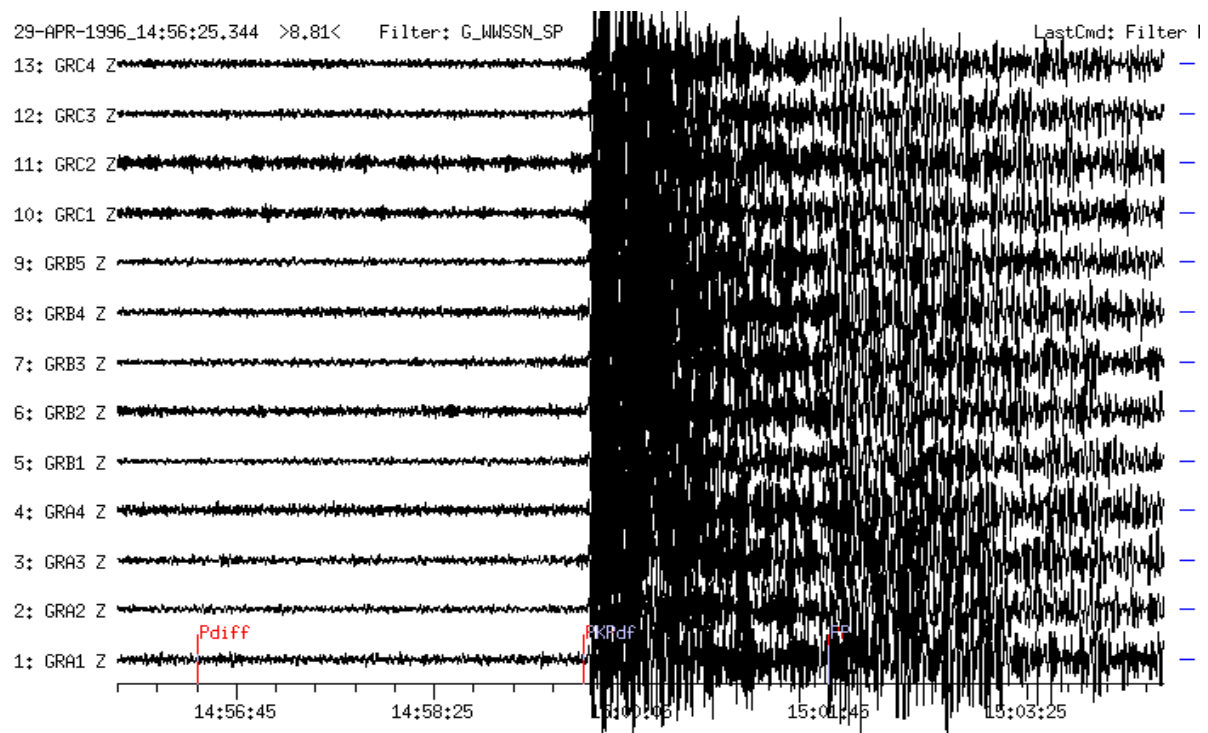
USGS NEIC-data: 1996-04-29 OT 14:40:41.2 6.516S 155.037E h = 44G  
 mb = 6.3 Ms = 7.5;  
 (D = 127.3 ° and BAZ = 47.5° from GRA1)

The example below presents clear Pdif onsets on long-period filtered seismograms as well as an excellent phase 4P' = P'P'P'P' = PKPPKPPKPPKP arriving about 63 min after the first onset. All records were made at the GRF-array (see Fig. 11.3).

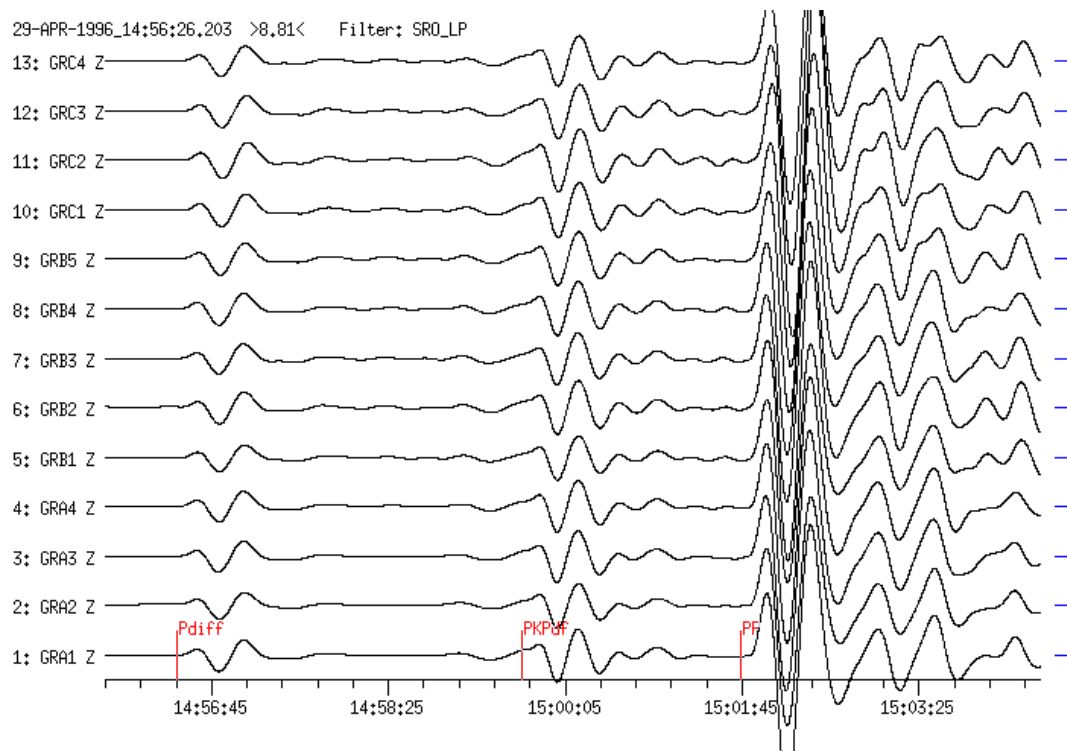




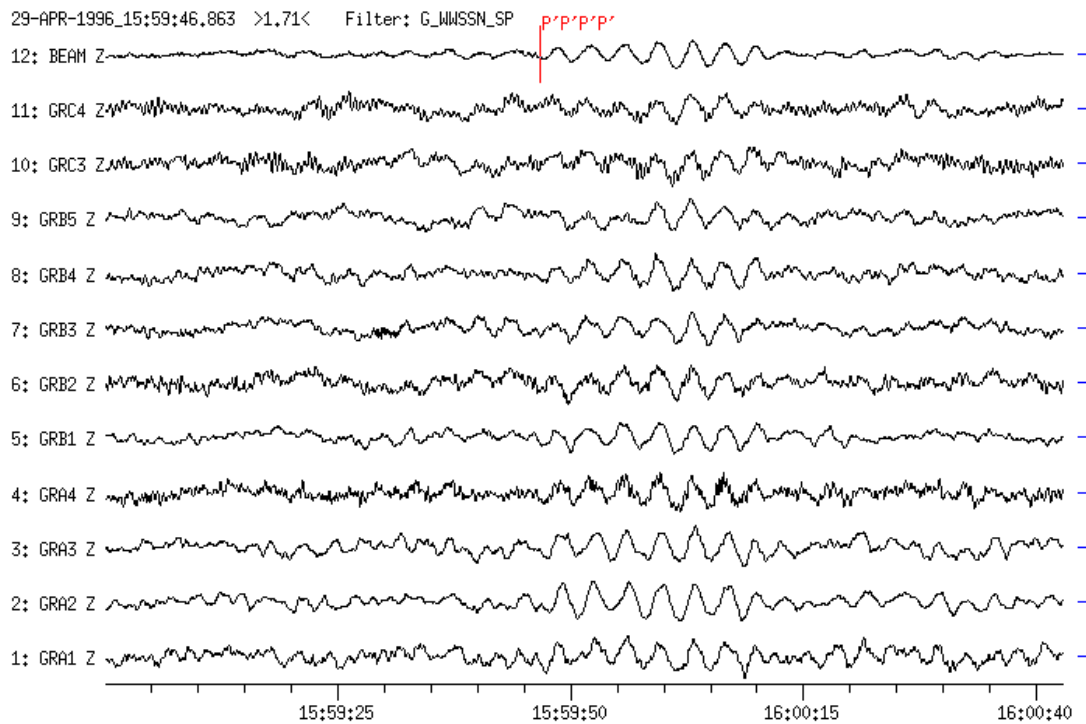
**Figure 6a** Broadband Z-component records at the GRF array within the distance range  $D = 127.2^\circ$  to  $127.8^\circ$  which show, besides the rather long-period Pdif, also PKPdf and PP.



**Figure 6b** The same record as Figure 6a, however, a short-period filter was applied (WWSSN-SP simulation). Pdif is no longer visible above the noise level.



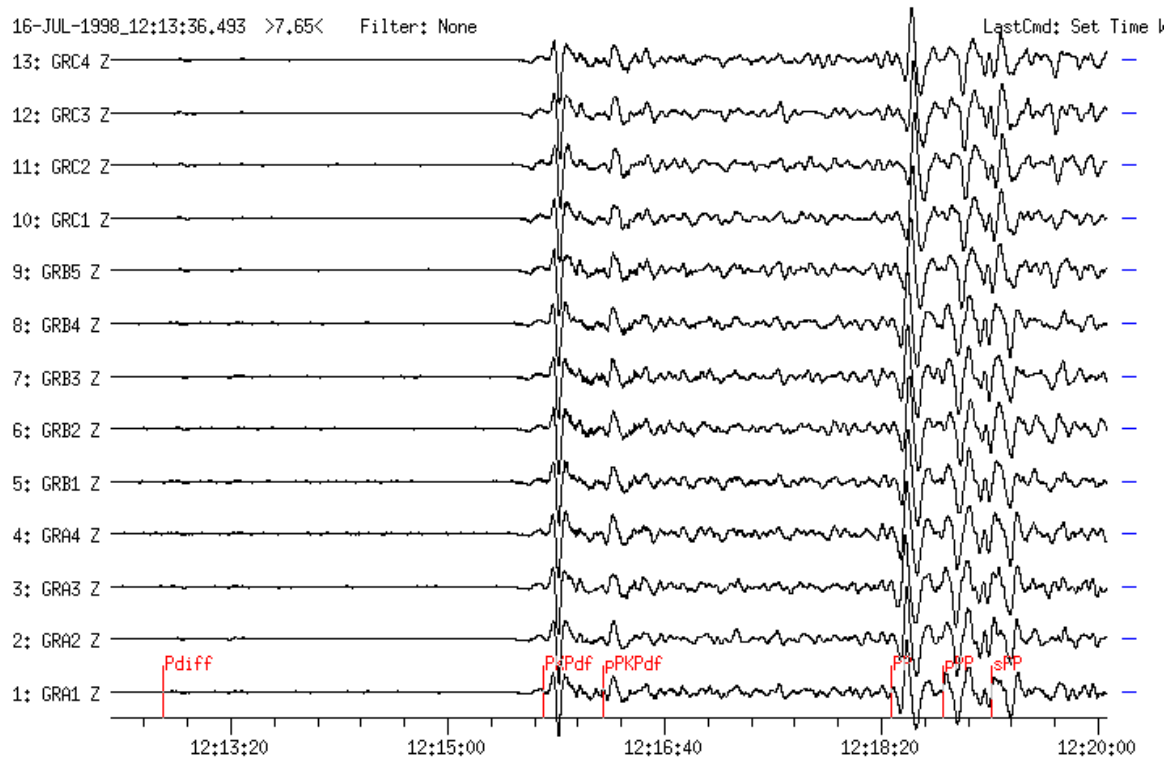
**Figure 6c** The same records as in Figures 6a and b, however, after applying a long-period filter (SRO-LP simulation). A distinct Pdfff appears on the record.



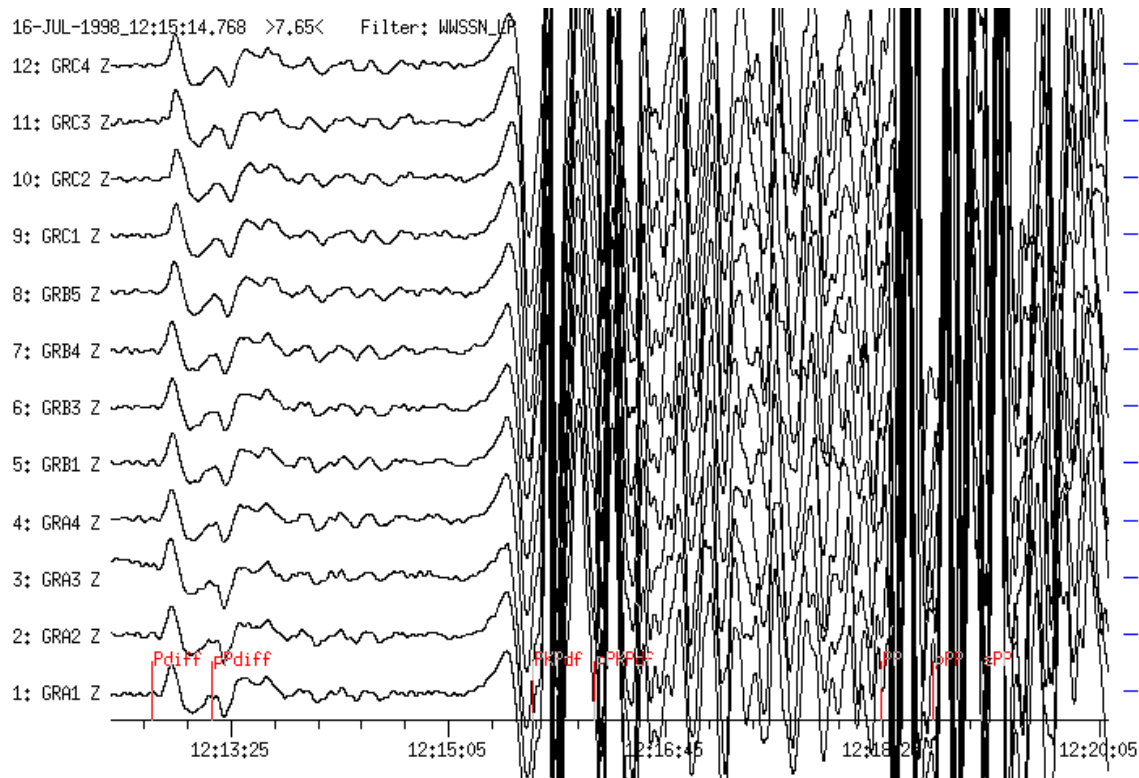
**Figure 6d** Short-period filtered (WWSSN-SP simulation) seismogram. The rare phase 4P' = P'P'P'P' (PKPPKPPKPPK) was clearly recorded at the GRF-array coming from opposite direction (BAZ = 226 and S = 2.56 s/°) about 63 min after the first P onset. Trace No.12 (the sum of all traces = BEAM) shows the best signal-noise-ratio.

## Example 7: Earthquake in Santa Cruz Island

USGS NEIC-data: 1998-07-16 OT 11:56:36 11.1S 165.9E h = 110G;  
(D = 136.2° and BAZ = 37.3° from GRA1).



**Figure 7a** Vertical-component velocity BB seismograms recorded at the GRF-array (distance range between 136.1 and 136.8°). Note the very weak Pdif. The depth phases of PKPdf and PP correspond to a depth of 110 km.



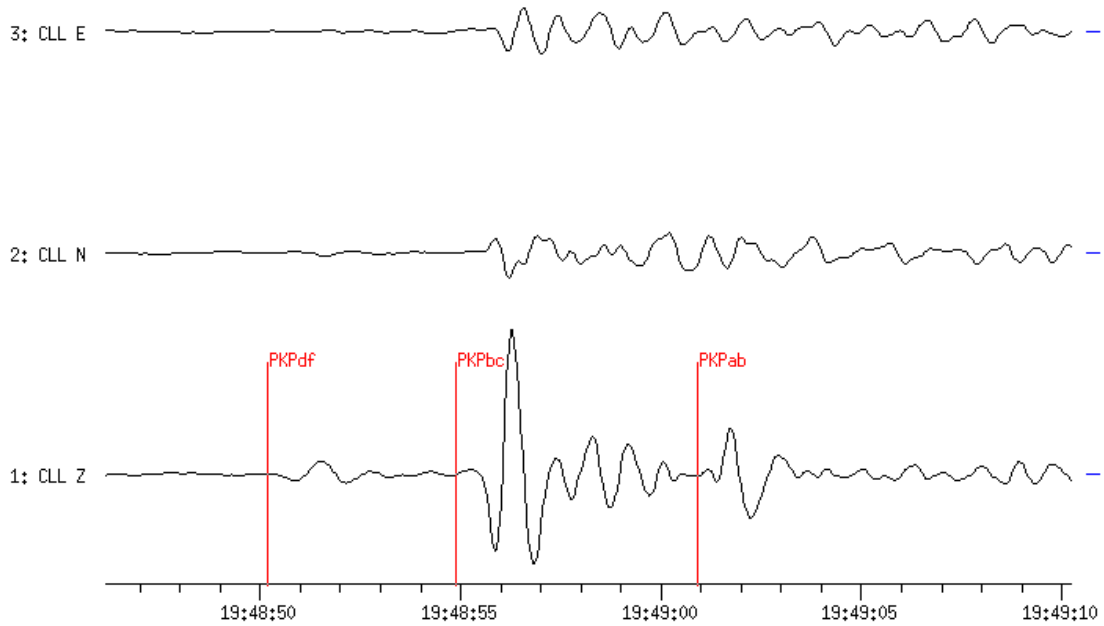
**Figure 7b** The same record as in Figure 7a, however, after long-period filtering (WWSSN-LP simulation). Note the pronounced onsets Pdif and pPdif. Also remarkable is the long-period and slowly emerging onset 24 sec before PKP.

### Example 8: Earthquake in the region of Fiji Islands

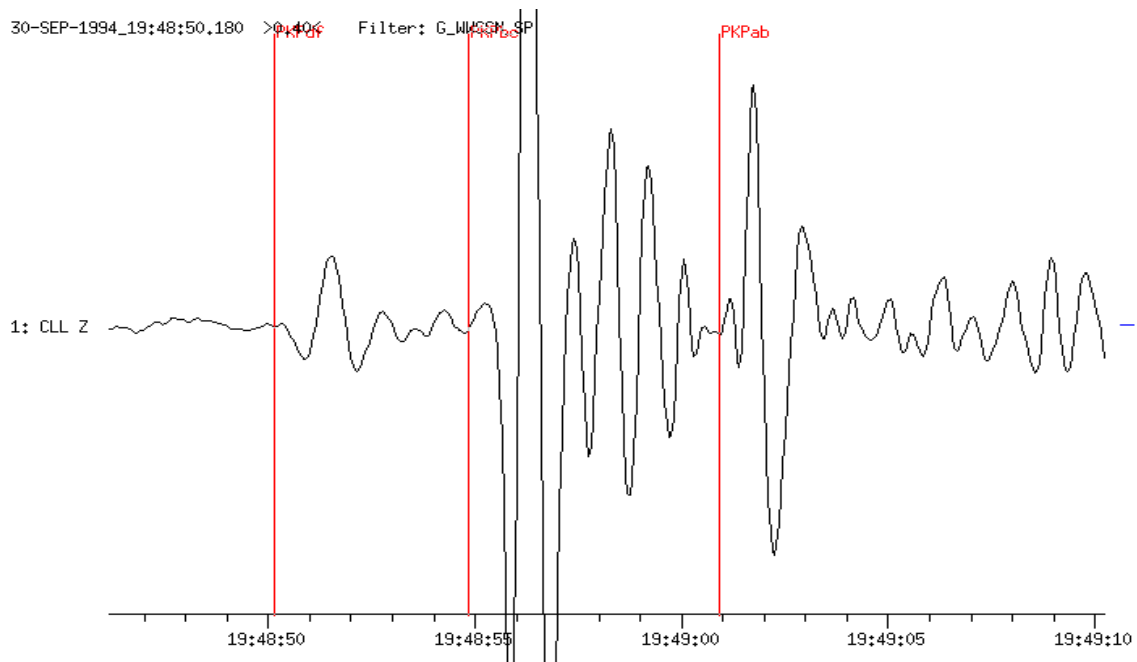
**USGS-QED-data: 1994-09-30 OT 19:30:06.9 21.102S 179.204W h = 500km(G), mb = 5.1;**

The example shows pronounced core phases in the distance range 148°-152°. Short-period P waves reappear in this distance range but with discontinuous travel-time branches. In the region of the caustic, i.e. around 144°, the whole energy of direct longitudinal core phases is concentrated thus forming a strong PKP onset. Its amplitudes are comparable with that of P waves at much shorter distances (D around 50°). At distances beyond the caustic point PKP onsets are separated into individual PKP branches. The energy distribution changes with increasing distance. PKPbc (or PKP1) is the dominant branch just beyond the caustic, up to about 153°. In records of weaker events, PKPbc is often the first visible onset since PKPdf (or PKIKP), preceding PKPbc, is too weak to be observed. With increasing distance near 160° PKPbc vanishes from the record and PKPab (or PKP2) dominates the seismogram. For more details see sub-section 11.5.2.4.

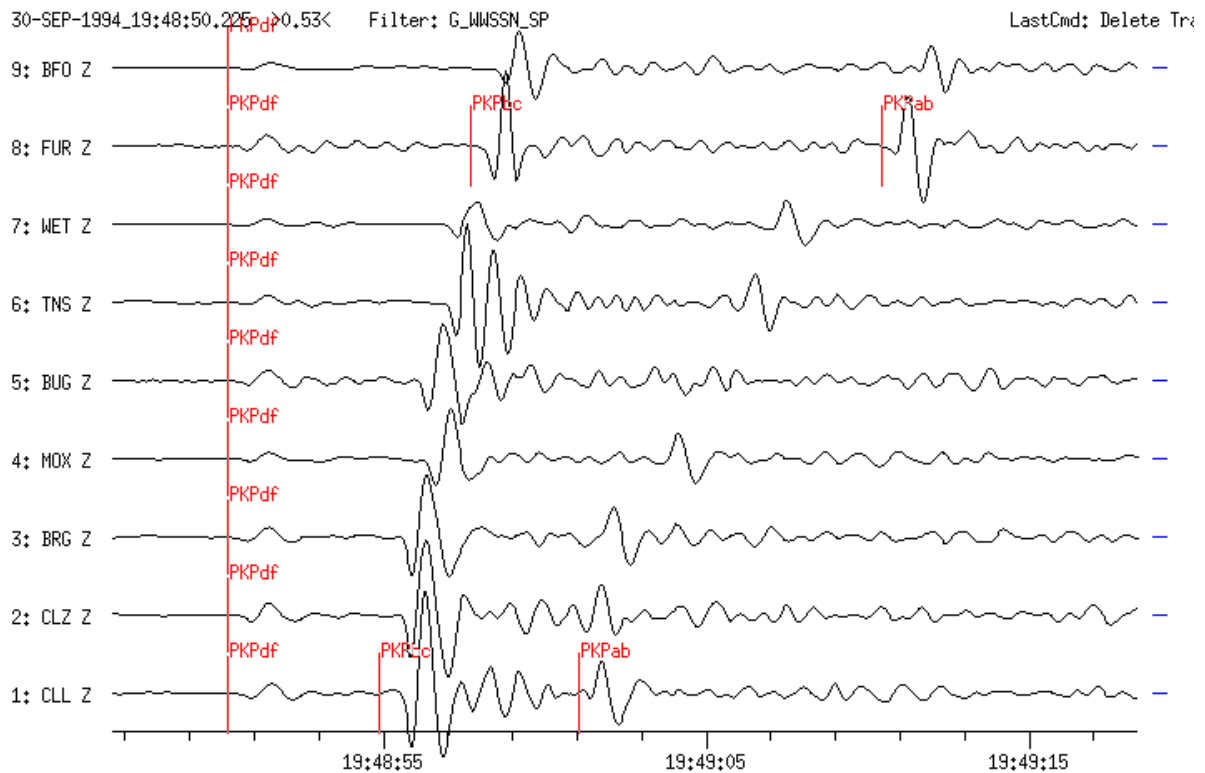
30-SEP-1994\_19:48:50.219 >0.40< Filter: G\_WWSSN\_SP



**Figure 8a** Short-period filtered three-component seismogram (WWSSN-SP simulation) recorded at the broadband station CLL at an epicentral distance of  $D = 148.4^\circ$  ( $BAZ = 22^\circ$ ). Phases PKPdf, PKPbc and PKPab are easy to analyse in this distance range. Travel-time curves for the different branches of these core phases allow to determine the epicentral distance from records of a single station only with an error less than  $2^\circ$ . Note the relatively very small amplitudes on the horizontal components because of the very small (steep) incidence angle at this large distance.



**Figure 8b** Vertical-component record of station CLL, displayed with higher magnification.



**Figure 8c** Short-period filtered seismograms (WWSSN-SP simulation) recorded at 9 GRSN-stations. All traces are time-shifted and aligned for PKPdf and sorted according to increasing distance. The epicentral distance for CLL is  $148.4^\circ$  and for the most distant station BFO  $152.2^\circ$ .

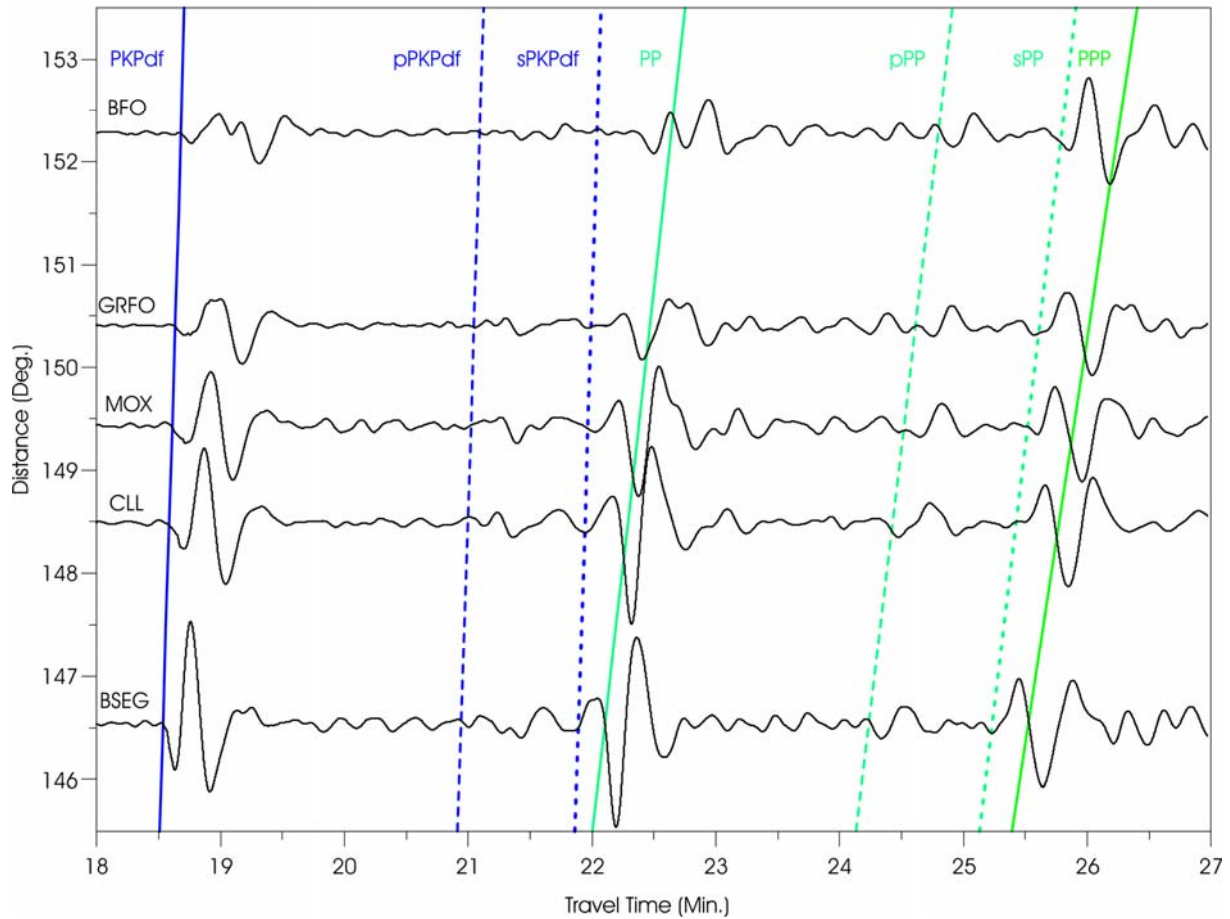
### Example 9: Deep earthquake in the region of Fiji Islands

**USGS NEIC-data: 2000-12-18 OT 01:19:21.6 21.18S 179.12W h = 628 km mb = 6.4;**  
**The epicentral distances of the GRSN stations range between  $146^\circ < D < 153^\circ$ .**

Shown are enlarged cut-outs of the the PKP-wave group from the seismograms presented in Fig. 11.64. The PKP-wave group consists between  $146^\circ < D < 156^\circ (160^\circ)$  of three distinct wave arrivals PKPab, bc and df. Fig. 11.59 shows the respective travel-time curves. The figures below demonstrate the effect of the record filter on the discrimination of these closely spaced wave arrivals and the possibility to pick their onset times. The longer the center period of applied narrowband band-pass filters, the worse is the discrimination and onset-time picking for these three core-phase arrivals. Therefore, PKP phases are best picked in SP records, however, the later phases PP and PPP from so distant earthquakes are usually recognizable only in LP or BB records.

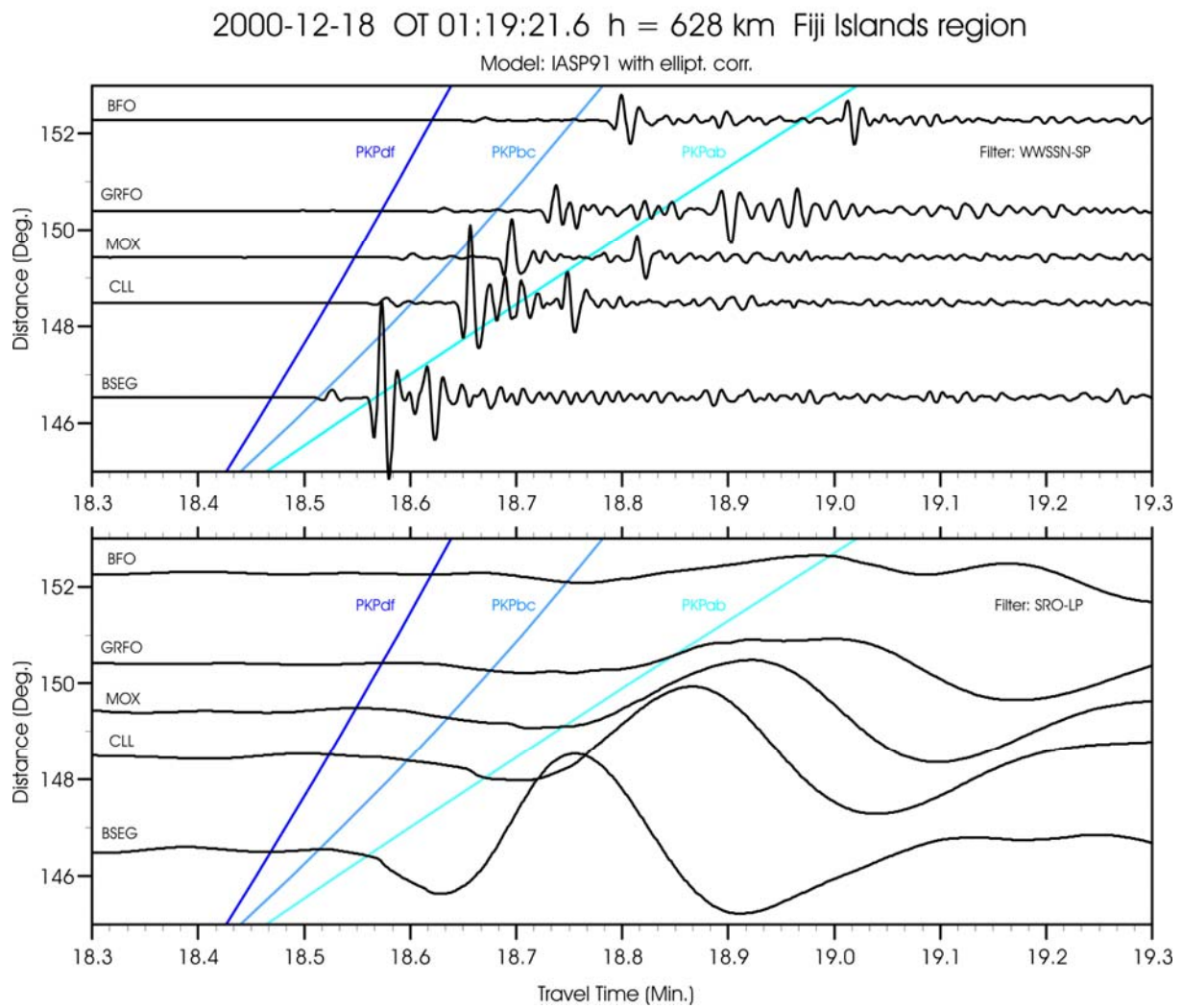
2000-12-18 OT 01:19:21.6 h = 628 km Fiji Islands region

Z-component, SRO-LP



**Figure 9a** Z component long-period (SRO-LP simulated) records of longitudinal wave groups from a deep ( $h = 628$  km) Fiji Islands earthquake at stations of the GRSN. Note the strong onsets of the depth phases sPKPdf and sPP prior to onsets PP and PPP, respectively, which follow closely.

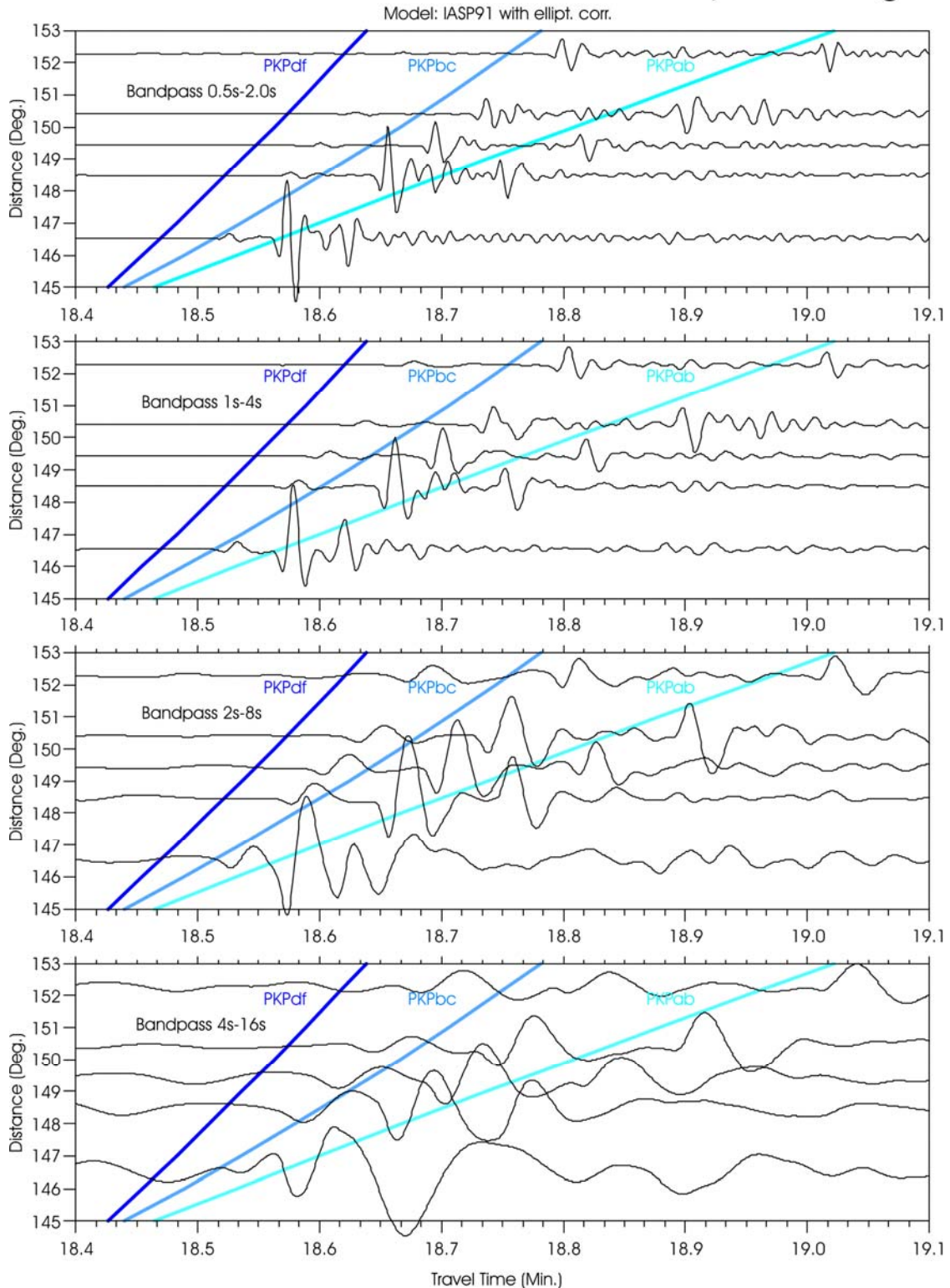




**Figure 9b** Enlarged cut-outs of the PKP-wave group as recorded in WWSSN-SP (upper traces) and SRO-LP filtered seismograms of the GRSN stations. Note that for all three core phases the theoretically expected onset times according to the IASP91 travel-time model are about 2 s too early.



# 2000-12-18 OT 01:19:21.6 h = 628 km Fiji Islands region



**Figure 9c** Enlarged cut-outs of the PKP-wave group of the deep Fiji Islands earthquake of 18 December 2000 recorded at stations of the GRSN. Shown are, from top to bottom, the bandpass filtered BB records of two octaves bandwidth with center periods around 1s, 2s, 4s and 8 s. For center periods larger than 3s the separation of the different PKP-wave arrivals and recognition of their onset times becomes less and less clear. Note that according to the IASP91 travel-time model the three core phases should arrive about three seconds earlier.

| Topic       | Record examples of underground nuclear explosions                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|-------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Compiled by | <p><b>Klaus Klinge</b> (formerly Federal Institute for Geosciences and Natural Resources, 30655 Hannover, Germany);<br/> E-mail: <a href="mailto:klaus.klinge@googlemail.com">klaus.klinge@googlemail.com</a></p> <p><b>Johannes Schweitzer</b>, NORSAR, P.O.Box 53, N-2027 Kjeller, Norway,<br/> Fax: +47-63818719, E-mail: <a href="mailto:johannes.schweitzer@norsar.no">johannes.schweitzer@norsar.no</a></p> <p><b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> |
| Version     | January, 2012; DOI: 10.2312/GFZ.NMSOP-2_DS_11.4                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |

## 1 Introduction

Nuclear explosions and in particular underground nuclear explosions had a dramatic influence on the development of modern seismology. All nuclear explosions are sources of elastic waves travelling over long distances in the Earth's atmosphere, in the Earth's oceans, and through the solid Earth, depending on the medium in which the explosion was conducted. These side effects of such explosions are well known and were used since the beginning of the nuclear age as tools to monitor the nuclear testing activity of foreign states (for details about today's monitoring activities see Chapter 17).

The development of seismic arrays (see Chapter 9), the homogenization of seismic instrumentation and the installation of global networks since the 1960s were triggered and financially supported by the wish to globally monitor test activities of nuclear underground explosions. Theoretical seismology has strongly benefited by the demand for more knowledge about the propagation and the attenuation of seismic waves in order to understand the differences between natural and artificial seismic sources.

Data from nuclear tests have been an important data source for numerous studies to investigate the 1D and 3D structure of the Earth's interior. Since, in contrast to earthquakes, the exact location and shot time of these explosions are often precisely known, they are very useful tools to investigate the structure of the Earth with elastic waves (see Ground Truth events in IS 8.6). Already the first test of a nuclear device had been used to investigate the crustal structure (Gutenberg, 1946) and since then the number of published studies using nuclear test observations to study in particular the P-wave velocity structure within the Earth is enormous.

Underground nuclear explosions (UNEs) are usually tests of nuclear weapon devices and performed at (known) test sites. However, some explosions were made outside of these recognized test sites for making cavities, moving large ground masses to redirect rivers or for conducting long-range seismic profiles. These underground explosions are per definition called Peaceful Nuclear Explosions (PNEs). Source information and waveform data from UNEs and PNEs can be retrieved from the open DTRA (<http://www.rdss.info/> and there go to Research Databases / Nuclear Explosions).

Thankfully, underground nuclear explosion activity has decreased in the last decades, but we think it is important to document some typical features of nuclear underground explosions as recorded at seismic stations.

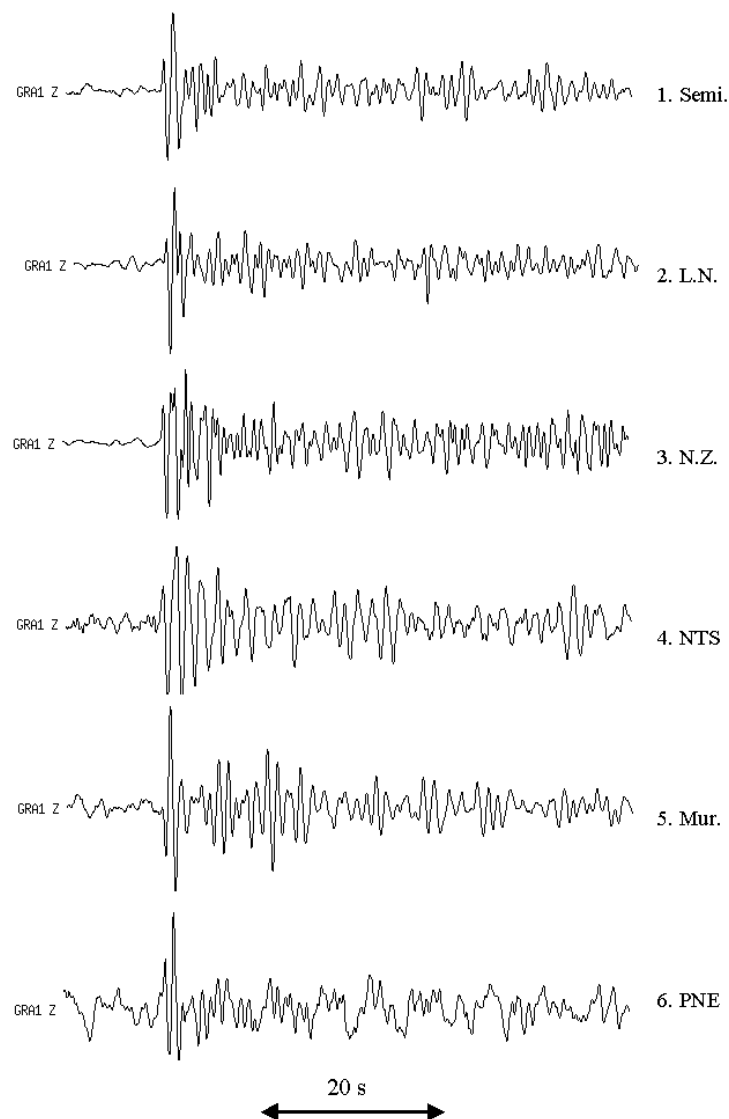
## 2 Record examples of underground nuclear explosions between 1978 and 1993 within the teleseismic distance range

Below, seismic records of underground nuclear explosions (UNEs) at 5 weapon test sites and from a peaceful nuclear explosion (PNE) are shown. All records were made in the teleseismic distance range  $D > 30^\circ$  by vertical-component seismographs at station GRA1 of the Gräfenberg broadband array in Germany. The original BB records were filtered in the short-period range (WWSSN-SP simulation filter). The time scale is given below the records. The amplitudes have been normalized and the records are presented in Figure 1 in the following order:

| No. | Date       | Time        | Latitude [deg] | Longitude [deg] | Depth [km] | $m_b$ | Location (Code Name)         | D [deg] | Reference              |
|-----|------------|-------------|----------------|-----------------|------------|-------|------------------------------|---------|------------------------|
| 1   | 1988-12-17 | 04:18:09.2  | 49.879         | 78.924          | 0          | 5.9   | Semipalatinsk                | 42.34   | UK AWE (1990)          |
| 2   | 1993-10-05 | 01:59:56.68 | 41.6322        | 88.6886         | 0          | 5.9   | Lop Nor                      | 52.49   | ISC (2001)             |
| 3   | 1988-12-04 | 05:19:53.3  | 73.3660        | 55.0010         | 0          | 5.9   | Novaya Zemlya                | 30.32   | Richards (2000)        |
| 4   | 1988-10-13 | 14:00:00.08 | 37.0890        | -116.0493       | 0          | 5.9   | Nevada Test Site (Dalhart)   | 81.83   | ISC (2001)             |
| 5   | 1987-11-19 | 16:31:00.2  | -21.845        | -138.941        | 0          | 5.7   | Mururoa                      | 143.58  | UK AWE (1993)          |
| 6   | 1978-10-08 | 00:00:00.0  | 61.55          | 112.85          | 1.545      | 5.2   | PNE, USSR / Siberia (Vyatka) | 52.78   | Sultanov et al. (1999) |

D is the distance to the reference site of the Gräfenberg Array GRA1. The estimated yields for these explosions range between approximately 20 and 150 kt TNT for the weapon tests and is about 15 kt TNT for the PNE (No. 6).

With the exception of the Mururoa test all other records show a clear positive (compressional) first arrival. This is what one expects from an explosion source at all distances and azimuths if sufficient recording bandwidth and signal-to-noise ratio (SNR) is provided (see Chapter 4, Figs. 4.9 ??? and 4.10 ??? and Figures 2 and 3 below). For the Mururoa test the waveforms are influenced by the PKP caustic (Hilbert Transformation of the signal!) in the distance range around  $D = 145^\circ$  (see Chapter 2, section 2.5.4.3 ??, Figs. 2.34 and 2.35 ??) and therefore the onset polarity cannot be read reliably. Note the remarkable differences in P waveforms, which are rather short and simple for events No. 1, 2, 5 and 6 but much more complex and longer for events No. 3 and 4. This is mainly due to the specific geology and/or complex topography at the test sites in Nevada and on Novaya Zemlya. Additionally, for event No. 3, we observe also later P energy, about 5 s after the first onset, at an epicentral distance of about  $30^\circ$  due to the seismic energy refracted, reflected and scattered in the upper mantle (see Figure 4 and Chapter 2, Fig. 2.29 ??) and for event No. 5 the distinguished onsets, about 5 and 10 s after the first onset are later PKP-type onsets (see Schlittenhardt, 1996).



**Figure 1** Short-period filtered seismograms (WWSSN-SP simulation filter) recorded at station GRA1 of the Gräfenberg broadband array in Germany from underground nuclear explosions in six different areas. For source parameters and distances to GRA1 see the list above.

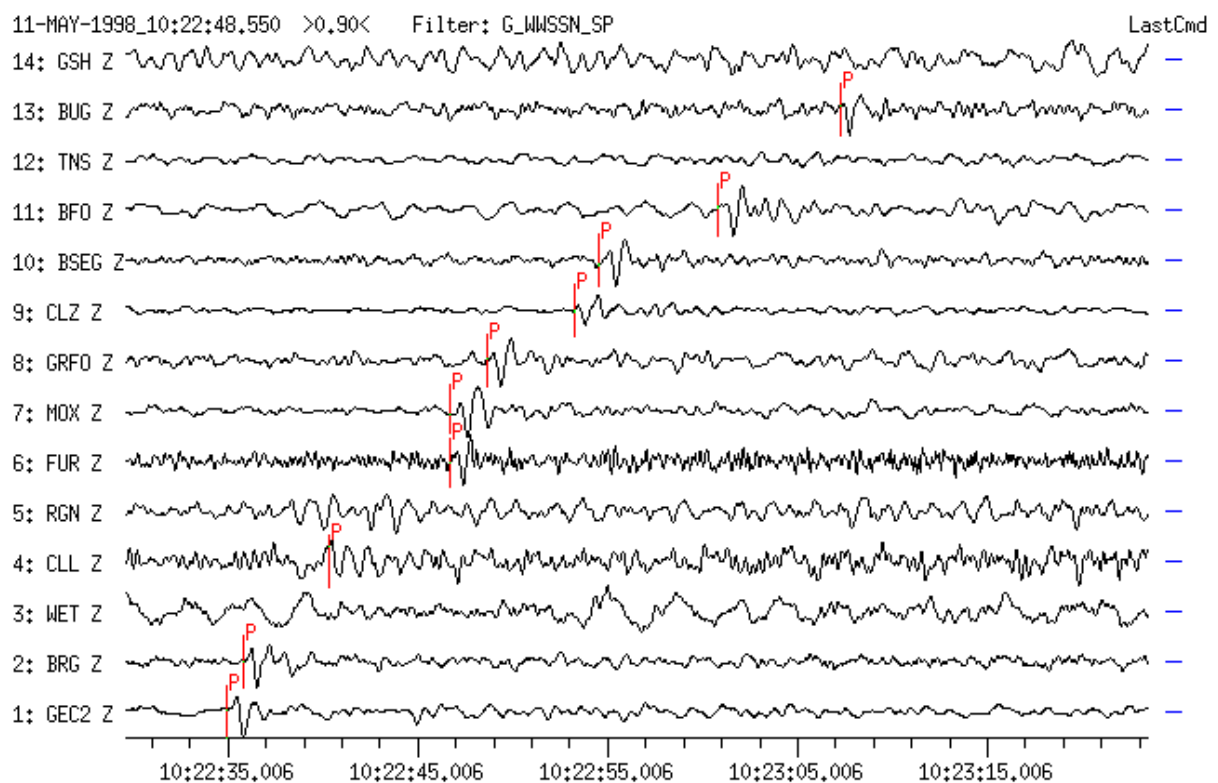
### 3 Records of the nuclear-weapon tests of India and Pakistan

Figures 2 and 3 show vertical component records of broadband stations of the German Regional Seismography Network (GRSN) from the underground nuclear weapons tests of India and Pakistan in 1998. All records were filtered narrowband in the short-period range around 1 Hz and sorted according to the epicentral distance. The source parameters are:

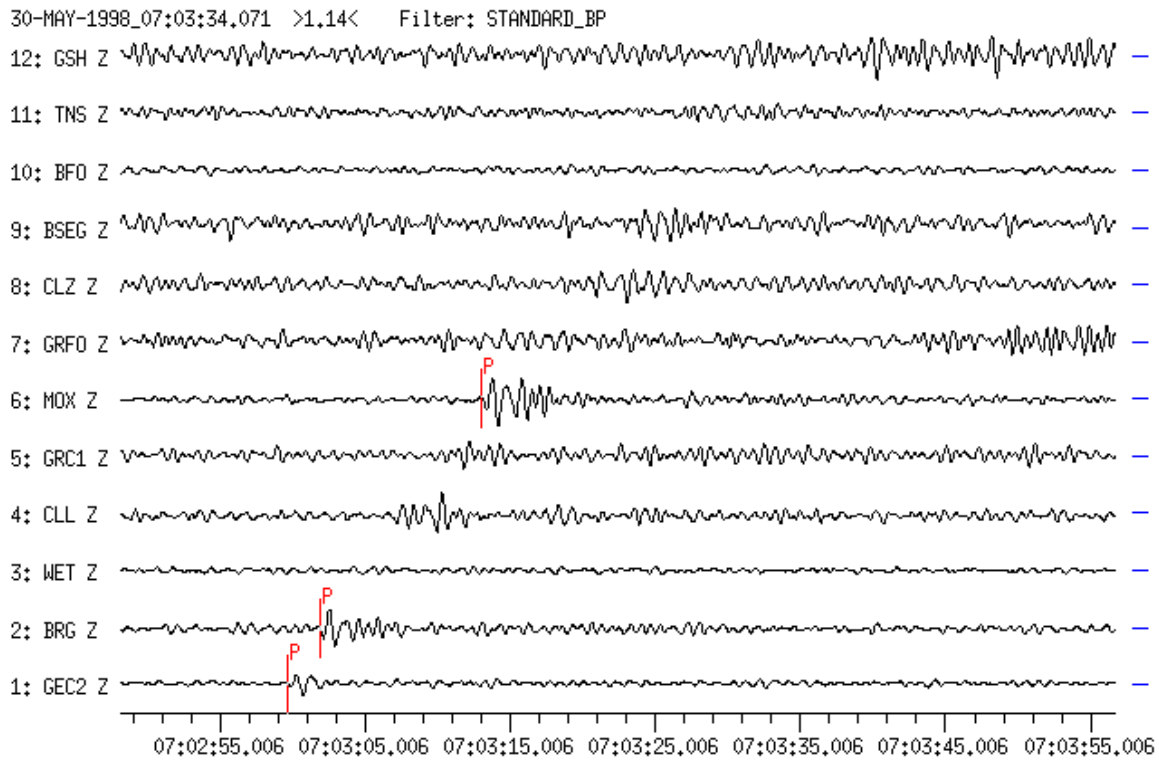
India: 1998-11-05, 10:13:44.2, 27.0780°N, 71.7190°E, depth 0 km, mb 5.2 (Barker et al., 1999) with epicentral distance and backazimuth from GRFO:  $D = 50.99^\circ$  and  $BAZ = 92.94^\circ$ .

Pakistan: 1998-05-30, 06:54:54.87, 28.4434°N, 63.7375°E, depth 0 km; mb 4.7 (ISC, 2001) with epicentral distance and backazimuth from GRFO:  $D = 44.91^\circ$  and  $BAZ = 98.12^\circ$ .

The P-wave onsets are generally simple. However, they are masked by noise at the more noisy stations. Surface-wave amplitudes were very weak and could not be analyzed at this large distance.



**Figure 2** Records of the Indian underground nuclear test on 11 May 1998. For source parameters see above. The broadband records were filtered with the WWSSN-SP response. Typical for explosions are the compressional first onset polarities at stations with high SNR.



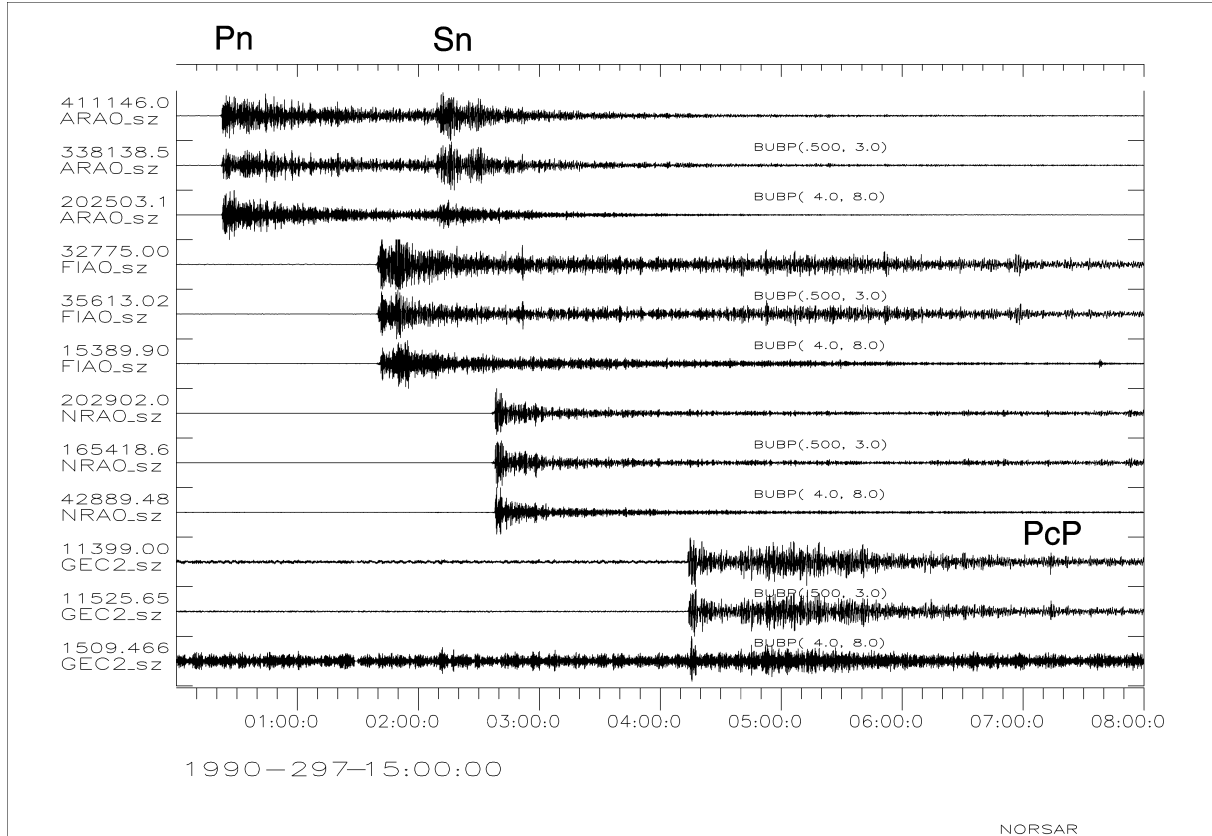
**Figure 3** Records of the Pakistani underground nuclear test of 30 May 1998. For source parameters see above. The relatively small amplitudes of this 0.5 magnitude units smaller explosion are disturbed by high noise levels at the stations and the energy of an aftershock sequence of an earthquake in the Afghanistan-Tajikistan border region (see, e.g., Kværna et al., 2002). Therefore, the broadband records were bandpass filtered between 0.8 and 1.2 Hz, which has a narrower (about 0.5 octaves) bandwidth than the standard WWSSN-SP filter (about 1 octave) used for the Indian explosion (Fig. 2). It results in a better visibility of the signal (higher SNR). However, the signal onset is so emergent that it is not possible to read reliable first motion polarities. Note the apparently negative first onsets at stations BRG and MOX! For related discussions see section 4.2 in Chapter 4 of this Manual.

#### 4 Records of underground nuclear explosions at regional distances ( $7^\circ < D \leq 30^\circ$ )

At shorter distances ( $D \leq 30^\circ$ ) records from underground nuclear explosions (UNEs) still contain a rather large amount of high-frequency energy. This is mainly due to the difference in the source process as compared to an earthquakes (see Fig. 3.5 ??? in Chapter 3 and the related discussions in Section 3.1.1.3 ?? of this Manual). Two examples are shown below.

#### 4.1 UNE at the Northern Novaya Zemlya Test Site

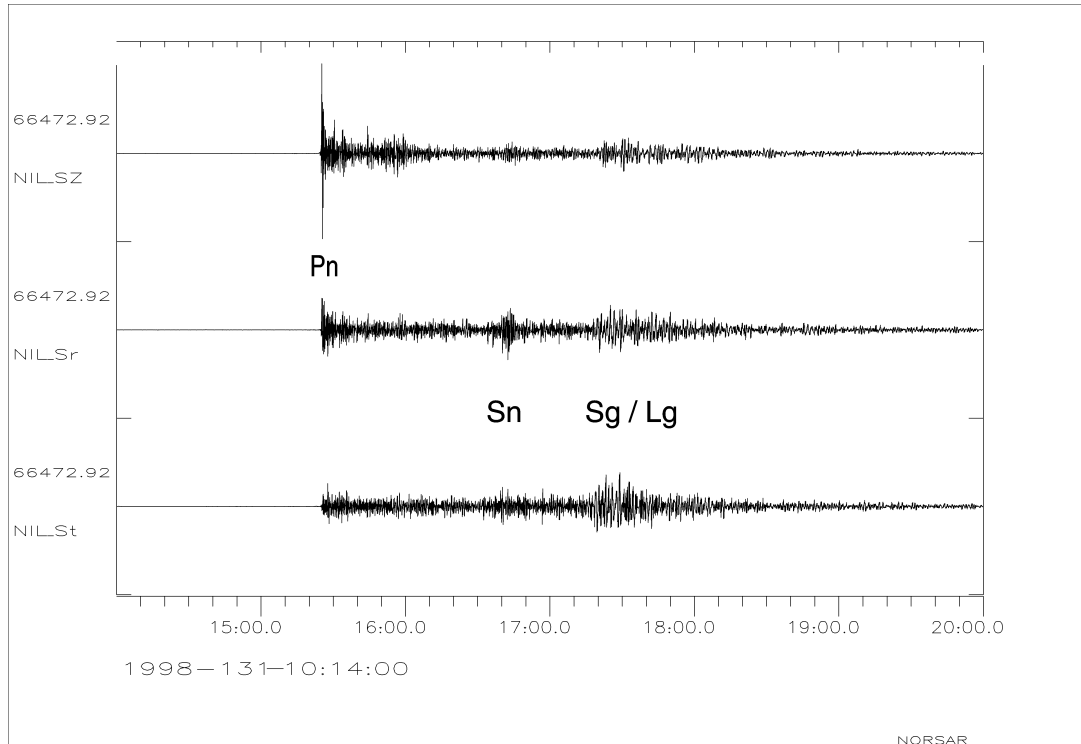
Source data: 1990-10-24, 14:57:58.5, 73.3310°N, 54.7570°E, depth 0 km, mb 5.7 (Richards, 2000). The distances to the recording stations shown in Fig. 4 are: D = 9.99° (ARCES), 15.98° (FINES), 20.41° (NORES) and 30.40° (GERES).



**Figure 4** Records of the key stations ARA0, FIA0, NRA0 and GEC2 of the small aperture short-period arrays ARCES, FINES, NORES and GERES from the Novaya Zemlya test of 24 October 1990. These arrays are specialized for regional signals and ARCES, FINES and GERES are part of the International Monitoring System (IMS) under the Comprehensive Nuclear-Test-Ban Treaty Organization (CTBTO) in Vienna (see Chapter 17). Shown are the vertical component records a) unfiltered, b) band-pass filtered between 0.5 and 3 Hz, and c) band-pass filtered between 4 and 8 Hz. Note the relatively strong high-frequency energy that is well developed in the P-wave group up to a distance of  $\sim 30^\circ$  but no longer visible above the level of signal coda for S waves beyond  $10^\circ$  distance. S waves are stronger attenuated than P waves. At GEC2 the core reflection PcP is nicely visible and the P-wave coda is dominated by energy refracted, reflected and scattered in the upper mantle. Each trace is normalized by its maximum amplitude, which is given in digital counts together with the channel name. The time axis is labeled at each minute after 15:00:00 with 10 s between two ticks.

## 4.2 UNE of India in 1998

In Figure 2 we showed data of the Indian UNE of 5 May 1998 as it was observed in Europe. This event could also be observed at regional distances, e.g., at the Pakistani station NIL (Figure 5).



**Figure 5** Z-R-T rotated three-component record at the IRIS/IDA station NIL, Pakistan. The epicentral distance to the Indian test site is about 740 km; the amplitudes are normalized by the largest signal, which is given in counts at each channel name; shown are unfiltered data. Note the strong P- but weak S-wave arrivals. The time axis is labelled for each minute after 10:14:00 with 10 s between two ticks.



## References

- Barker, B., Clark, M., Davis, P., Fisk, M., Hedlin, M., Israelsson, H., Khalturin, V., Kim, W.-Y., McLaughlin, K., Meade, C., Murphy, J., North, R., Orcutt, J., Powell, C., Richards, P., Stead, R., Stevens, J., Vernon, F., and Wallace, T, (1999). Monitoring nuclear tests. *Science*, **281**, 1967-1968.
- Gutenberg, B. (1946). Interpretation of records obtained from the New Mexico atomic bomb test, July 16, 1945. *Bull. Seism. Soc. Am.*, **36**, 327-330.
- ISC (2001). International Seismological Centre, On-line Bulletin, <http://www.isc.ac.uk/Bull>, Internatl. Seism. Cent., Thatcham, United Kingdom.
- Kværna, K., Ringdal, F., Schweitzer, J., and Taylor, L. (2002). Optimized seismic threshold monitoring — part 2: teleseismic processing. *Pure Appl. Geophys.*, **159**, 989-1004.
- Schlittenhardt, J. (1996). Array analysis of core-phase caustic signals from underground nuclear explosions: Discrimination of closely spaced explosions. *Bull. Seism. Soc. Am.*, **86**, 1A, 159-171.
- Sultanov, D. D., Murphy, J. R., and Rubinstein, Kh. D. (1999). A seismic source summary for Soviet Peaceful Nuclear Explosions. *Bull. Seism. Soc. Am.*, **89**, 3, 640-647.
- UK AWE (1993). *U.K. Atomic Weapons Establishment AWE Report No. O 11/93* (France).

|                |                                                                                                                                                                                                                                                                                                                                                         |
|----------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Topic</b>   | <b>Estimating the epicenters of local and regional seismic sources, using the circle and chord method</b><br>(Tutorial with exercise by hand and movies)                                                                                                                                                                                                |
| <b>Author</b>  | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a><br><b>Siegfried Wendt</b> , Geophysical Observatory Collm, University of Leipzig, D-04779 Wermsdorf, Germany, E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a> |
| <b>Version</b> | October 2011; DOI: 10.2312/GFZ.NMSOP-2_EX_11.1                                                                                                                                                                                                                                                                                                          |

## 1 Aim

This is a tutorial with related exercise. It aims at making you familiar with the basic “circle and chord” method used for determining the epicenter of near seismic sources when a “flat Earth” approach is appropriate. The method is applied both to sources inside and outside of the recording networks. With respect to phase names and essential phase features one should also consult section 2.6.1 in Chapter 2 and record examples in DS 11.2.

## 2 Data

- Available are two sections of vertical component short-period records of stations of the former Potsdam seismic network from a local earthquake inside the network (see Figure 2) and a strong rock-burst in a mine located outside of the network (see Figure 3).
- Travel-time curves of the main crustal phases Pn, Pg, Sn, Sg and Lg from a near surface source up to an epicentral distance of 400 km (see Figure 4). These curves are reasonably good average curves for Central Europe. For any stations in this exercise at distances beyond 400 km, you may linearly extrapolate the curve without much error.
- Map with the positions of the recording stations and a distance scale (see Figure 5).

## 3 Procedure

- Identify the seismic phases in short-period records of near seismic sources.
- By means of a local travel-time curve for near-surface events determine the source distance **d** from the best fit with the identified seismic phases.
- If no local travel-time curves are available, a first rough estimate of the hypocenter distance **d** or of the epicentral distance **D** (both in km) may be found using the following “rules-of-thumb”:

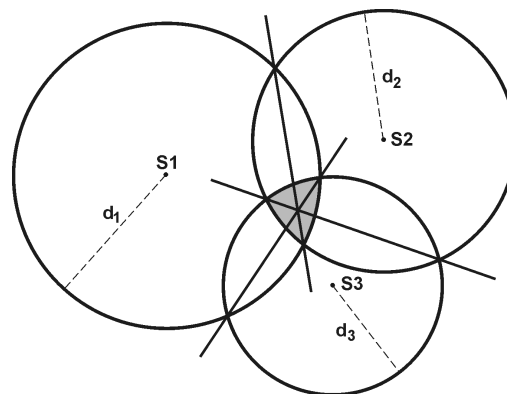
$$\mathbf{d} \approx \mathbf{t}(\mathbf{Sg} - \mathbf{Pg}) \times 8 \quad \text{or} \quad (1)$$

$$\mathbf{D} \approx \mathbf{t}(\mathbf{Sn} - \mathbf{Pn}) \times 10 \quad (2)$$

with  $t$  as the travel-time difference in seconds between the respective seismic phases. These rules are approximations for a single layer crust with an average Pg-wave velocity of 5.9 km/s, a sub-Moho velocity of about 8 km/s and a velocity ratio  $v_s/v_p = \sqrt{3}$ . If in your area of study the respective average P- and S-wave crustal velocities  $v_p$  and  $v_s$  deviate significantly from these assumptions you may calculate  $d$  more accurately from the relationship:

$$d = t(S_g - P_g) (v_p v_s)/(v_p - v_s). \quad (3)$$

- Draw circles with a compass around each station  $S_i$ , which is marked on a distance-true map projection, with the radius  $d_i$  determined from the records of each station.
- The circles will usually cross at two points, not one point (the thought epicenter) thus forming an area of overlap (see Figure 1, shaded area) within which the epicenter most probably lies.
- Usually, it is assumed, that the best estimate of the epicenter position is the “center of gravity” of this shaded area of overlap. The best estimate of the epicenter is found by drawing so-called “chords”, i.e., straight lines connecting the two crossing points of each pair of circles. The crossing point (or smaller area of overlap) of the chords should be the best estimate of the epicenter (see Figure 1).



$$d \approx t(S_g - P_g) \times 8$$

$$\text{or } d \approx t(S_n - P_n) \times 10$$

**Figure 1** Principle of epicenter estimation by using the “circle and chord” method.  $S$  – station sites,  $d$  – hypocentral distance of the event determined for each station according to travel-time curves (as given in the Figure 4) or the “rules-of-thumb” (as Equations (1) and (2)).

#### Notes:

- 1) In the absence of independent information on the source depth and depth-dependent travel-times the distance  $d$  determined as outlined above is not the epicenter but the *hypocenter distance*. Therefore, for sources at depth the circles will necessarily overshoot, the more so the deeper the focus. Iterative reduction of the overshoot permits to estimate the source depth (see section 7 with linked movie).
- 2) However, even for a surface source, an ideal crossing of circles at a single point would require that all phases have been properly identified, their onset times been picked without error and that the travel-time curves/model for the given area (including the

effects of lateral variations) are exactly known and taken into account. This, however, is rarely the case. Therefore, do not expect your circles to cross all at one point.

- 3) Yet, despite note 2, the circles should at least come close to each other in some area, overlapping or not, within about 10 to 20 km, if the epicenter is expected to lie within the network and the hypocenter within the crust. If not, one should check again the phase interpretation and resulting distance estimates and also compare for all stations the consistency of related estimates of origin time (see tasks below). Any obvious outliers should be re-evaluated.
- 4) For seismic sources outside the network the circle crossing will be worse, the error in epicenter estimation larger, particularly in the direction perpendicular to the azimuth of the connection line (great circle) between the network center and the source. However, the distance control, based on travel-time differences S-P, is better than the azimuth control. Azimuth estimates are more reliable if the source is surrounded by stations on at least three sides, i.e., with a maximum azimuthal gap less than  $180^\circ$ .
- 5) With only two stations one gets two possible solutions for the three unknowns (epicenter coordinates  $\lambda$  and  $\phi$  and origin time OT) unless the source direction can be independently determined from polarity readings in three-component records of each station (see EX 11.2). If more than 3 stations are available, the estimates of both epicenter and hypocenter will improve.

## 4 Tasks, data and approaches

### 4.1 Phase identification and travel-time fit (consult also section 5 for guidance)

- Identify the main local phases Pn, Pg, Sn and/or Sg/Lg in the records of the Potsdam seismic network (Figures 2 and 3) by using the travel-time curves given in Figure 4.
- Use plots of the t-D travel-time curves on transparent sheet and of the record section on paper with identical time resolution (e.g., 1 mm/s, as for the originally analog records). Overlay the t-D curves on the records so that the D(km) axis is strictly perpendicular and the t(s) axis parallel to the record trace zero lines!
- Move the t(s) axis over a given station record trace until you find a best fit for the first arrival and onsets of several later wave groups characterized by significant change/increase in amplitude and/or period. Mark with pencil the best fitting onset times with dots on the record together with their (assumed) phase names.

#### Notes:

- 1) When searching for the best fit remember that the beginning of the later wave group with the largest amplitudes in the record is usually the onset of Sg, whereas in the early parts of the record it is Pg that is usually the largest wave. For distances  $< 400$  km Pn is usually much smaller than Pg, although for deeper crustal earthquakes with appropriate rupture orientation and for distances  $> 400$  km Pn may become strong or even the dominating P-wave amplitude (see Figs. 11.44 to 11.46 in Chapter 11 and several network record sections in DS 11.1).
- 2) From the onset-time differences Sg-Pg and/or Sn-Pn you may roughly estimate the distance of the event by using the “rules-of-thumb” (Equations (1) to (3) above). If your rough estimate from S-P is  $d < 150$  km then the first arrival should never be interpreted as Pn but rather as Pg (unless it is a deeper crustal event or the crust is less than 30 km thick). If  $d > 150$  km, try to get the best fit to the onsets by assuming that

the first arrival is Pn, however remember that its amplitude is usually smaller than that of the following stronger Pg for  $d < 400$  km.

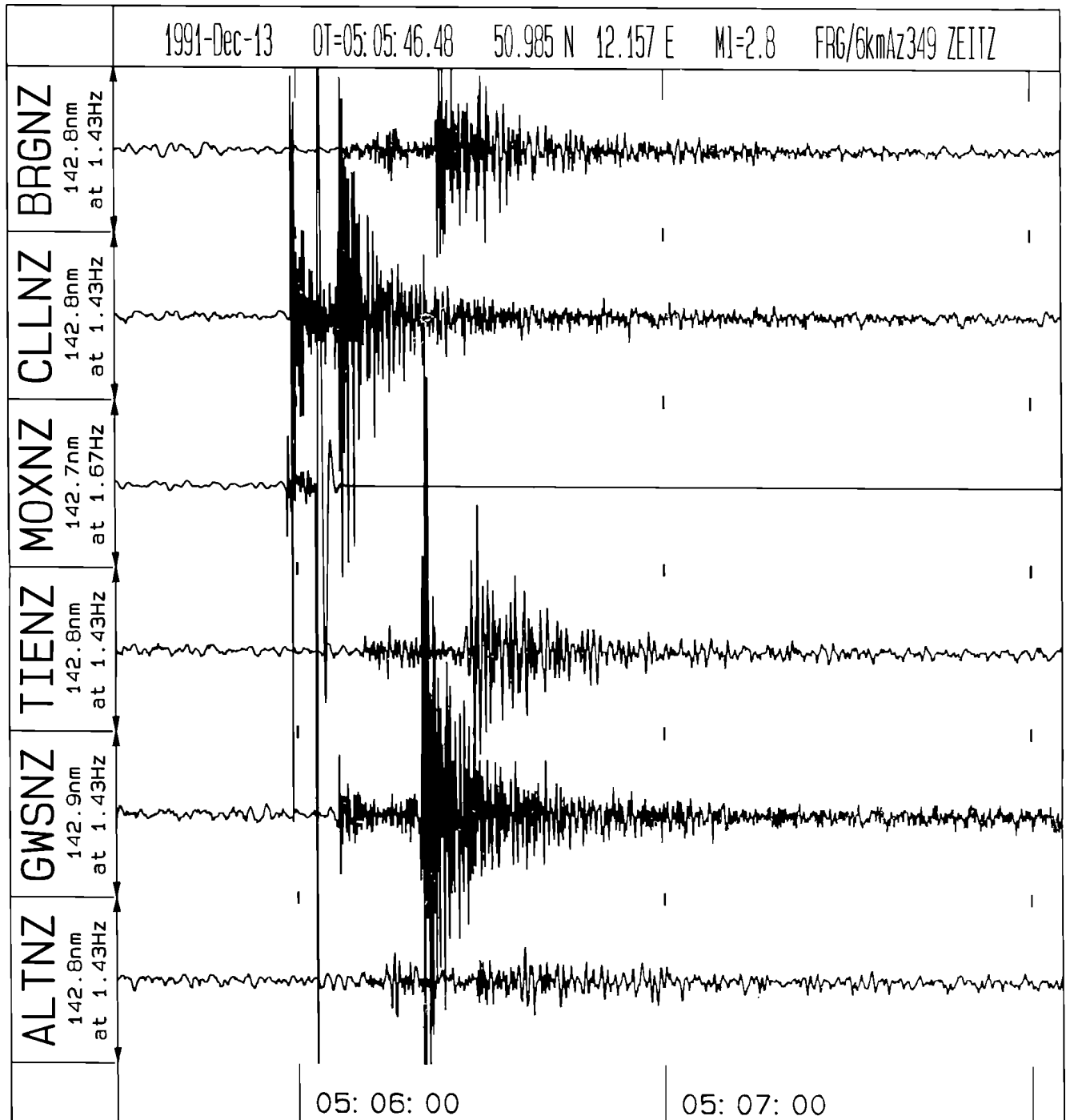
- 3) The above said is true for near-surface events in a single-layer crust with average P-wave velocity of 5.9 km/s and sub-Moho velocity of 8 km/s. The **cross-over distance**  $x_{co}$  beyond which Pn becomes the first arrival is then approximately  $x_{co} \approx 5 z_m$  with  $z_m$  as the Moho depth. In the case of different average crustal and sub-Moho P-wave velocities,  $\bar{v}_c$  and  $v_m$ , you may use the relation  $x_{co} = 2 z_m \{ (v_m + \bar{v}_c) / (v_m - \bar{v}_c) \}^{1/2}$  to calculate the cross-over distance of Pn. However, be aware that for deeper crustal events Pn may over-take already at smaller distances!

#### 4.2 Estimation of distance and origin time (consult also section 5 for guidance)

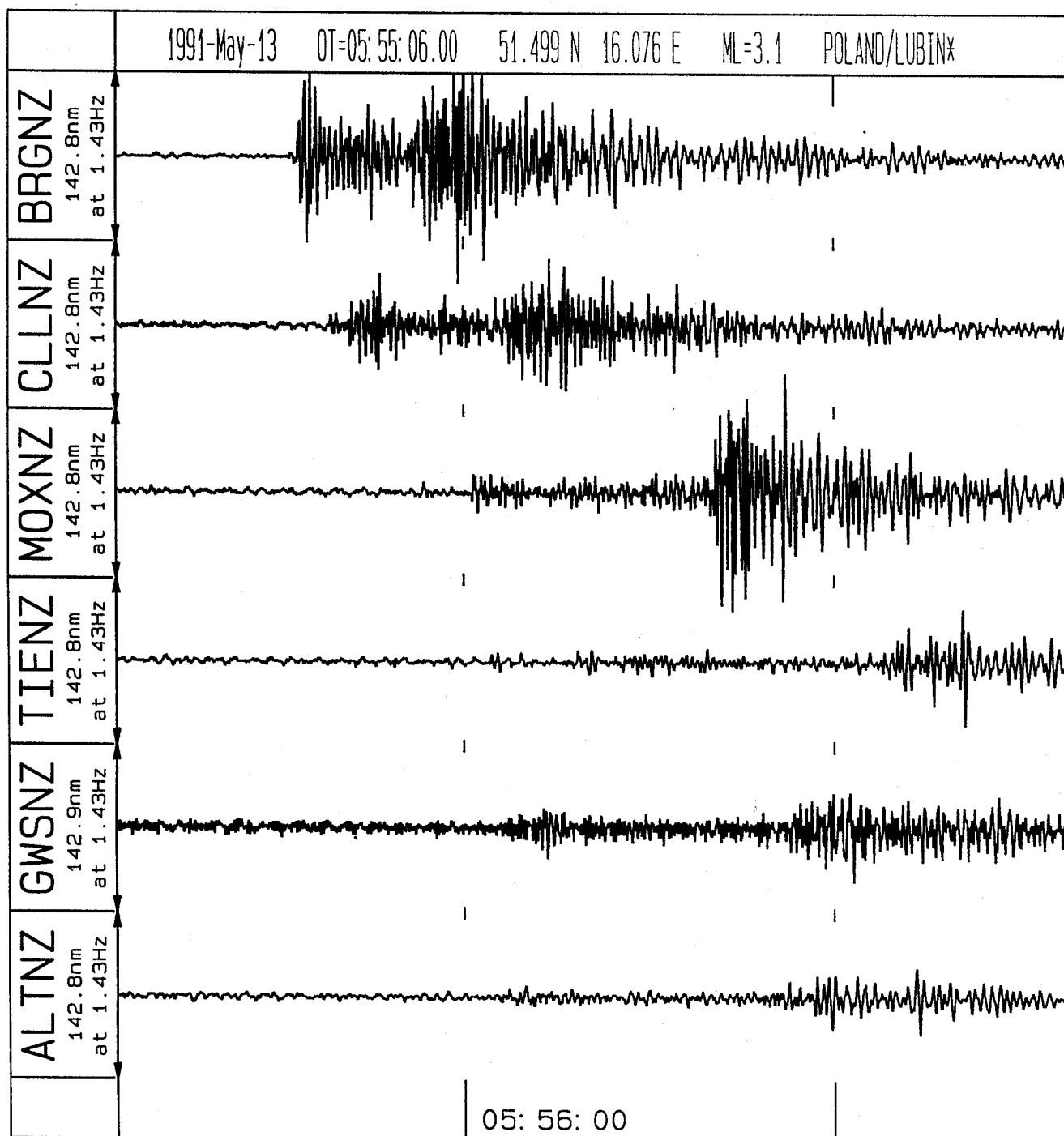
- Write down on the record plots for each station, next to the identified main wave onsets, its assumed phase name as well as the hypocentral distance corresponding to your S-P time reading, respectively the best travel-time fit. Mark on each record the estimated origin time, which is the time of the D-axis position on the record for your best phase travel-time fit.
- Check, whether your marks for the estimated origin times on the different records are roughly the same (in vertical line) for all stations. This is a good check of the accuracy and reproducibility of your phase identifications and estimated distances. For any “outliers” check the phase identification and distance estimate again until you get agreement between the origin times within about  $\pm 3$ s.
- Compare your best estimate of origin time OT (average of all origin times estimated from the records of each station) with the OT computer solution given in the head lines of Figures 2 and 3.
- If your average OT deviates by more than about 3 s from the computer solution reconsider your interpretation.

#### 4.3 Epicenter location

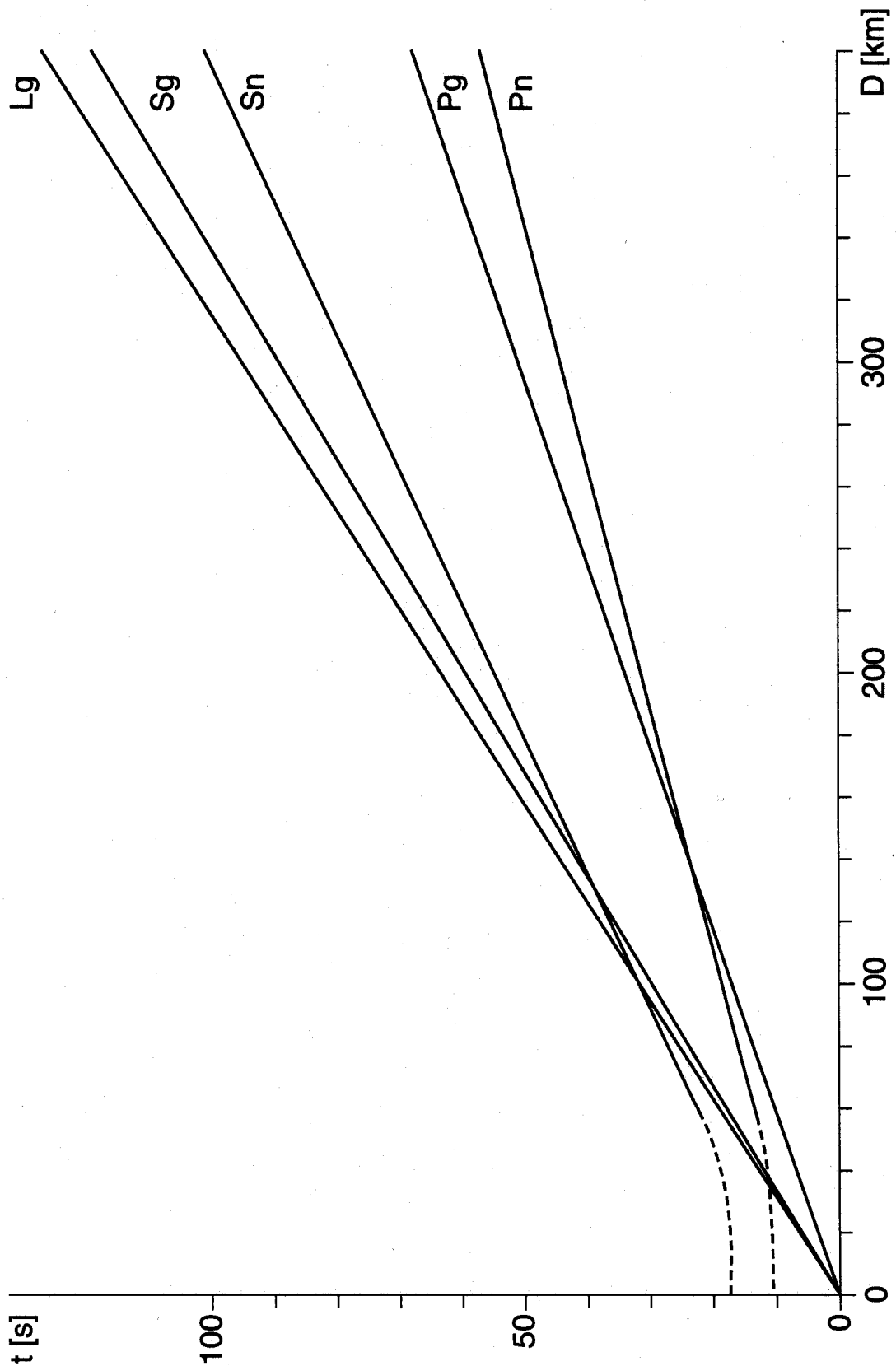
- Take a compass and draw circles around each station position (see Figure 5) with the radius  $d_i$  in km as determined for the distance of the source from the station  $S_i$ . Use the distance scale given on the station map.
- Connect the crossing points of each pair of circles by chords. Estimate the coordinates  $\lambda$  and  $\phi$  (in decimal units of degree) from the chord crossings.
- Compare your coordinates with the ones given in the headlines of Figures 2 and 3. If your solutions deviate by more than  $0.2^\circ$  for the earthquake within the network and by more than  $0.4^\circ$  for the mining rock-burst outside the network, reconsider your phase interpretation, distance estimates and circle-drawings.
- Compare your solutions for phase interpretations, station distances, circle plots and estimated epicenter positions with the solutions derived by the authors when applying the same procedure (see section 6).



**Figure 2** Recordings of a near earthquake situated within the seismic network of stations shown in Figure 5. The record time scale in hh:mm:ss is given below. Note that the second strong onset in the record of station MOX has a shape and frequency that strongly differs from all other records. It is not a natural wave onset but a malfunction of the seismograph, which responds to the strong Sg onset with its own impulse eigen response and then stops recording.

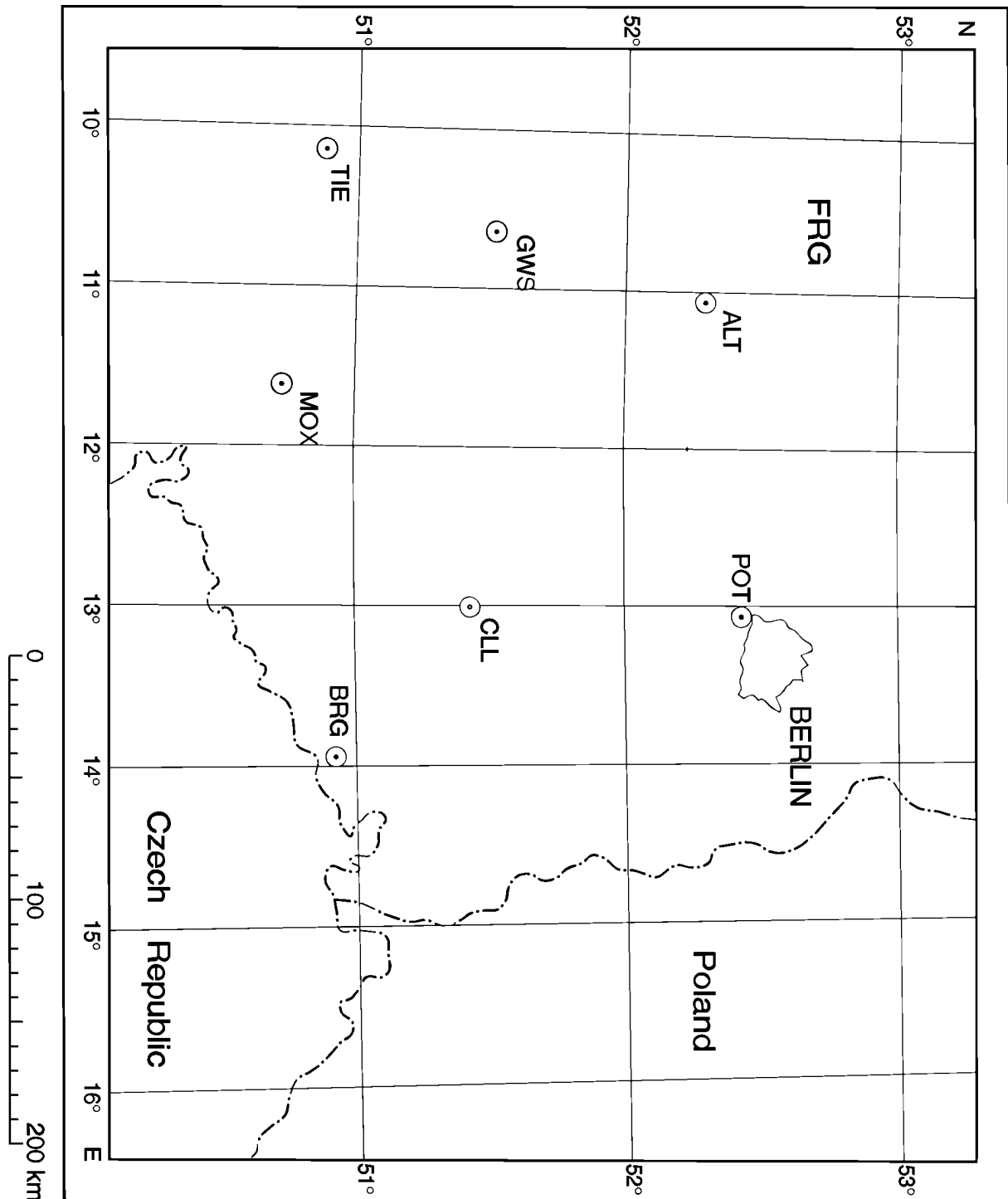


**Figure 3** Recordings at regional distances from a strong mining rock-burst situated outside the seismic network of stations shown in Figure 5. Time and minute intervall are given on the abszissa.



**Figure 4** Travel-time curves for the main phases observed in records of near-surface local and near regional seismic sources. They are good average curves for Central Europe with a crustal thickness of about 30 km.



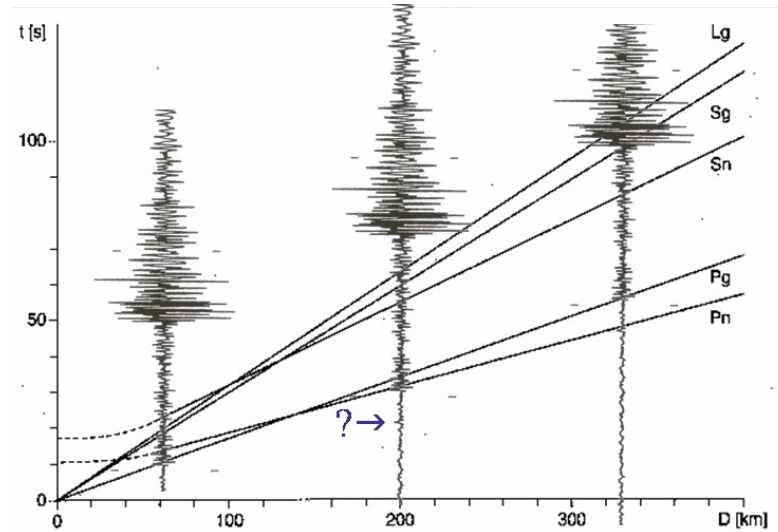


**Figure 5** Map of parts of Central Europe with codes and positions (circles) of the seismic stations that recorded the seismograms shown in Figures 2 and 3 (on the map projection all distances are true).

## 5 Some additional guidance

### 5.1 Phase identification and distance estimates by matching a seismic record with travel-time curves or using the S-P time difference

Figure 6 illustrates the identification of seismic phases in a record of a shallow local event by stepwise improved matching of recognizable (although sometimes small) wave onsets with travel-time curves and the estimation of the epicentral distance  $D$  (when source depth is assumed to be zero) from the best-fitting match.



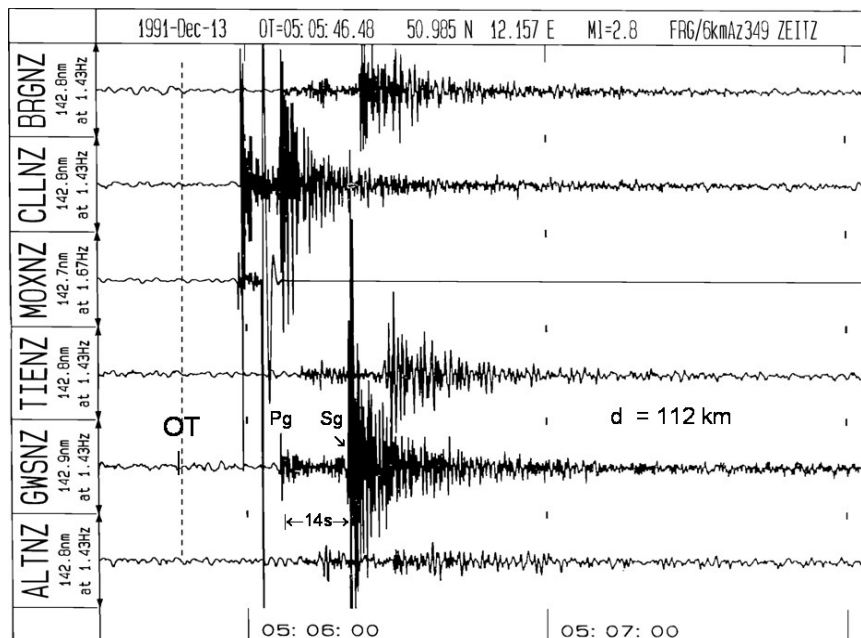
**Figure 6** Stepwise movement of a seismic record over a local travel-time curve until a best fit for three recognizable onsets and the Lg wave group is achieved at a distance  $D \approx 327$  km. For discussion, see text.

At the first two positions we related only the first clear record onset in the early part of the seismogram with the first P-wave travel-time curve. This would be the Pg curve at 60 km distance, respectively the Pn curve at 200 km distance. But then there is no match of the onset of the largest wave group in the later part of the record with any of the well established travel-time curves. Thus, for a better match one has to move the record further to larger distances. However, as outlined above, at distances about 5 times crustal thickness, Pn is expected to arrive as the first P phase, although (at distances  $< 400$  km) with usually much smaller amplitudes than that of Pg. Therefore, we may hypothesize that the small onset marked with ?  $\rightarrow$  in the second position maybe Pn. Although its SNR  $< 2$ , its waveform differs clearly from that of the preceding noise). Indeed, with this assumption one gets a good fit for Pn, Pg and Sg, as well as the following Lg-wave maximum, at position 3. The distance read at this position is  $D \approx 327$  km.

When considering that at distances  $< 400$  km Pg and Sg are usually the two most distinct first arrivals in the earlier and later part of the record one would already get - according to Equation (1) - for the record in Figure 5 with  $t(\text{Sg-Pg}) \approx 40$  s a distance of  $\approx 320$  km. This is already a good first estimate and requires to search for a possible Pn arrival prior to Pg. If, however, in records made under normal crustal conditions,  $t(\text{Sg-Pg}) < 18$  s then one can be

rather sure that these have all been recorded at distances  $d < \approx 140$  km. No other phases can then be expected to be first arrivals in the P- and S-wave groups in this near distance range.

Figures 7 and 8 illustrate for the two seismic events to be located by the same small network these two situations. Distance estimates are made only with the rule-of-thumb (1). On such a record section the first distance estimate should be made only from a record with undoubtedly clear onsets of Pg and Sg. Deducting from the onset-time of Pg its travel-time according to the t-D curve in Figure 6 for the determined distance one gets a first estimate of the origin-time OT. It should be marked on the record trace by a bar. The line along this bar and perpendicular to all other record traces then guides fairly well phase identification, OT and distance estimates on the other station records. Just place the D axis of the travel-time curves, corresponding to  $t = 0$  from a surface source, on/respectively close to this first OT reference line and move the t-D diagram up and down until you get a best fit also with the most distinct phase onsets in the other records. Mark also their respective OT estimates. They may differ from the first one within a few seconds due to reading errors, worse SNR and/or significant lateral velocity differences.

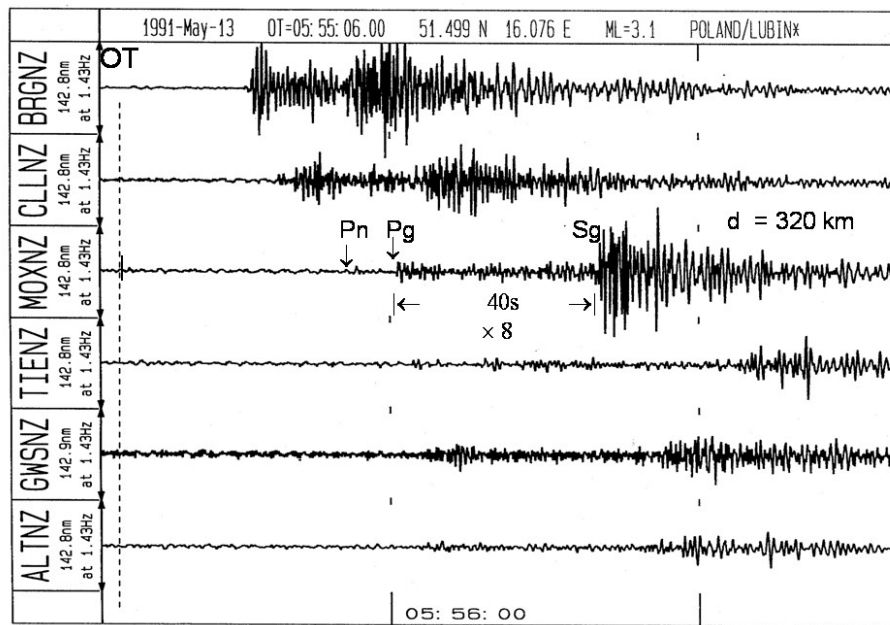


**Figure 7** Section of records from a local seismic event in the Pg-Sg distance range. Note the marking of the first origin time OT prior to the clearest Pg at distance of 112 km. This guides also phase identification and d estimates at the other stations when using the t-D curve in Figure 6. Note the distinct wave group between Pg and Sg in records of BRGNZ, TIENZ and ALTNZ. This is not generally present and not explained by the simple travel-time model of Figure 6. Travel-time modeling for an average crustal model makes it likely that this is an SmP wave, i.e., an S-to-P conversion at an overcritical reflection angle from the Moho.

In contrast to Figure 7, Figure 8 shows a record section from another event outside of the network. All stations were at distances larger than 140 km when Pn is already the first arrival. This is easily found out by a quick check of the time differences  $t(S-P)$  between the two most distinct phases, Pg and Sg, which is everywhere  $> 18$  s. The follow-up procedure of more detailed phase identification and d determination by moving the travel-time curve of Figure 6 over the records along the first OT line estimate is the same as described above. For more

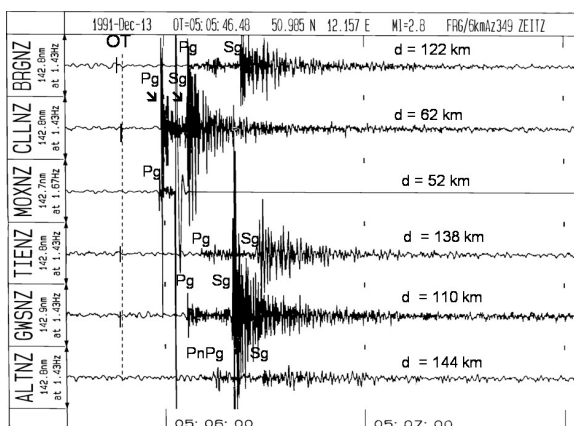
record examples in the local and regional distance range see section 11.5.1 in Chapter 11 and DS 11.1.

Digital records now permit to stretch the time and amplitude scales at will. This may significantly ease phase identification and allows much more accurate picking of onset times (see IS 11.4).

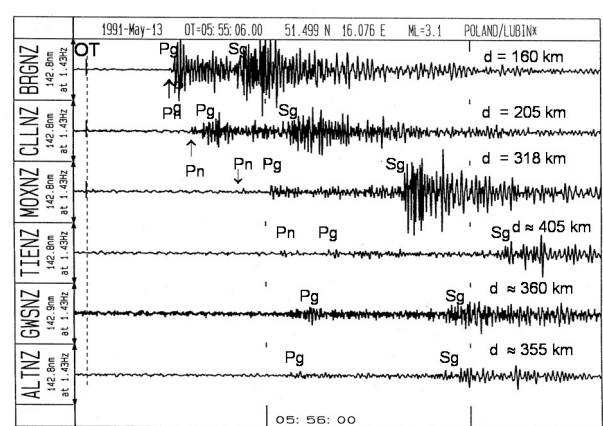


**Figure 8** Section of records from a seismic event in the Pn distance range. Note the rather vague onsets of Pg and Sg in the lower three record traces. They will allow only rough estimates of  $d$ .

## Solutions

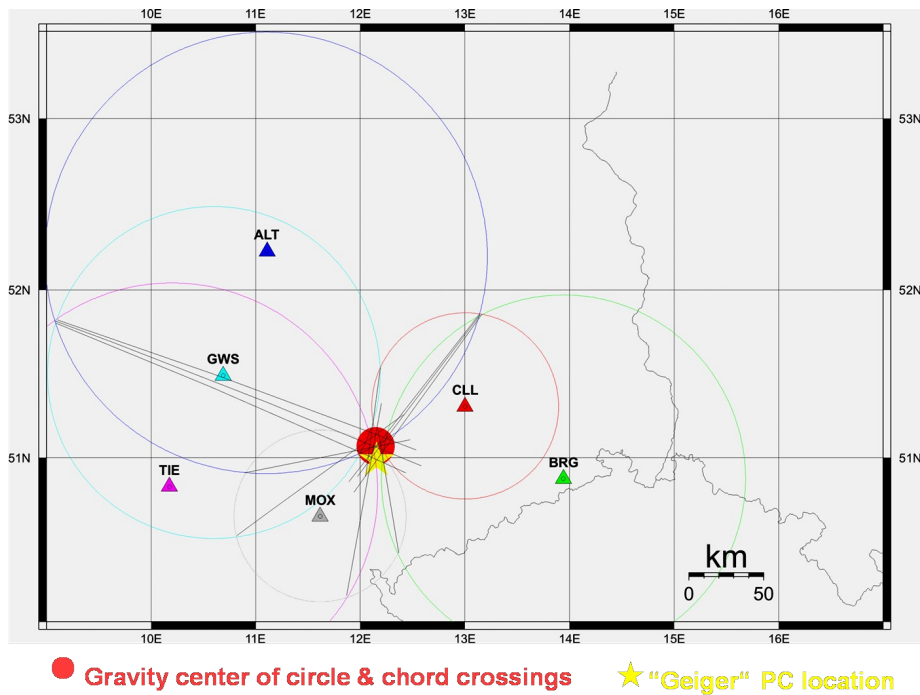


**Figure 9** OTs, identified phases and calculated distances for a seismic event in the local Pg-Sg distance range.

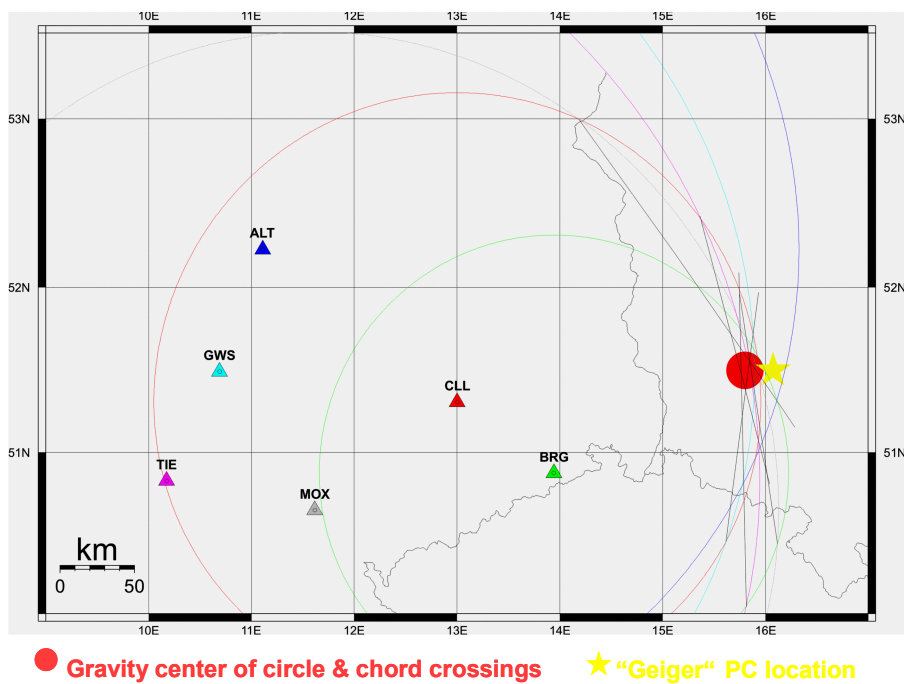


**Figure 10** The same as in Figure 9 for an event in the Pn distance range. Note that in the record at 405 km Pn and Pg have already about the same amplitude, whereas at shorter distances Pn is much smaller or not even

recognizable above the noise.



**Figure 11 Location** by the circle and chord method for the event **within the network**, using only records in the local Pg-Sg distance range and the  $d$  values given in Figure 9, determined by the "rule of thumb" (1). Note the good agreement between this rough solution by hand and the PC assisted record analysis, phase picks and event location.



**Figure 12 Location** by the circle and chord method for the event **outside the network**, using records in the Pn distance range and the  $d$  values given in Figure 10. Note the much worse circle crossing. It results in larger location uncertainty (long axis of the error ellipse) in N-S

direction, which is perpendicular to the connection line between the network center and the source. This applies to both the manual and the PC solution. Again the two agree rather well.

## 6 Determination of source depth

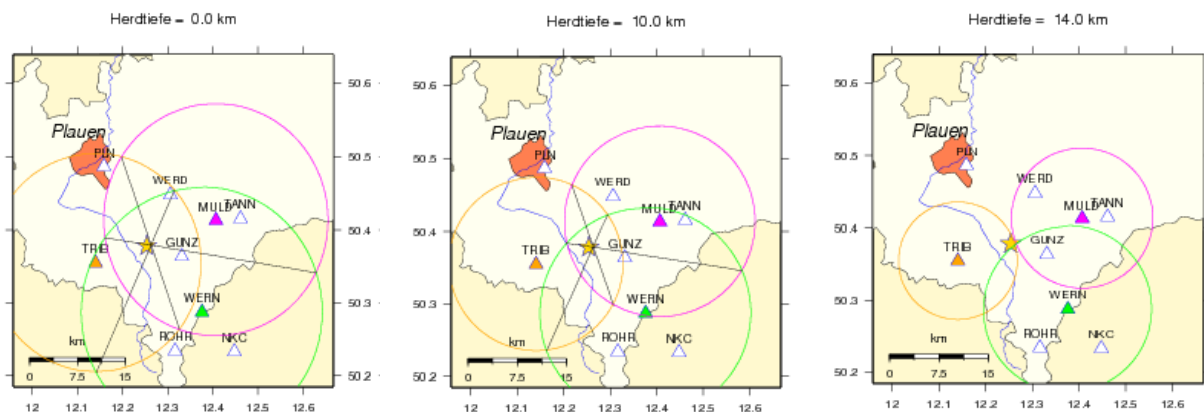
The above exercise was run on the assumption of a (near) surface source for which effects of source depth  $h$  are negligible. However, as mentioned earlier, the overshoot of the distance circles drawn around the stations is largely due to the negligence of  $h$  and the fact that for  $h > 0$  it holds  $d$ (= hypocentral distance)  $>$   $D$ (= epicentral distance).

However, the records in Figure 9 relate to a local earthquake of larger source depth, in contrast to the records in Figure 10, which result from a more shallow strong mining explosion in Poland. Therefore, the overlap of circles in Figure 11 is larger than in Figure 12, although quite some uncertainties are introduced in the latter case due to more difficult phase identification, often worse SNR, less accurate onset time picking as well as pronounced lateral velocity differences towards the boundary of the East European Platform, which are not accounted for by the used simple one-dimensional travel-time model.

If, however, large circle overshoot hints to  $h \gg 0$  km one should re-calculate the station radii  $d_i$  by stepwise increasing  $h$  until for  $d_{imin}$  the “true” (better: the most likely) source depth  $h_{true}$  can approximately be estimated via the relationship

$$h_{true} = (R_i^2 - d_{imin}^2)^{1/2}.$$

The circles drawn around the stations with  $d_{imin}$  are expected to have zero (or negligible) overshoot, provided that the wave propagation conditions are sufficiently homogeneous and well represented by the assumed velocity model and further that the station reading errors of phase onset times are negligible. This is illustrated by the 3 plots in Figure 13 for another local earthquake.



**Figure 13** Sequence of circle & chord plots for an earthquake in the Vogtland swarm earthquake region of South-Eastern Germany. Assumed are different source depths for fixed hypocentral distances  $d_i$  of the  $i$  stations as calculated from the S-P travel-time differences. For  $h = 14$  km the circles cross almost ideally in one point at the epicenter.

Allowing for finite source depth also means that the travel-time curves for the direct P and S waves would no longer be straight lines with zero travel time at  $D = 0$  km. Rather, P and S would have bended travel-time curves with different finite travel times also at the epicenter. Accordingly, for constant S-P times and thus  $d$  the epicentral distance  $D$  decreases with growing depth until one gets at the epicenter position  $D = 0$  that  $d = h$ . This also explains why reliable depth estimates require records from stations in the near source area at  $D < h$ .

Both effects are well illustrated by the movies [\*\*3D-wave-prop\\_travel-times\\_qt\*\*](#) and [\*\*circles\\_depth\\_qt\*\*](#). You can activate these movies by right mouse click on the respective file name in bold blue. Note in the final travel-time-record plot shown in the first movie that actual phase onset times at some of the stations may differ by several tenths of a second from the theoretically calculated ones for an average 1-D crustal velocity model of the considered area.

The movies can also be downloaded via the summary listing *Download Programs & Files* (see Overview on the NMSOP-2 cover page and follow related instructions).

| Topic   | Earthquake location at teleseismic distances from<br>3-component records (Tutorial with exercise by hand)                                                                                                                                                                                                                                               |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a><br><b>Siegfried Wendt</b> , Geophysical Observatory Collm, University of Leipzig, D-04779 Wermsdorf, Germany, E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a> |
| Version | October 2012; DOI: 10.2312/GFZ.NMSOP-2_EX_11.2                                                                                                                                                                                                                                                                                                          |

## 1 Aim

This exercise and tutorial aims at making you familiar with the basic concept of locating seismic events by means of teleseismic records from single 3-component stations. Often the results are comparably good or even better than for uncalibrated single seismic arrays.

The exercise uses teleseismic events only although the procedure outlined below is the same for local seismic events. In the latter case local travel-time curves as in Exercise EX 11.1 have to be used for phase identification and distance determination. Note however that azimuth determinations for local events are less reliable when short-period records are used. They are much more influenced by local heterogeneities in the crust than teleseismic long-period or broadband records. Accordingly, particle motion might deviate significantly from linear polarization (see Chapter 2, Fig. 2.6) and the azimuth of wave approach for local events sometimes deviates more than  $20^\circ$  ( $+ 180^\circ$ ) from the backazimuth AZI towards the source (see Chapter 11, Fig. 11.23).

## 2 Data

The following data are used in the exercise:

- two 3-component earthquake records: a) Kirnos BB-displacement seismograms (Figure 2) and b) long-period (WWSSN-LP) seismograms (Figure 3);
- simplified differential body-wave travel-time curve (with respect to the P-wave first arrival) for the distance range  $0^\circ < D \leq 100^\circ$  (Figure 4) according to the classical Jeffreys-Bullen (1970) travel-time curves;
- IASP91 table (Kennett, 1991) of travel-time differences pP-P and sP-P, respectively, as a function of epicentral distance D (in degree) and source depth h (in km) (see Table 1);
- global map of epicenter distribution with isolines of epicentral distance D (in  $^\circ$ ) and principal directions of backazimuth AZI from station CLL (Germany).

## 3 Procedure

### 3.1 Assessment of the earthquakes distance and depth range, identification of essential phases, estimation of epicentral distance D and depth h



- Identification of the very first as well as of later secondary arrivals from teleseismic events in broadband or long-period filtered records. At least P and S have to be identified. P is the first arrival (up to about 100°). For teleseismic events ( $D > 20^\circ$ ) P has its largest amplitude in the vertical component. S is the first arriving shear wave, up to about 83°, and has its largest amplitudes in horizontal components. For larger distances SKS becomes the first arriving shear wave (see t-D curves in Figure 4 and record examples in Figures 10c, 12d, 14e and 15b of DS 11.2 ). Misinterpreting it as S might result in significant underestimation of the epicentral distance.
- Determination of epicentral distance D by using the travel-time difference  $t(S-P)$  according to travel-time tables (e.g., IASPEI91; Kennett 1991) or by fitting best a set of differential travel-time curves as in Figure 4 with the identified phases in the record. Note that the records and the t-D-curves must have the same time scale. In the distance range  $20^\circ < D < 85^\circ$  the following *rule-of-thumb* allows to determine D with an error  $< 3^\circ$ :

$$D [^\circ] = [t(S-P)_{\text{min}} - 2] \times 10. \quad (1)$$

- Note, that the travel-time difference  $t(S-P)$  but also the time difference between P or PKP (beyond 105°) with other secondary phases is influenced by the source depth h. Unrecognized significant source depth might result in underestimating D by several degrees. Accordingly, it is important to assess first whether an event was deep or shallow (i.e., probably within the crust).
- For a first rough discrimination between deep and shallow earthquakes one should compare the amplitudes of body waves with that of (dispersed!) surface waves. If the latter are well developed and significantly larger in amplitude than the earlier body waves, then the event can be considered a crustal earthquake. If, however, the long-period surface waves have much smaller amplitudes than the earlier body waves or are not recognizable at all, then the earthquake is surely deeper than 100 km.
- The identification of depth phases (see Chapter 2, Fig. 2.43) is difficult for shallow events, because then they follow within a few seconds to the primary P or S onset. One may then use travel-time curves or tables for surface focus ( $h = 0$  km) or “normal depth” events ( $h = 33$  km) for estimating the distance. In any event, if  $D \gg h$ , the error in distance estimate is usually negligible.. They are crucial for improving hypocenter locations.
- However, in the case of relatively weak or absent surface waves one should look for depth phases! They are crucial for improving hypocenter locations (see discussion in section 11.5.4 of Chapter 11 and Figure 7 in IS 11.1). Examples for depth phases are given in DS 11.2, Figures 7b, 9a, 16b and in DS 11.3, Figures 3b-d and 5a. If depth phases such as pP and/or sP have been identified, h can be calculated rather reliably when the epicentral distance is roughly known. One may use for it either the differential travel-times pP-P or sP-P (see Table 2), or, if such tables are not at hand, roughly estimate h by another rule-of-thumb:

$$h [\text{km}] \approx 0.5 t(\text{pP-P}) [\text{s}] \times n \quad (2)$$

with  $n = 7$ , or 8, or 9 for  $h < 100$  km, 100–300 km, or...  $\times 9$   $h > 300$  km.

- Generally, when interpreting seismic records you should proceed as follows:

**Take interest!****Be curious!****Ask questions!**

to your seismic record

**1. Is the event NEAR ( $D < 20^\circ$ ) or TELESEISMIC ( $D > 20^\circ$ )?****Criteria:**

- |                          |                                    |                                     |
|--------------------------|------------------------------------|-------------------------------------|
| • <b>Frequencies</b>     | on SP records $f \geq 1$ Hz        | $f \leq 1$ Hz                       |
| • <b>Amplitudes</b>      | on LP records <b>not or weaker</b> | <b>large, also for later phases</b> |
| • <b>t(S-P)</b>          | $< 3.5$ min                        | $> 3.5$ min                         |
| • <b>Record duration</b> | $< 20$ min                         | $> 20$ min                          |
- (for magnitudes  $< 5$ ; may be longer for strong earthquakes)

**2. Is the event SHALLOW or DEEP ( $> 70$  km)?****Criteria:**

- |                        |                             |                                       |
|------------------------|-----------------------------|---------------------------------------|
| • <b>Surface waves</b> | on LP records <b>strong</b> | <b>weak or none</b>                   |
| • <b>Depth phases</b>  | usually <b>not clear</b>    | <b>well separated and often clear</b> |
| • <b>Waveforms</b>     | usually <b>more complex</b> | <b>more impulsive</b>                 |

**3. Is the event  $D < 100^\circ$  or  $D > 100^\circ$  ?****Criteria:**

- |                            |                                  |    |                                      |
|----------------------------|----------------------------------|----|--------------------------------------|
| • <b>Surface wave max.</b> | after P arrival $< 45 \pm 5$ min | or | $> 45 \pm 5$ min (Table 5 in DS 3.1) |
| • <b>Record duration</b>   | on LP records $< 1.5$ hours      | or | $> 1.5$ hours                        |
- (may be larger for very strong earthquakes)

**4. Is the first strong horizontal arrival S or SKS ?****Criteria:**

- |                               |                                     |                          |
|-------------------------------|-------------------------------------|--------------------------|
| • <b>Time difference to P</b> | $< 10 \pm 0.5$ min                  | $\approx 10 \pm 0.5$ min |
| • <b>Polarization</b>         | large horiz. A <b>in R and/or T</b> | <b>in R only</b>         |

**Warning !** If the first strong horizontal arrival follows P after  $\approx 10 \pm 0.5$  min it may be SKS.

**Check polarization** (see 11.2.4.3 in Chapter 11). R and T are the radial and transverse horizontal components, respectively. Misinterpreting SKS as S may yield D estimates up to  $20^\circ$  too short. Look also for later multiple S arrivals (SP, SS, SSS) with better D control.

**5. What are the first longitudinal and transverse onsets for  $D > 100^\circ$  ?**

Beyond  $100^\circ$  epicentral distance phase interpretation and distance determination from travel-time differences between secondary phases becomes more difficult. A long-period P may still arrive first, up to about almost  $150^\circ$ , yet with steadily decreasing amplitude and therefore recognizable only in records of strong earthquakes. But this P did not travel a direct mantle path, rather it has been diffracted around the core-mantle boundary. Therefore, it is termed Pdif (old nomenclature Pdiff; see Figs. 11.59 and 11.63 in Chapter 11). First onsets in SP records, however, are beyond  $100^\circ$  usually the later arriving PKiKP and PKPdf (Fig. 11.59), or, somewhat later PP. The latter is often the first strong longitudinal Z-component arrival in both SP, LP and BB records (see Figs. 11.60 and 11.63). The first strong arrivals on horizontal (R) components are PKS or SKS.

Misinterpretation of these first strong longitudinal and transverse wave arrivals as direct P and S, respectively, may result in epicentral distance estimates up to more than  $70^\circ$  too short! This, however, can be avoided by taking the criteria 3. above into account. Also note, that the travel-time difference between PKPdf and PKS or SKS is (almost) independent of distance. Therefore, these first arriving P and S waves do not allow distance to be estimated from their

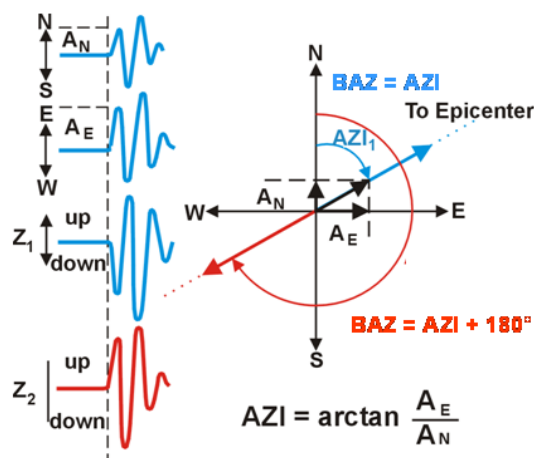
onset-time difference. Rather look for later arriving multiple-reflected S waves such as SS, SSS, etc. They are usually well developed in this distance range on horizontal LP records, have a large slowness difference with respect to earlier arriving core phases and thus allow D to be estimated with an error of usually  $< 2^\circ$ .

### 3.2 Estimation of backazimuth AZI

- Identify the proper direction of P-wave first motion in the three components Z, N, E. Make sure by exact time correlation that you really compare the same first half cycles in all three records! This is particularly important, if in one of the horizontal components the first onset is very weak or near to zero. Then one might be misled and associate the stronger amplitude of a later half cycle with the first motion in the other components and gets a wrong backazimuth.
- Determine the **direction of particle motion AZI** in different horizontal quadrants from the amplitudes of first motions in the horizontal component records according to the formula

$$\text{AZI} = \arctan (|A_E/A_N|) \quad (3)$$

- If seismograph components have been calibrated properly and avail of identical frequency responses (which is the case for the records used in the two exercises below) then one just calculates the ratio between measured trace amplitudes. However, as demonstrated in Figure 1, the direction AZI may either show **towards the epicenter**, in case the first motion in Z is down (-, dilatational; see blue record traces), or away from the epicenter in the direction of wave propagation if the first motion in Z is up (+, compressional; see red record trace). In the latter case, the backazimuth BAZ, measured clockwise from north towards the source epicenter as seen from the station) is  $\text{BAZ} = \text{AZI} + 180^\circ$ .
- But the conversion from AZI into BAZ depends on the quadrant of particle motion and thus the polarity patterns in the three components. Table 1 guides you how to determine the quadrant of horizontal particle motion and then to calculate BAZ from measured AZI.



**Figure 1** Synthetic example of P-wave first motions in 3-component records (left) from which the azimuth AZI (clockwise from north towards the direction of wave-propagation), respectively the backazimuth BAZ (clockwise from north towards the direction of the source as seen from the station) can be calculated.

**Table 1** How to identify the quadrant of AZI measurement and to calculate BAZ from AZI.

| Z | N-S | E-W | AZI                      | BAZ              |
|---|-----|-----|--------------------------|------------------|
| – | +   | +   | 1 <sup>st</sup> quadrant | BAZ = AZI        |
| – | –   | +   | 2 <sup>nd</sup> quadrant | BAZ = 180° - AZI |
| – | –   | –   | 3 <sup>rd</sup> quadrant | BAZ = 180° + AZI |
| – | +   | –   | 4 <sup>th</sup> quadrant | BAZ = 360° - AZI |
| + | +   | +   | 1 <sup>st</sup> quadrant | BAZ = 180° + AZI |
| + | –   | +   | 2 <sup>nd</sup> quadrant | BAZ = 360° - AZI |
| + | –   | –   | 3 <sup>rd</sup> quadrant | BAZ = AZI        |
| + | +   | –   | 4 <sup>th</sup> quadrant | BAZ = 180° - AZI |

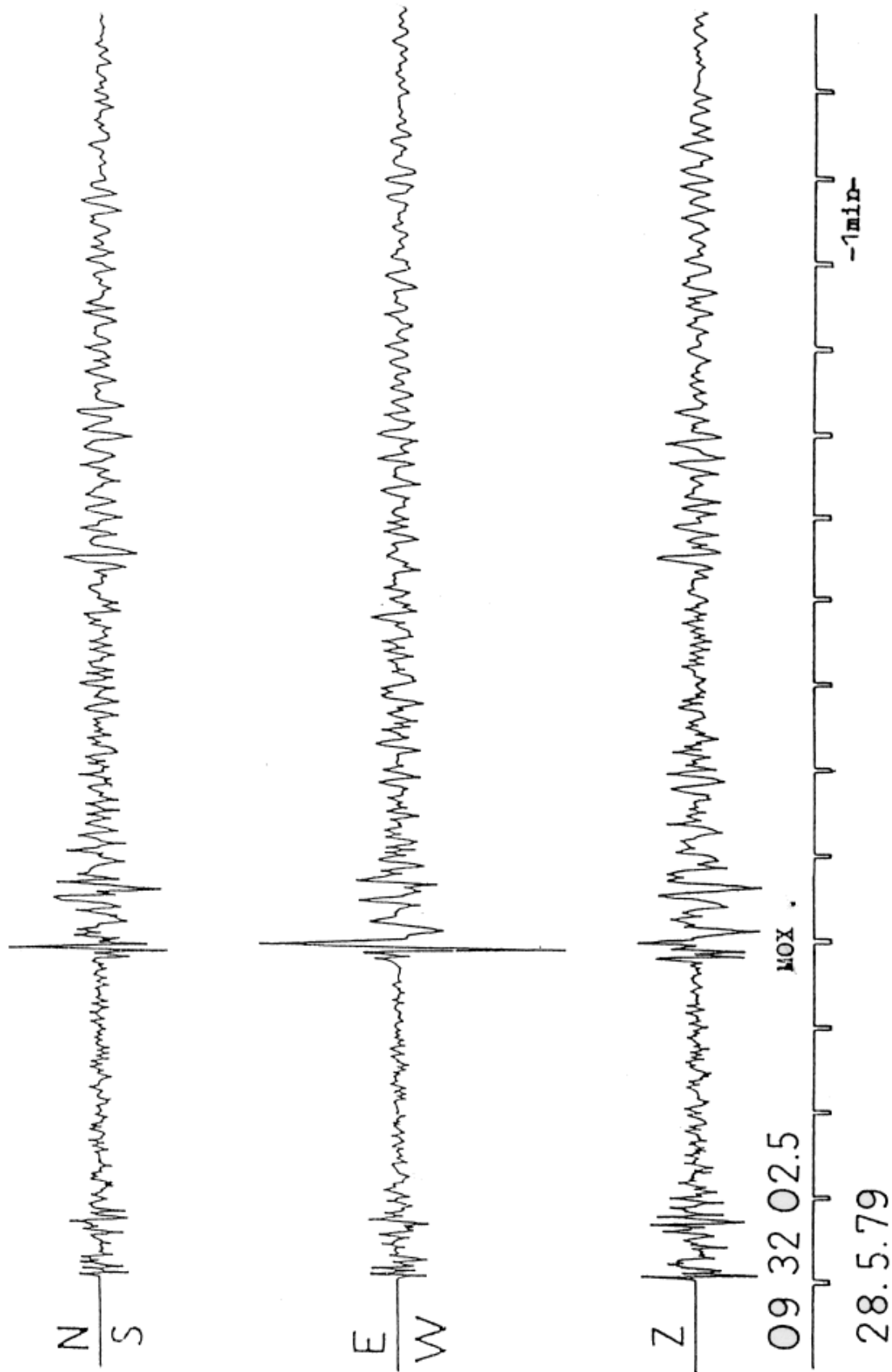
- If this 180° ambiguity has been resolved by taking into account the P-wave vertical component first motion polarity then one may also calculate BAZ from horizontal component records of either later cycles of P with larger amplitudes or even by using the amplitude ratio in E/N from other later phases that are polarized in the vertical propagation plane such as PP, SKS, SP etc.

### 3.3 Event location using the estimated epicentral distance D and backazimuth BAZ

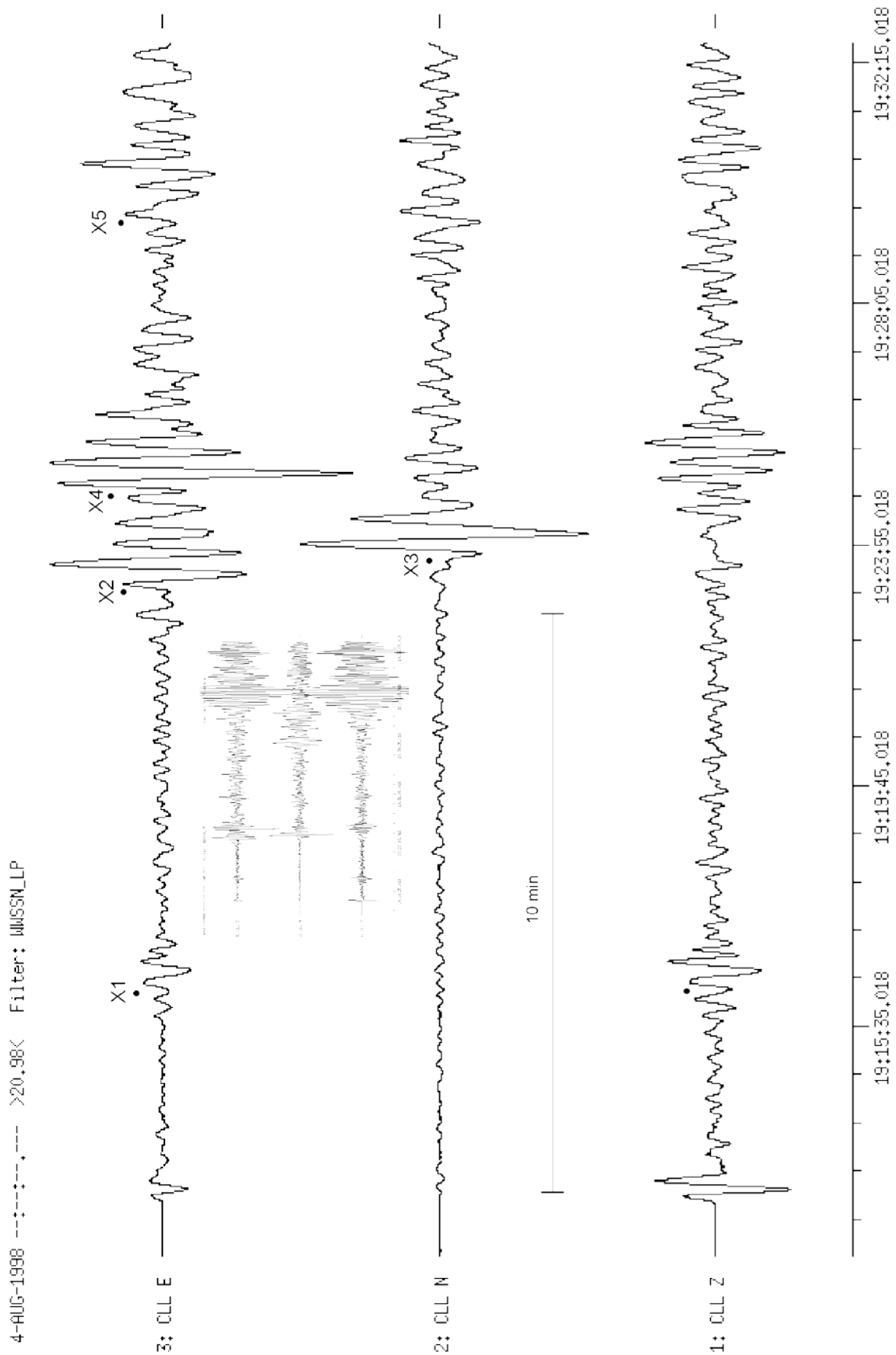
- You may use a sufficiently large globe (diameter about 0.5 to 1 m), mark there the position of your station and then use a bendable ruler with the same scale in degree as your globe and an azimuth dial to find your event location on the globe.
- Another possibility is that you get a regional (Figure 5) or global map projection (Figure 6) which shows isolines of equal azimuth and distance from your station. Such maps can nowadays easily be calculated and plotted for any station with known co-ordinates together with the seismicity pattern.

## 4 Data used for the exercise

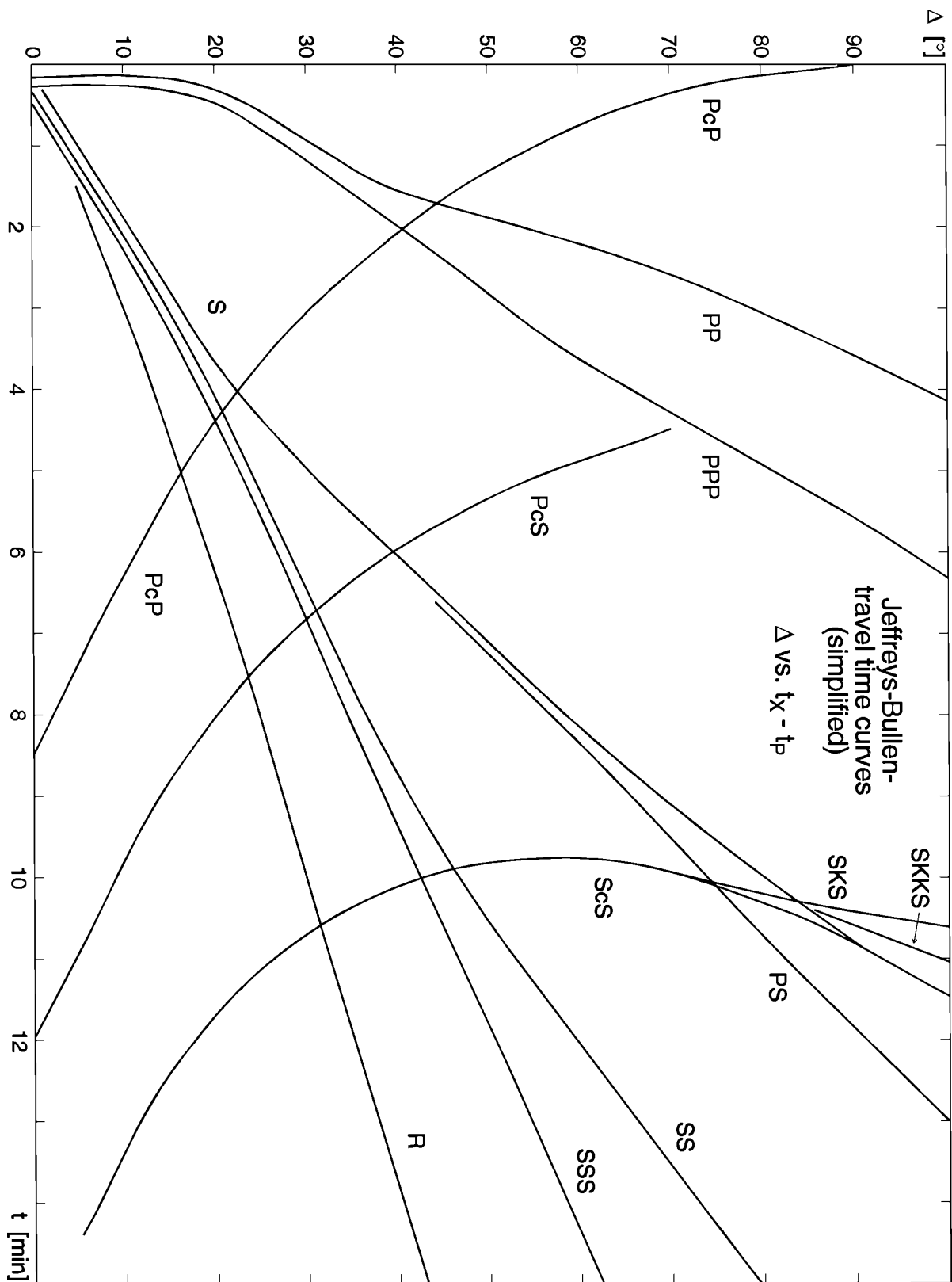
- 3-component broadband displacement records of an earthquake plotted in Figure 2;
- 3-component long-period WWSSN records of an earthquake plotted in Figure 3;
- Simplified differential travel-time curve according to the Jeffreys-Bullen model for the distance range up to 100°;
- Regional map for Europe and the Mediterranean with earthquake epicenters and isolines of D and BAZ with respect to station MOX, Germany;
- World map with epicenters and isolines of D and BAZ from station CLL, Germany;
- Table 2 of travel-time differences pP-P and sP-P as a function of distance D and depth h according to the IASP91 travel-time tables (Kennett, 1991).



**Figure 2** 3-component record of a Kirnos BB-displacement seismograph at station MOX, Germany. For response see Fig. 3.11 in Chapter 3. Original time-scale of the analog record was 15 mm/minute. All components were properly calibrated and had identical magnification.

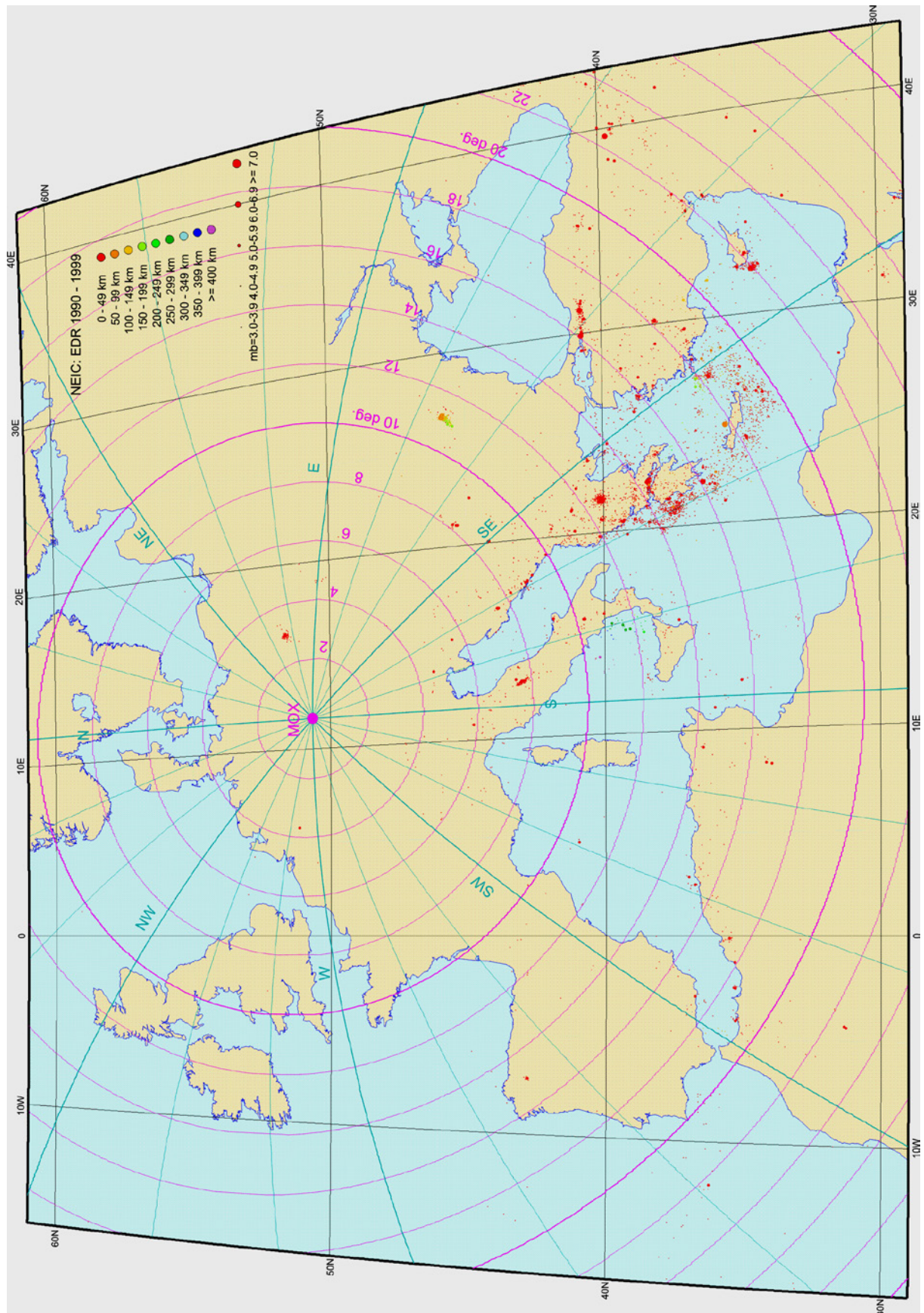


**Figure 3** A long-period 3-component body-wave record section (WWSSN-LP simulation) of station CLL, Germany. **Insert:** Complete event record at compressed scale.



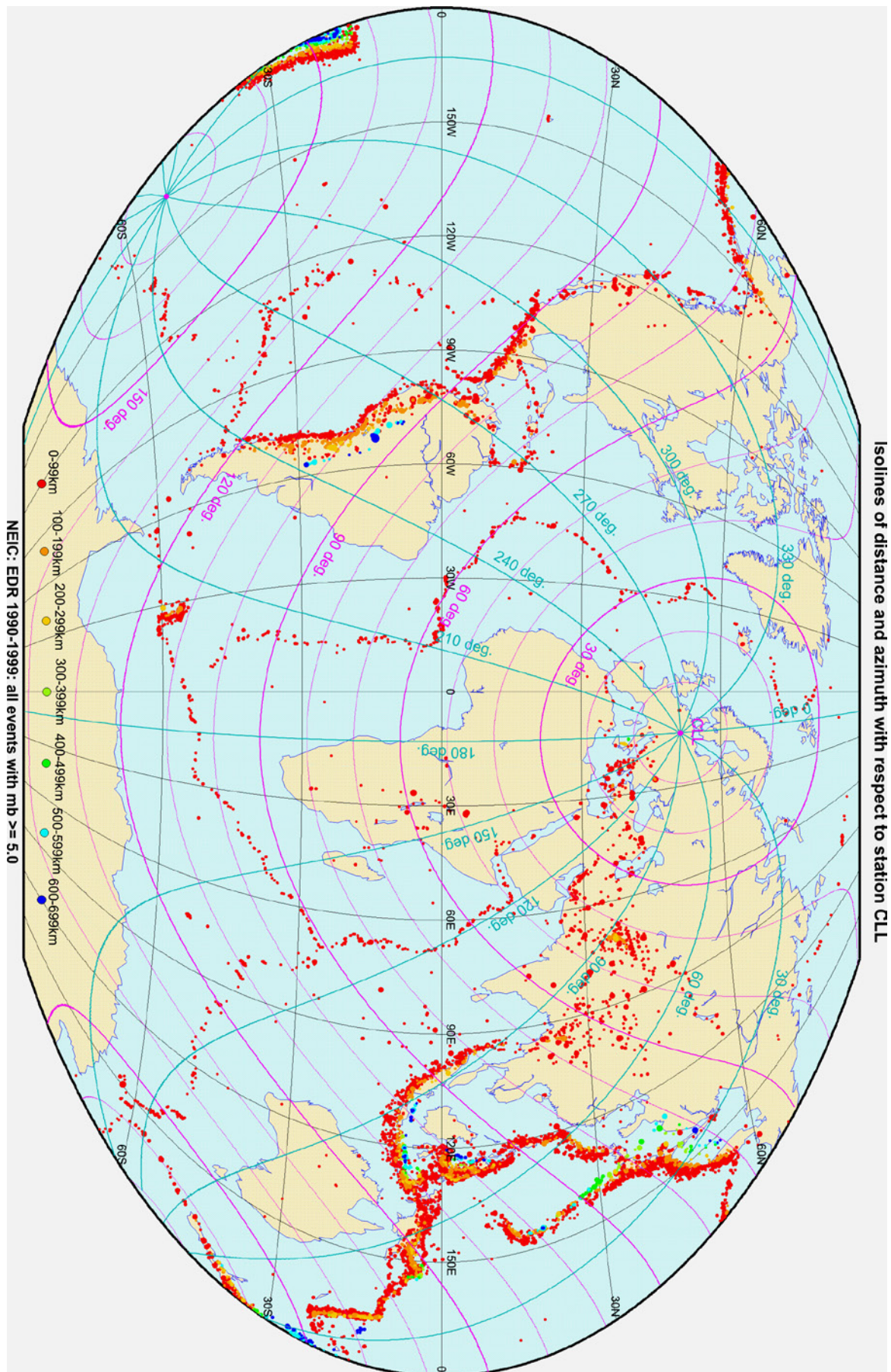
**Figure 4** Simplified Jeffreys-Bullen differential travel-time curve in the distance range  $0^\circ < D \leq 100^\circ$ . Original time scale was 15 mm = 1 min.





**Figure 5** Regional map for Europe and the Mediterranean with earthquake epicenters and isolines of D and AZI with respect to station MOX, Germany.





**Figure 6** World map with epicenters and isolines of D and AZI fort station CLL, Germany.

**Table 2** Travel-time differences pP-P and sP-P as a function of distance D and depth h according to the IASP91 travel-time tables (Kennett, 1991).

| pP-P<br>$\Delta$     | 15. | 35.  | 50.  | 100. | 150. | 200. | 250. | 300.  | 400.  | 500.  | 600.  | 700.  |
|----------------------|-----|------|------|------|------|------|------|-------|-------|-------|-------|-------|
|                      |     |      |      |      |      |      |      |       |       |       |       |       |
| sP-P<br>$\Delta$     | 15. | 35.  | 50.  | 100. | 150. | 200. | 250. | 300.  | 400.  | 500.  | 600.  | 700.  |
|                      |     |      |      |      |      |      |      |       |       |       |       |       |
| Depth of source [km] | 15. | 35.  | 50.  | 100. | 150. | 200. | 250. | 300.  | 400.  | 500.  | 600.  | 700.  |
|                      |     |      |      |      |      |      |      |       |       |       |       |       |
| 2                    | 5.9 | 12.7 | 15.4 | 22.8 | 27.0 | 35.7 | 46.9 | 53.1  | 63.7  | 81.2  | 89.2  | 106.0 |
| 4                    | 5.9 | 12.7 | 15.5 | 24.4 | 32.4 | 40.0 | 50.1 | 57.5  | 70.6  | 81.2  | 89.2  | 106.0 |
| 6                    | 5.9 | 12.7 | 15.6 | 24.8 | 33.5 | 42.0 | 52.4 | 60.8  | 75.8  | 81.2  | 89.2  | 106.0 |
| 8                    | 5.9 | 12.7 | 15.6 | 25.0 | 34.3 | 43.5 | 54.6 | 63.7  | 80.6  | 95.1  | 106.6 | 115.9 |
| 10                   | 5.9 | 12.7 | 15.7 | 25.3 | 35.1 | 45.0 | 57.0 | 66.7  | 86.0  | 101.4 | 114.0 | 125.3 |
| 12                   | 5.9 | 12.8 | 15.7 | 25.7 | 36.3 | 46.7 | 59.8 | 70.5  | 91.4  | 107.7 | 121.3 | 134.6 |
| 14                   | 5.9 | 12.8 | 15.8 | 26.8 | 38.0 | 49.0 | 64.4 | 75.9  | 97.2  | 114.1 | 130.3 | 143.9 |
| 16                   | 6.0 | 13.1 | 16.7 | 28.6 | 40.4 | 52.4 | 69.0 | 80.9  | 103.0 | 121.7 | 139.3 | 153.3 |
| 18                   | 6.1 | 13.4 | 17.2 | 30.5 | 43.6 | 56.5 | 72.0 | 84.6  | 108.5 | 129.6 | 148.2 | 162.8 |
| 20                   | 6.3 | 13.9 | 18.1 | 31.9 | 45.5 | 58.9 | 75.8 | 87.3  | 113.0 | 135.2 | 155.5 | 171.4 |
| 22                   | 6.4 | 14.0 | 18.2 | 32.2 | 46.0 | 59.9 | 77.1 | 90.7  | 116.5 | 139.1 | 159.6 | 177.1 |
| 24                   | 6.6 | 14.5 | 19.0 | 34.0 | 48.6 | 63.0 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 26                   | 6.6 | 14.5 | 19.0 | 34.1 | 49.0 | 63.7 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 28                   | 6.6 | 14.6 | 19.1 | 34.2 | 49.2 | 64.0 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 30                   | 6.6 | 14.6 | 19.1 | 34.3 | 49.3 | 64.1 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 32                   | 6.6 | 14.6 | 19.1 | 34.3 | 49.3 | 64.1 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 34                   | 6.6 | 14.6 | 19.1 | 34.3 | 49.3 | 64.1 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 36                   | 6.6 | 14.6 | 19.1 | 34.3 | 49.3 | 64.1 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 38                   | 6.6 | 14.6 | 19.1 | 34.3 | 49.3 | 64.1 | 81.1 | 97.1  | 118.8 | 142.4 | 163.5 | 181.3 |
| 40                   | 6.7 | 14.7 | 19.4 | 34.8 | 50.1 | 65.3 | 80.2 | 94.6  | 122.4 | 147.2 | 170.3 | 191.1 |
| 42                   | 6.7 | 14.8 | 19.4 | 34.9 | 50.3 | 65.6 | 80.5 | 95.1  | 122.9 | 148.0 | 171.3 | 192.3 |
| 44                   | 6.7 | 14.8 | 19.5 | 35.0 | 50.5 | 65.8 | 80.8 | 95.5  | 123.5 | 148.7 | 172.3 | 193.6 |
| 46                   | 6.7 | 14.8 | 19.5 | 35.2 | 50.7 | 66.1 | 81.2 | 95.9  | 124.1 | 149.5 | 173.3 | 194.8 |
| 48                   | 6.7 | 14.9 | 19.6 | 35.3 | 50.9 | 66.3 | 81.5 | 96.3  | 124.7 | 150.3 | 174.3 | 196.0 |
| 50                   | 6.7 | 14.9 | 19.6 | 35.4 | 51.1 | 66.6 | 81.8 | 96.7  | 125.2 | 151.0 | 175.2 | 197.2 |
| 52                   | 6.7 | 14.9 | 19.7 | 35.5 | 51.3 | 66.8 | 82.2 | 97.1  | 125.8 | 151.8 | 176.1 | 198.3 |
| 54                   | 6.7 | 15.0 | 19.7 | 35.6 | 51.4 | 67.1 | 82.5 | 97.5  | 126.3 | 152.5 | 177.0 | 199.5 |
| 56                   | 6.8 | 15.0 | 19.8 | 35.7 | 51.6 | 67.3 | 82.8 | 97.8  | 126.8 | 153.2 | 177.9 | 200.6 |
| 58                   | 6.8 | 15.0 | 19.8 | 35.8 | 51.8 | 67.6 | 83.1 | 98.2  | 127.4 | 153.8 | 178.8 | 201.6 |
| 60                   | 6.8 | 15.0 | 19.9 | 36.0 | 51.9 | 67.8 | 83.4 | 98.6  | 127.8 | 154.5 | 179.6 | 202.6 |
| 62                   | 6.8 | 15.1 | 19.9 | 36.1 | 52.1 | 68.0 | 83.6 | 98.9  | 128.3 | 155.1 | 180.4 | 203.6 |
| 64                   | 6.8 | 15.1 | 20.0 | 36.2 | 52.3 | 68.2 | 83.9 | 99.2  | 128.8 | 155.7 | 181.2 | 204.6 |
| 66                   | 6.8 | 15.1 | 20.0 | 36.3 | 52.4 | 68.4 | 84.2 | 99.6  | 129.3 | 156.3 | 182.0 | 205.5 |
| 68                   | 6.8 | 15.2 | 20.1 | 36.3 | 52.6 | 68.6 | 84.4 | 99.9  | 129.7 | 156.9 | 182.7 | 206.5 |
| 70                   | 6.8 | 15.2 | 20.1 | 36.4 | 52.7 | 68.8 | 84.7 | 100.2 | 130.1 | 157.5 | 183.4 | 207.4 |
| 72                   | 6.8 | 15.2 | 20.1 | 36.5 | 52.8 | 69.0 | 84.9 | 100.5 | 130.6 | 158.1 | 184.1 | 208.2 |
| 74                   | 6.9 | 15.2 | 20.2 | 36.6 | 53.0 | 69.2 | 85.2 | 100.8 | 131.0 | 158.6 | 184.8 | 209.1 |
| 76                   | 6.9 | 15.3 | 20.2 | 36.7 | 53.1 | 69.4 | 85.4 | 101.1 | 131.4 | 159.1 | 185.5 | 209.9 |
| 78                   | 6.9 | 15.3 | 20.3 | 36.8 | 53.3 | 69.6 | 85.7 | 101.4 | 131.8 | 159.7 | 186.2 | 210.7 |
| 80                   | 6.9 | 15.3 | 20.3 | 36.9 | 53.4 | 69.8 | 85.9 | 101.7 | 132.2 | 160.2 | 186.8 | 211.5 |
| 82                   | 6.9 | 15.3 | 20.3 | 37.0 | 53.5 | 69.9 | 86.1 | 101.9 | 132.6 | 160.7 | 187.4 | 212.3 |
| 84                   | 6.9 | 15.3 | 20.4 | 37.0 | 53.6 | 70.1 | 86.3 | 102.2 | 132.9 | 161.2 | 188.0 | 213.0 |
| 86                   | 6.9 | 15.4 | 20.4 | 37.1 | 53.8 | 70.3 | 86.5 | 102.5 | 133.3 | 161.7 | 188.7 | 213.8 |
| 88                   | 6.9 | 15.4 | 20.4 | 37.2 | 53.9 | 70.5 | 86.8 | 102.8 | 133.7 | 162.2 | 189.2 | 214.5 |
| 90                   | 6.9 | 15.4 | 20.5 | 37.2 | 54.0 | 70.5 | 86.9 | 102.9 | 133.9 | 162.4 | 189.5 | 214.8 |
| 92                   | 6.9 | 15.4 | 20.5 | 37.3 | 54.0 | 70.6 | 86.9 | 103.0 | 134.0 | 162.5 | 189.7 | 215.0 |
| 94                   | 6.9 | 15.4 | 20.5 | 37.3 | 54.0 | 70.6 | 87.0 | 103.1 | 134.1 | 162.6 | 189.9 | 215.3 |
| 96                   | 6.9 | 15.4 | 20.5 | 37.3 | 54.1 | 70.7 | 87.1 | 103.1 | 134.2 | 162.8 | 190.1 | 215.5 |
| 98                   | 6.9 | 15.4 | 20.5 | 37.3 | 54.1 | 70.7 | 87.1 | 103.2 | 134.3 | 162.9 | 190.2 | 215.7 |
| 100                  | 6.9 | 15.4 | 20.5 | 37.3 | 54.1 | 70.7 | 87.2 | 103.2 | 134.3 | 163.0 | 190.3 | 215.7 |

## 5 Tasks

To run the exercise by hand you should first produce two paper copies of the records to be analyzed (Figures 2 and 3, respectively) and a transparent overlay copy of the differential travel-time curve in Figure 4 and. All copies should be made so as to have identical time scale.

### 5.1 Event record No. 1 (Figure 2)

5.1.1 Assess, whether the source was shallow ( $< 70$  km) or deep (Look for surface waves!)

- Shallow source?
- Deep source?

5.1.2 Look for possible depth phases. Are there any clear depth phases?

- Yes?
- No?
- Comments on the possible depth range of the EQ if you suspect pP and/or sP to arrive in the complex wave group after P?

5.1.3 If you do not find any clear depth phases use the differential travel-time curve in Figure 4 (which is for surface foci), and try to identify the principal phases in the vertical and horizontal component record.

- Which phases you have identified?
- Give reasons for your interpretation?

5.1.4 a) Measure the time difference  $t(S-P)$ , b) estimate the epicentral distance  $D$  by using Equation (1) and c) by matching  $t(S-P)$  with the  $t-D$  curve of Figure 4.

- a)  $t(S-P) =$
- b)  $D = \dots\dots^\circ$
- c)  $D = \dots\dots^\circ$

5.1.5 Determine a) the quadrant of P-wave first motion horizontal particle motion, b) AZI from the amplitude ratio  $|A_E/A_N|$  using Equation (1), and c) calculate BAZ taking into account the relationships between AZI, quadrant and BAZ in Table 1.

- a) Quadrant =
- b) AZI (MOX) =  $\dots\dots^\circ$
- c) BAZ(MOX) =  $\dots\dots^\circ$

5.1.6 Locate the epicenter on the map given in Figure 5 and name the source region.

- Source region?
- Discussion?

## 5.2 Event record No. 2 (Figure 3)

5.2.1 Assess, whether the event was shallow ( $< 70$  km) or deep by looking for possible surface waves in the full record, which is inserted at strongly compressed time scale.

- Shallow event?
- Deep event?

5.2.2 Look for possible depth phases. Are there any clear depth phases?

- Yes?
- No?
- Comments?

5.2.3 Measure the time difference (in min) between the P-wave first arrival and the five marked onsets of stronger secondary wave arrivals in the records. Write down the time differences  $X_i$ -P (in min) in the order of their appearance.

- $X_1$ -P = ,  $X_2$ -P = ,  $X_3$ -P = ,  $X_4$ -P = ,  $X_5$ -P = ?

5.2.4 In order to match these travel-time differences with the differential t-D curve in Fig. 4 plot them with the same time scale as your overlay on the edge of a sheet of paper. Then place the P-wave onset mark on distance scale and move it there along until you match the onset marks of the later onsets with relevant travel-time curves in Figure 4. a) Read the distance D (in  $^\circ$ ) related to this best fit with as many as possible of your onset marks and b) identify the related phases and write their names on the onsets!

a)  $D$  (CLL) = ..... $^\circ$ ,

b)  $X_1$  = ,  $X_2$  = ,  $X_3$  = ,  $X_4$  = ,  $X_5$  =

- Comments, which support your phase interpretations ?

5.2.5 Determine a) the quadrant of particle motion, b) AZI and c) BAZ as in task 4.1.5.

a) Quadrant =

b) AZI (CLL) = .....  $^\circ$

c) BAZ(CLL) = .....  $^\circ$

5.2.6 Locate the event as in task 5.1.6 by using Figure 6

- Name of source region/country?

## 6 Solutions

**Note:** Your estimates for the travel-time differences should be within about 0.2 min, for D within  $2^\circ$  and for the backazimuth AZI within about  $5^\circ$  of the solutions given below. They were derived by hand by the authors of this tutorial exercise, following the outlined procedures. The solutions given below are numbered according to the tasks in section 5.

## Event No. 1:

## 5.1.1

The earthquake is deeper than 70 km because no long-period and dispersive surface waves have been recorded.

## 5.1.2

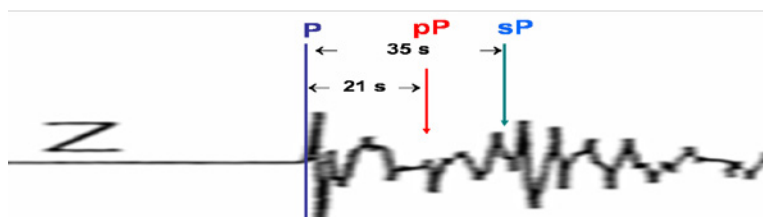
On this record with low time resolution it is difficult to recognize for sure depth phases. However, they might arrive within the complex wave group that follows within about 40 s after the P onset. The ISC calculated a hypocenter depth  $h = 111$  km and an epicentral distance  $D = 20.24^\circ$  for station MOX. According to the IASP91 (Kennett, 1991) differential travel-time table for depth phases pP should then arrive about 21 s and sP about 35 s after P.

| pP-P     |     | Depth of source [km] |      |      |      |      |      |      |      |      |      |      |      |
|----------|-----|----------------------|------|------|------|------|------|------|------|------|------|------|------|
| $\Delta$ | 15. | 35.                  | 50.  | 100. | 111  | 150. | 200. | 250. | 300. | 400. | 500. | 600. | 700. |
|          | s   | s                    | s    | s    | s    | s    | s    | s    | s    | s    | s    | s    | s    |
| 2        | 3.6 | 7.5                  |      |      |      |      |      |      |      |      |      |      |      |
| 4        | 3.6 | 7.5                  |      |      |      |      |      |      |      |      |      |      |      |
| 6        | 3.6 | 7.5                  |      |      |      |      |      |      |      |      |      |      |      |
| 8        | 3.6 | 7.5                  |      |      |      |      |      |      |      |      |      |      |      |
| 10       | 3.6 | 7.6                  |      |      |      |      |      |      |      |      |      |      |      |
| 12       | 3.6 | 7.6                  | 7.9  |      |      |      |      |      |      |      |      |      |      |
| 14       | 3.6 | 7.6                  | 8.0  |      |      |      |      |      |      |      |      |      |      |
| 16       | 3.8 | 8.1                  | 9.3  | 20.1 | 26.2 |      |      |      |      |      |      |      |      |
| 18       | 4.0 | 8.5                  | 10.1 | 16.2 | 25.9 | 32.6 | 38.8 |      |      |      |      |      |      |
| 20       | 4.3 | 9.2                  | 11.5 | 19.0 | 25.6 | 33.2 | 39.6 | 45.4 |      |      |      |      |      |
| 22       | 4.3 | 9.3                  | 11.7 | 19.4 | 26.9 | 34.4 | 41.7 | 48.4 | 59.9 |      |      |      |      |
| 24       | 4.5 | 10.0                 | 12.7 | 21.4 | 29.8 | 37.8 | 45.3 | 52.2 | 64.1 | 82.7 |      |      |      |
| 26       | 4.6 | 10.0                 | 12.8 | 22.1 | 31.3 | 40.2 | 48.7 | 55.8 | 68.2 | 83.4 | 93.4 |      |      |

| sP-P     |     | Depth of source [km] |      |      |      |      |      |      |       |       |       |       |  |
|----------|-----|----------------------|------|------|------|------|------|------|-------|-------|-------|-------|--|
| $\Delta$ | 15. | 35.                  | 50.  | 100. | 150. | 200. | 250. | 300. | 400.  | 500.  | 600.  | 700.  |  |
|          | s   | s                    | s    | s    | s    | s    | s    | s    | s     | s     | s     | s     |  |
| 2        | 5.9 | 12.7                 | 15.4 | 22.8 | 27.0 | 35.7 |      |      |       |       |       |       |  |
| 4        | 5.9 | 12.7                 | 15.5 | 24.4 | 32.4 | 40.0 | 46.9 | 53.1 | 63.7  |       |       |       |  |
| 6        | 5.9 | 12.7                 | 15.6 | 24.8 | 33.5 | 42.0 | 50.1 | 57.5 | 70.6  | 81.2  | 89.2  |       |  |
| 8        | 5.9 | 12.7                 | 15.6 | 25.0 | 34.3 | 43.5 | 52.4 | 60.8 | 75.8  | 88.6  | 98.5  | 106.0 |  |
| 10       | 5.9 | 12.7                 | 15.7 | 25.3 | 35.1 | 45.0 | 54.6 | 63.7 | 80.6  | 95.1  | 106.6 | 115.9 |  |
| 12       | 5.9 | 12.8                 | 15.7 | 25.7 | 36.3 | 46.7 | 57.0 | 66.7 | 86.0  | 101.4 | 114.0 | 125.3 |  |
| 14       | 5.9 | 12.8                 | 15.8 | 26.8 | 38.0 | 49.0 | 59.8 | 70.5 | 91.4  | 107.7 | 121.3 | 134.6 |  |
| 16       | 6.0 | 13.1                 | 16.7 | 28.6 | 40.4 | 52.4 | 64.4 | 75.9 | 97.2  | 114.1 | 130.3 | 143.9 |  |
| 18       | 6.1 | 13.4                 | 17.2 | 30.5 | 43.6 | 56.5 | 69.0 | 80.9 | 103.0 | 121.7 | 139.3 | 153.3 |  |
| 20       | 6.3 | 13.9                 | 18.1 | 31.9 | 45.5 | 58.9 | 72.0 | 84.6 | 108.5 | 129.6 | 148.2 | 162.8 |  |
| 22       | 6.4 | 14.0                 | 18.2 | 32.2 | 46.0 | 59.9 | 73.8 | 87.3 | 113.0 | 135.2 | 155.5 | 171.4 |  |
| 24       | 6.6 | 14.5                 | 19.0 | 34.0 | 48.6 | 63.0 | 77.1 | 90.7 | 116.5 | 139.1 | 159.6 | 177.1 |  |

Accordingly, the first two sharp recognizable onsets after P on the Z component record are most likely the depth phases pP and sP (see Figure 7), not PP and PPP, as plotted in the simplified J-B-differential t-D curves of Figure 4. According to more recent travel-time models such as IASP91 (see DS 2.1) PP and PPP will appear only for  $D \geq 30^\circ$ .





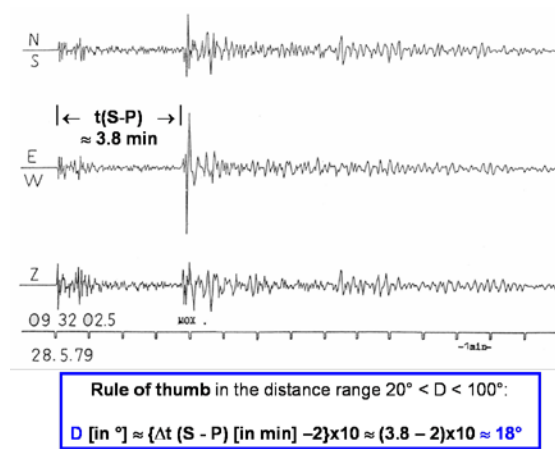
**Figure 7** Cutout of the vertical component record in Figure 2, stretched in time-scale. This allows better recognition of the depth phase onsets pP and sP at the expected arrival times after P for an earthquake at 111 km depth (as calculated by the ISC).

### 5.1.3

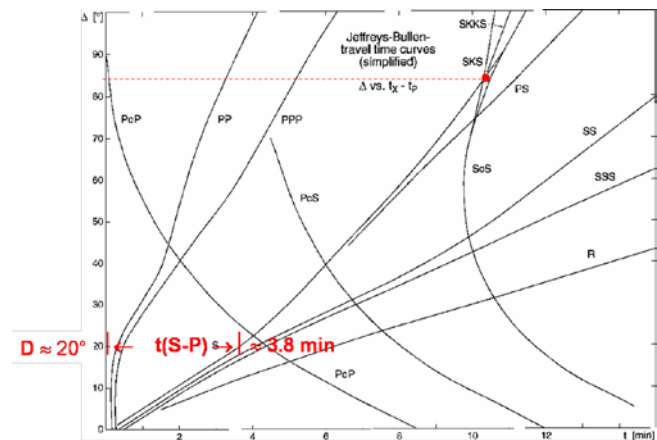
You should have identified at least P on the Z component, the S wave as the largest onset on the E component, and SS (with longer period than S) on the N component.

### 5.1.4

The results are shown in Figures 8 and 9. Note that the ISC calculated for MOX, based on the global network location,  $D = 20.24^\circ$ .



**Figure 8**



**Figure 9**

### 5.1.5

- 2<sup>nd</sup> quadrant of P-wave first particle motion;
- $\text{AZI}(\text{MOX}) = \arctan(|A_E/A_N|) \approx \arctan(4/3) \approx 53^\circ$
- $\text{BAZ} \approx 125^\circ$ .

### 5.1.6

Using the results from 5.1.4 and 5.1.5 (Figure 9) together with the map in Figure 5 the earthquake was located in the coastal area of **southern Turkey**. Locating a sub-crustal earthquake there makes sense, because the African Plate is sub-ducted underneath southern Turkey. Figure 10 shows the excellent agreement of the single station 3-component broadband solution with the global network solution of the ISC. **The two epicenter positions differ by only about  $0.7^\circ$ !**

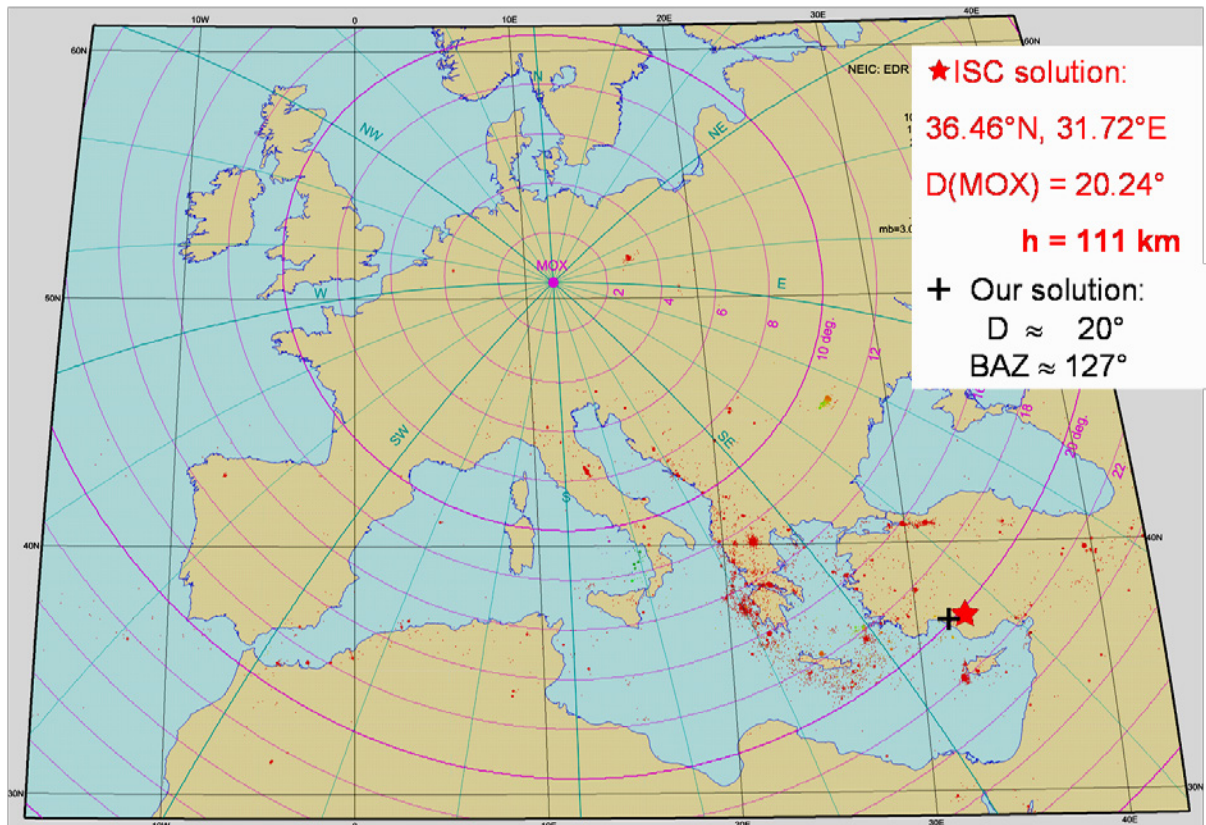


Figure 10

**Event No. 2:**

## 5.2.1

Shallow teleseismic earthquake with strong surface waves (see insert of Figure 3;  $A_{\max}$  after about 37 min).

## 5.2.2

No depth phases recognizable in the LP records (NEIC reported for this earthquake a hypocenter depth of  $h = 19$  km).

## 5.2.3

$X1-P \approx 3.65$  min,  $X2-P \approx 10.5$  min,  $X3-P \approx 11.3$  min,  $X4-P \approx 12.2$  min,  $X5-P \approx 17$  min

## 5.2.4

- $D(CLL) \approx 93^\circ \pm 1^\circ$  for the best match of the travel-time differences given under 5.2.3 with the with the travel-time curve shown in Figure 4. NEIC-PDE gives for station CLL  $92.6^\circ$ .
- The identified phases, which match best the travel-time differences to P given under 5.2.3 are:  $X1 = PP$ ,  $X2 = SKS$ ,  $X3 = S$ ,  $X4 = PS/SP$ ,  $X5 = SS$
- Both SKS and PS are strongest in the horizontal component E where also P has its largest (radial) horizontal amplitude. At about the time of the PS arrival in N there appears also in Z a clear energy arrival (SP↓). S arrives later than SKS and is strongest in the N (transverse) component only.

The summary results are plotted in Figure 11.

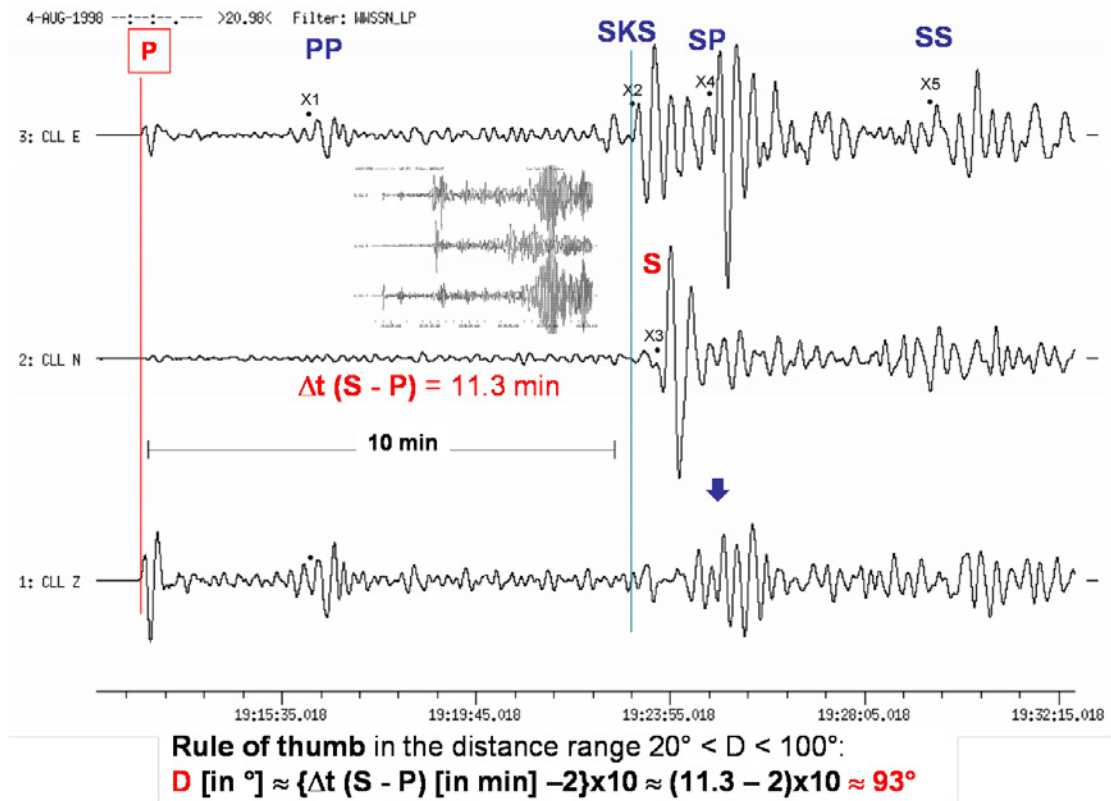


Figure 11

## 5.2.5

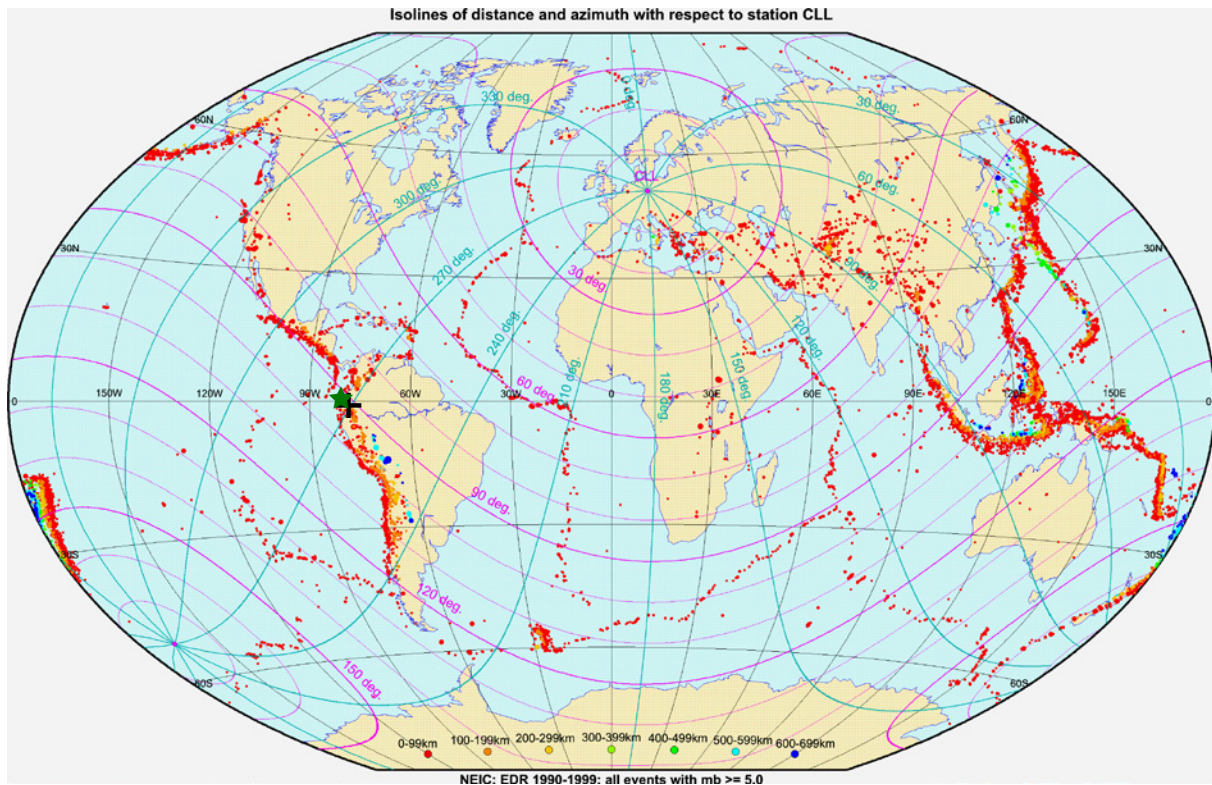
- Quadrant ? Transition from 1<sup>st</sup> to 2<sup>nd</sup> because of large eastward P-wave motion; first motion in N-S component not measurable
- AZI (CLL)  $\approx 90^\circ$
- BAZ (CLL)  $\approx 270^\circ$  because + first motion in Z and E relates to a source in the W.

**Comment:** Recognizable P-wave upward (+) motion in the N component begins only more than 6 s later than in the Z component, thus being related to the strong downward (-) P-wave motion! Therefore, the azimuth might be a few degrees further north. In fact, NEIC-PDE gives for CLL AZI =  $272.3^\circ$ .

## 5.2.6

Using the map in Figure 6 for station CLL the source is located near **Ecuador**. NEIC-PDE gives as epicenter coordinates  $0.59^\circ\text{S}$  and  $80.39^\circ\text{W}$ , i.e., near coast of Ecuador. The difference between these two locations is  $< 3^\circ$  (see Figure 12) and would be even less, if the weak N-S first motion could have been resolved by appropriate interactive amplitude scaling.





+ CLL solution  $\Rightarrow$  **Ecuador,  $D \approx 93^\circ$ , BAZ  $\approx 270^\circ$  + a few degrees N?**

★ NEIC solution: **Ecuador,  $0.59^\circ\text{S}$ ,  $80.39^\circ\text{W}$ ;  $D(\text{CLL}) = 92.6^\circ$ ,  $\text{BAZ}(\text{CLL}) = 272.3^\circ$**

Figure 12

## Acknowledgment

Figures 2 and 4 as well as Table 2 had been compiled by W. Strauch and K. Wylegalla for an earlier version of this exercise.

## References

- Jeffreys, H., and Bullen, K. E. (1940, 1948, 1958, 1967, and 1970). Seismological Tables. British Association for the Advancement of Science, *Gray Milne Trust*, London, 50 pp.
- Kennett, B. L. N. (Ed.) (1991). IASPEI 1991 Seismological Tables. Research School of Earth Sciences, Australian National University, 167 pp.

| Topic   | Identification and analysis of short-period core phases<br>(tutorial with exercise by hand)                                                                                                                                                                                                                                                                                                           |
|---------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a><br><b>Siegfried Wendt</b> , Universität of Leipzig, Institut für Geophysik und Geologie, Geophysikalisches Observatorium Collm, D-04779 Wermsdorf, Germany, E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a> |
| Version | June 2002; DOI: 10.2312/GFZ.NMSOP-2_EX_11.3                                                                                                                                                                                                                                                                                                                                                           |

## 1 Aim

This manual exercise aims at making you familiar with the identification of both direct and multiple-reflected longitudinal core phases and their use in location and magnitude determination. Clear short-period vertical component PKP seismograms of a single station contain all information needed to determine *source depth*  $h$ , *epicentral distance*  $D$  and *magnitude*  $m_b$  with an accuracy of  $\pm 30$  km, better  $\pm 1.5^\circ$  and  $\pm 0.3$  magnitude units, respectively. In case of strong seismic sources and the availability of identically calibrated horizontal components with good signal-to-noise ratio, additionally the backazimuth to the source can be determined with an accuracy of about  $\pm 5^\circ$ -  $10^\circ$  and thus the approximate location. Additionally, the identification of late reflected core phases and their use in distance determination is practiced. These phases are very suitable for calculating the epicentral distance since their relative travel-time difference to the related first arrival P or PKP is nearly independent of source depth.

## 2 Data

- Figure 1: Compilation of typical analog short-period recordings at station MOX, Germany, of the different direct core phases from earthquakes between  $135^\circ < D < 160^\circ$ ;
- Figure 2: Plots of digital broadband records of the German Regional Seismograph Network (GRSN), filtered according to a WWSSN-SP response, from a Fiji-Island earthquake within the distance range  $148.4^\circ$  (CLL) to  $152.2^\circ$  (BFO);
- Figure 3: Three record examples with PKPab, bc and df phases to be analyzed;
- Figure 4: Four records with later longitudinal core phases to be evaluated;
- Figure 5: Travel times and paths of the direct longitudinal core phases and their relationship to the P-wave velocity model of the Earth;
- Figure 6: Ray path of the reflected core phases  $P'P'$  (or PKPPKP) and PKKP;
- Figure 7: Differential travel-time curves pPKP-PKP for  $D = 150^\circ$ ;
- Figure 8: Differential travel-time curves PKPbc-PKPdf (PKP1-PKIKP) and PKPab-PKPdf (PKP2-PKIKP);
- Figure 9: Differential travel-time curves PKKP-P and PKKP-PKP, respectively;
- Figure 10: Differential travel-time curves PKPPKP-P.
- Figure 11: Record with identified onsets.
- Figure 12: Magnitude calibration functions for PKPdf (old PKIKP), PKPbc (old PKP1) and PKPab (old PKP2).

Note 1: All differential travel-time curves given in this exercise have been calculated according to the earth model IASP91 (Kennett and Engdahl, 1991). The more recent model AK135 (Kennett et al., 1995) yields still better travel-times for core phases. The difference, however, is more significant for absolute and usually negligible for differential travel times.

Note 2: The first record example in Figure 1 illustrates that at  $D < 145^\circ$  small amplitude precursors PKPpre of waves scattered from the core-mantle boundary (CMB) may occur. In the case of crustal earthquakes PKPdif may additionally be followed closely by depth phases. Together this may mimic a core phase triplication typical for  $D > 146^\circ$ . Yet, in the case of deep earthquakes with sharp onsets and no or small signal coda, the triple group of phases is usually rather distinct and its typical pattern easily recognizable.

Note 3: The typical three-phase pattern PKPdif (alternative name PKIKP), PKPbc (old name PKP1) and PKPab (old name PKP2) is however well developed between  $146^\circ < D < 155^\circ$  only. Around  $145^\circ$  all three phases arrive at the same time and thus superpose to a rather strong impulsive onset.

Note 4: Beyond  $154^\circ$  still a weak intermediate phase between PKPdif and PKPab may be observed up to about  $160^\circ$  along the extrapolation of the PKPbc travel-time branch. It is, however, not PKPbc proper but rather the phase PKPdif which is diffracted around the inner-core boundary (see last record example in Figure 1).

Note 5: In the last record example of Figure 1 the well developed depth phase obviously relates to the strongest direct phase PKPab (old PKP2). The depth phases pPKP may, in the case of maximum possible source depth around 700 km, follow the related direct phases after up to about 2.5 minutes. When the primary core phases are rather strong, two or three related depth phases may be discernable.

Note 6: More examples based on plots from digital seismic records are given in DS 11.3.

Additionally, for the magnitude determinations, measured trace amplitudes have to be converted into ground motion amplitudes. For this the frequency dependent amplitude-magnification of the seismograph has to be known. It is given in Table 1.

**Table 1** Magnification MAG of ground displacement in the seismic records of earthquakes No. 1, 2 and 3 in Figure 3 when these are reproduced with a time scale of 1 mm/s.

| Period (in s) | 0.9    | 1.0    | 1.1    | 1.2    | 1.3    | 1.4    | 1.6    | 1.8    | 2.0    | 2.4    |
|---------------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| MAG Event 1   |        | 52,440 |        | 49,210 |        | 44,090 | 37,550 | 30,660 | 24,420 | 14,000 |
| MAG Event 2   |        | 26,220 |        | 24,600 |        | 22,040 | 18,770 | 15,330 | 12,210 | 7,000  |
| MAG Event 3   | 62,020 | 56,000 | 49,980 | 43,960 | 37,940 |        |        |        |        |        |

### 3 Procedure

All steps of determining  $h$ ,  $D$ ,  $AZI$  and  $mb$  will be practiced.

**Source depth** and **epicentral distance** are determined by reading the relative onset-time difference in seconds or minutes, respectively, between identified phases and using the corresponding differential travel-time curves. Note that these curves in Figures 7 and 8 have been presented with the same time-resolution as the analog records in Figure 3, i.e. with 1

mm/s. If a transparent overlay is produced from these curves, the source depth and the epicentral distance, respectively, from the records depicted in Figure 3 can be directly determined by matching the related curves with the identified onsets. Make sure that the depth and distance axes, respectively, are kept perpendicular to the center line of the record trace when matching. When reading the time differences PKKP-P and P'P'-P in Figure 4 be aware, that the time difference between subsequent traces from top to bottom is 15 minutes. Full minutes start at the left side of the 2 second long gaps or „faintings“ on the record traces.

The **backazimuth** is determined according to the instructions given under 3.2 in EX 11.2.

For epicenter **location** based on the estimated epicentral distance D and the backazimuth AZI one may use either a sufficiently large globe (diameter about 0.5 to 1 m), mark there the position of the considered station and then use a bendable ruler with the same scale in degree as your globe and an azimuth dial to find the source location on the globe. Another possibility is to use a global map projection which shows isolines of equal backazimuth and distance from your station (as Figure 5 in EX 11.2). Such maps can nowadays easily be calculated and plotted by means of computers.

The **magnitude mb(PKP)** is determined according to an experimental calibration function for magnitude determinations based on short-period readings of various PKP phases in the distance range 145° to 164°. It has been developed by S. Wendt (Bormann and Wendt, 1999). Its world-wide testing is recommended. The following relationship is used:

$$mb(PKP) = \log_{10} (A/T) + Q(\Delta, h)_{PKP} \quad (1)$$

with amplitude A in  $\mu m$  ( $10^{-6}$  m). If more than one PKP phase PKPab, PKPbc and/or PKPdc can be identified and A and T been measured then the individual phase magnitudes should be determined first and then the average magnitude be calculated. The latter provides a more stable estimate.

## 4 Tasks

- 4.1 Train yourself first by matching the travel-time curve overlay of Figure 8 with the onsets marked in Figure 1, taking into account the distance and focal depth given for each earthquake. Also consult Figure 2 and the related notes 1 to 6 in section Data above. Then mark on the records in Figure 3 for all three earthquakes the onset times of recognizable phases and give them names, both for the early and the late arrivals (depth phases).
- 4.2 Measure the time difference pPKP-PKP for the strongest PKP arrival and its respective depth phase and determine the source depth for all three earthquakes by using the differential travel-time curves given in Figure 7. Note: If the pPKP group is less distinct and its different onsets can not be well separated then relate the depth phase to the strongest direct PKP arrival.
- 4.3 Determine the epicentral distance D (in °) of the three earthquakes shown in Figure 3 by using the differential travel-time curves shown in Figure 8 taking into account for each event the source depth determined under 4.2.

- 4.4 Measure the time difference between the P-wave first arrivals and the PKKP or P'P' phases marked in the four records presented and determine the epicentral distance  $D$  for these earthquakes by using the differential travel-time curves shown in Figures 9 and 10, respectively.
- 4.5 Estimate from the three-component record in Figure 3 (event No. 3) for the strongest phase the backazimuth AZI according to relationship and instructions given under 3.2 of EX 11.2.
- 4.6 Try to find for event No. 3, which had been recorded at the station CLL in Germany, the source location on the map shown in Figure 5 of EX 11.2. Give the name of the source area.
- 4.7 Measure the trace amplitudes  $B$  (in mm) and related periods  $T$  for all identified phases of direct PKP in Figure 3 and convert them into “true ground motion” amplitudes  $A$  (in nm) by means of Table 1 given above in section Data.
- 4.8 Determine  $\log(A/T)$  and estimate the related values for  $\sigma(D, h)$  from Figure 12. Note that these values are valid for amplitudes in  $\mu\text{m}$  only! Correct them for nm.
- 4.9 Give the individual magnitude estimates for PKPab, PKPbc and PKPdf,
- 4.10 Calculate the average  $\bar{m}_b(\text{PKP})$ .
- 4.11 Compare your results with the respective solutions given by the NEIC for these three earthquakes and assess the achievable accuracy of respective individual source parameter calculations at single stations, even when based on analog recordings only and simple analysis tools.

## 5 Solutions

5.1 See Figure 11.

5.2 NEIC gave for the three earthquakes the following hypocentral depths:

- No. 1  $h = 435 \text{ km}$
- No. 2  $h = 235 \text{ km}$
- No. 3  $h = 540 \text{ km}$

Your own depth estimates should be within about  $\pm 30 \text{ km}$  of these values. But this does not mean that your value is worse than that given by NEIC. It may be even better, because NEIC mostly does not use depth phases to constrain its solutions from direct P-wave readings.

5.3 NEIC calculated for the stations which recorded the three earthquakes in Figure 3 the following epicentral distances  $\Delta = D$  (in  $^\circ$ ):

- No. 1  $D = 148.5^\circ$
- No. 2  $D = 159.5^\circ$
- No. 3  $D = 150.3^\circ$

Using the recommended travel-time curves, your estimates should be within  $\pm 1.5^\circ$  of these values.

5.4 NEIC calculated for the station CLL which had recorded the earthquakes shown in Figure 4 the following epicentral distances:

- No. 1       $D = 98.5^\circ$   
 No. 2       $D = 111.2^\circ$   
 No. 3       $D = 66.3^\circ$   
 No. 4       $D = 57.3^\circ$

Using the recommended travel-time curves, your estimates should be within  $\pm 1.5^\circ$  to NEIC.

5.5  $AZI \approx 22^\circ$ . Your own estimate should be within  $\pm 5^\circ$  of this value.

#### 5.6 Fiji Islands

|                        |                                                   |
|------------------------|---------------------------------------------------|
| 5.7 Event No. 1: PKPdf | B = 0.5 mm, T = 1.5 s $\rightarrow$ A = 12.2 nm   |
| PKPbc                  | B = 3.9 mm, T = 1.0 s $\rightarrow$ A = 74.4 nm   |
| PKPab                  | B = 3.0 mm, T = 1.0 s $\rightarrow$ A = 57.2 nm   |
| Event No. 2: PKPdf     | B = 2.9 mm, T = 2.4 s $\rightarrow$ A = 414.3 nm  |
| PKPab                  | B = 8.2 mm, T = 2.0 s $\rightarrow$ A = 671.6 nm  |
| Event No. 3: PKPdf     | B = 0.95 mm, T = 1.3 s $\rightarrow$ A = 21.6 nm  |
| PKPbc                  | B = 20.4 mm, T = 1.0 s $\rightarrow$ A = 329.0 nm |
| PKPab                  | B = 6.5 mm, T = 1.0 s $\rightarrow$ A = 116.1 nm  |

|                        |                    |                                    |
|------------------------|--------------------|------------------------------------|
| 5.8 Event No. 1: PKPdf | $\log(A/T) = 0.9$  | $\sigma_{PKPdf}(\Delta, h) = 3.95$ |
| PKPbc                  | $\log(A/T) = 1.89$ | $\sigma_{PKPbc}(\Delta, h) = 3.15$ |
| PKPab                  | $\log(A/T) = 1.76$ | $\sigma_{PKPab}(\Delta, h) = 3.39$ |
| Event No. 2: PKPdf     | $\log(A/T) = 2.24$ | $\sigma_{PKPdf}(\Delta, h) = 3.8$  |
| PKPab                  | $\log(A/T) = 2.53$ | $\sigma_{PKPab}(\Delta, h) = 3.55$ |
| Event No. 3: PKPdf     | $\log(A/T) = 1.22$ | $\sigma_{PKPdf}(\Delta, h) = 3.94$ |
| PKPbc                  | $\log(A/T) = 2.52$ | $\sigma_{PKPbc}(\Delta, h) = 3.23$ |
| PKPab                  | $\log(A/T) = 2.06$ | $\sigma_{PKPab}(\Delta, h) = 3.55$ |

5.9 Event No. 1:  $mb(PKPdf) = 4.85$ ;  $mb(PKPbc) = 5.04$ ;  $mb(PKPab) = 5.15 \rightarrow \bar{mb}(PKP) = 5.0$   
 Event No. 2:  $mb(PKPdf) = 6.0$ ;  $mb(PKPab) = 6.1 \rightarrow \bar{mb}(PKP) = 6.0$   
 Event No. 3:  $mb(PKPdf) = 5.19$ ;  $mb(PKPbc) = 5.75$ ;  $mb(PKPab) = 5.61 \rightarrow \bar{mb}(PKP) = 5.5$

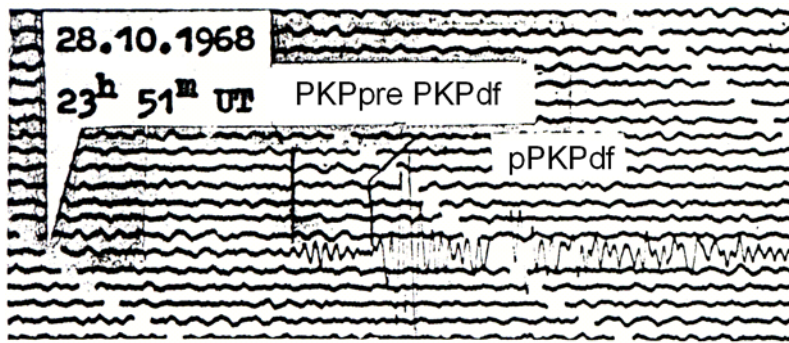
5.10 NEIC gave for these three events, based on teleseismic P-wave readings only:

- Event No. 1:  $mb = 5.0$ ,  
 Event No. 2:  $mb = 5.5$ ,  
 Event No. 3:  $mb = 5.3$

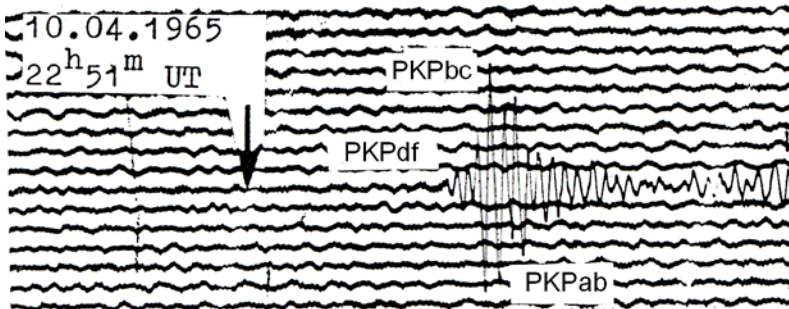
5.11 Using the calibration curves for PKP waves one can get quick  $mb$  estimates from readings of PKP amplitudes at individual stations with simple analysis tools which are within about  $\pm 0.5$  magnitudes units to the  $mb$  estimates of global seismological services. Your distance estimates should also be within  $\pm 1.5^\circ$  even when using only low-resolution analog data and visual time picks. The general source area can be determined properly, even for very



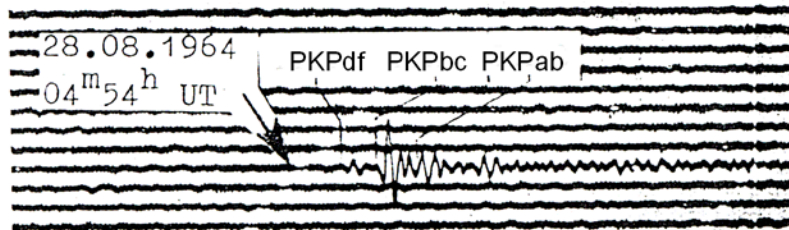
distant events, on the basis of properly mutually calibrated 3-component recordings of single seismic stations.



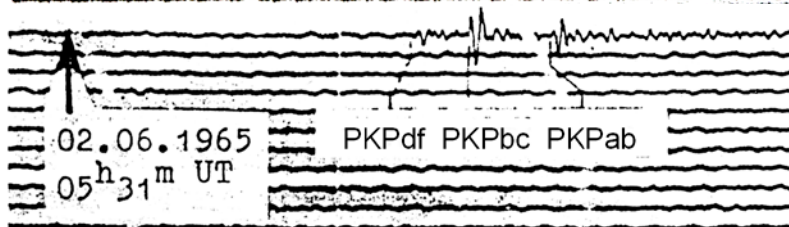
$D = 135.5^\circ$   
 $h = 60 \text{ km}$   
 $mb = 5.9$   
 Santa Cruz Islands



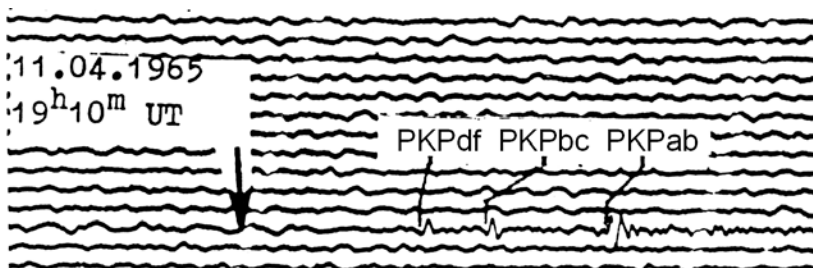
$D = 146.7^\circ$   
 $h = 543 \text{ km}$   
 $mb = 5.9$   
 Fiji Islands



$D = 148.3^\circ$   
 $h = 580 \text{ km}$   
 $mb = 5.4$   
 Fiji Islands

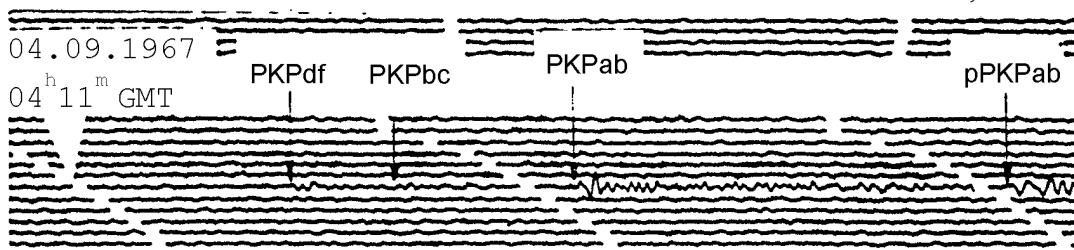


$D = 151.3^\circ$   
 $h = 539 \text{ km}$   
 $mb = 5.6$   
 Fiji Islands

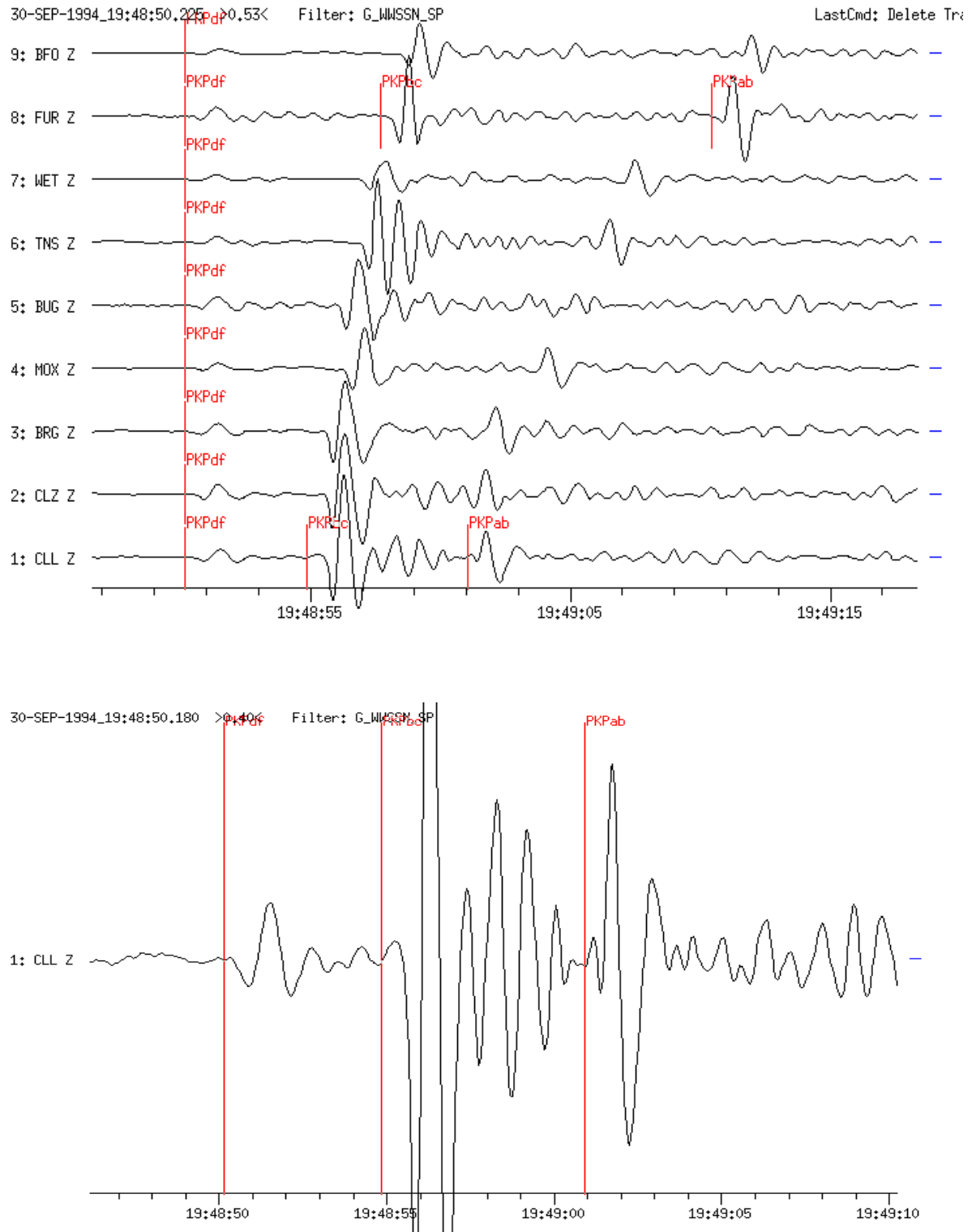


$D = 153.6^\circ$   
 $h = 581 \text{ km}$   
 $mb = 5.5$   
 South of Fiji Islands

$D = 159.5^\circ$ ,  $h = 231 \text{ km}$   
 $mb = 5.5$ , Kermadec Islands



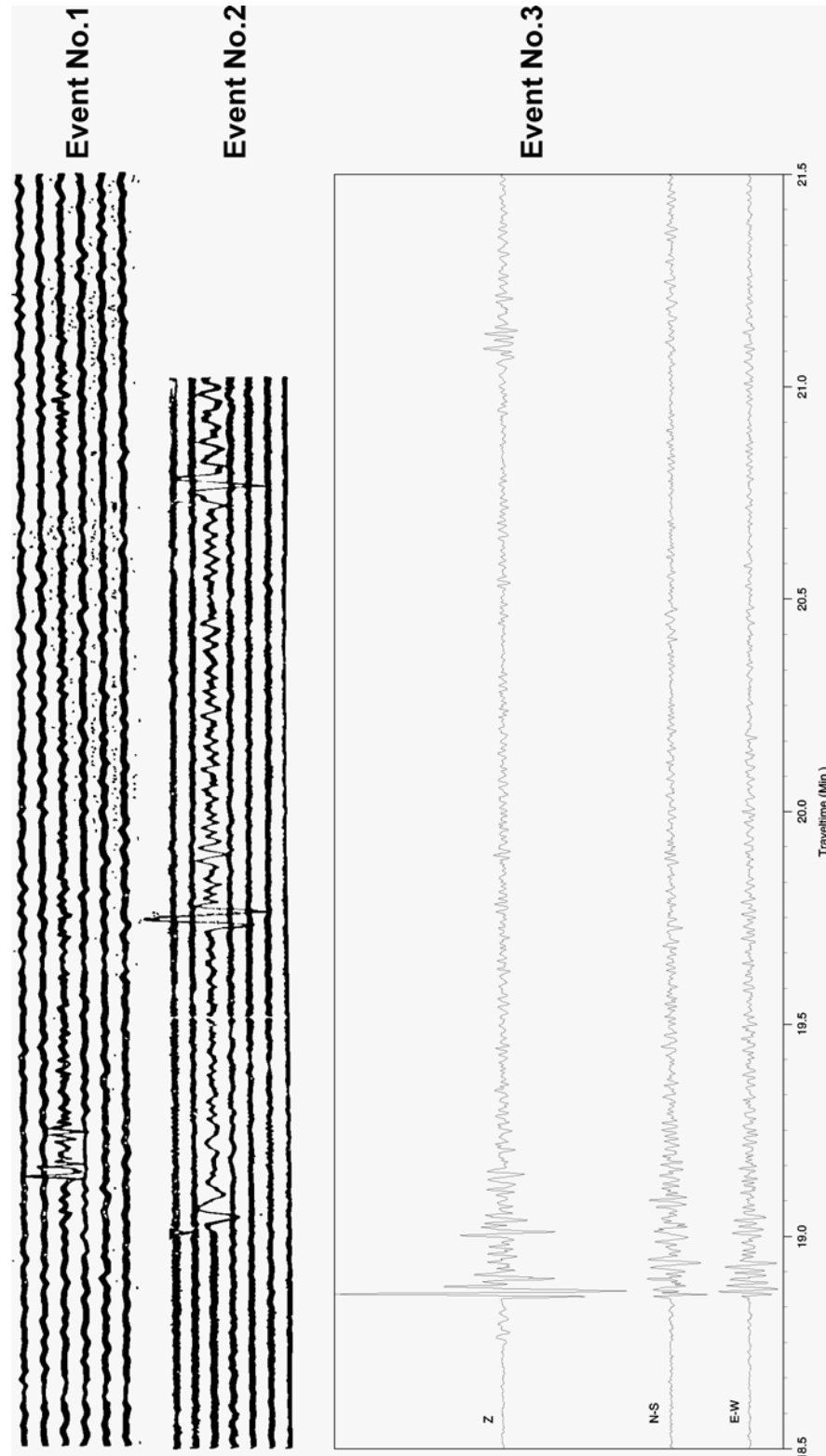
**Figure 1** Examples of short-period analog records of stations CLL and MOX of longitudinal core phases in the distance range  $135^\circ < D < 160^\circ$ . Time scale: 1 mm/s on all records.



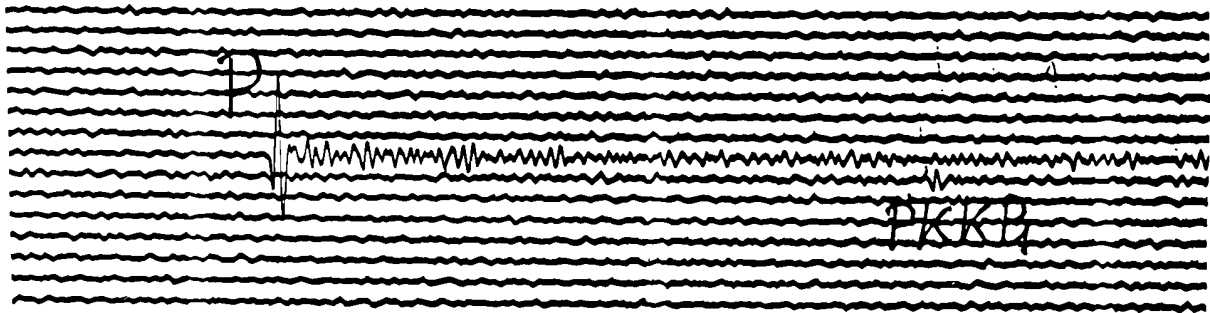
**Figure 2 Top:** Short-period filtered seismograms (WWSSN\_SP simulation) recorded at 9 GRSN stations from a deep earthquake in the Fiji Islands (Sept. 30, 1994,  $m_b = 5.1$ ,  $h = 643$  km). All traces are time-shifted and aligned with respect to PKPdf and sorted according to epicentral distance  $D$  which is  $148.4^\circ$  for CLL and  $152.2^\circ$  for the most distant station BFO.



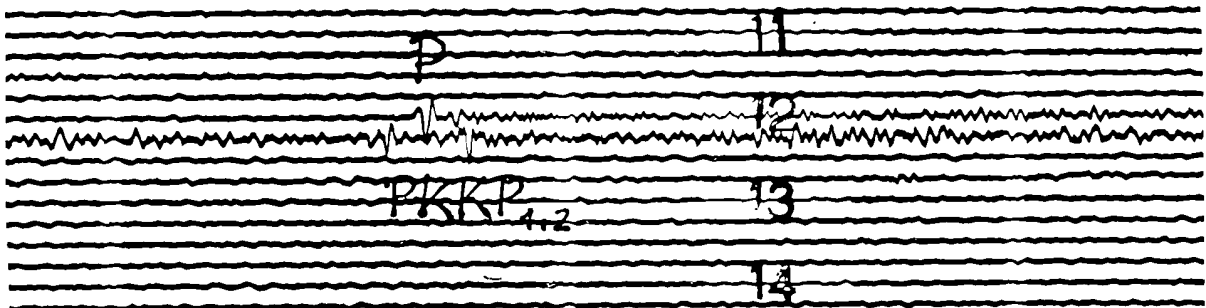
**Bottom:** The same trace of station CLL as above but with enlarged amplitude and time resolution.



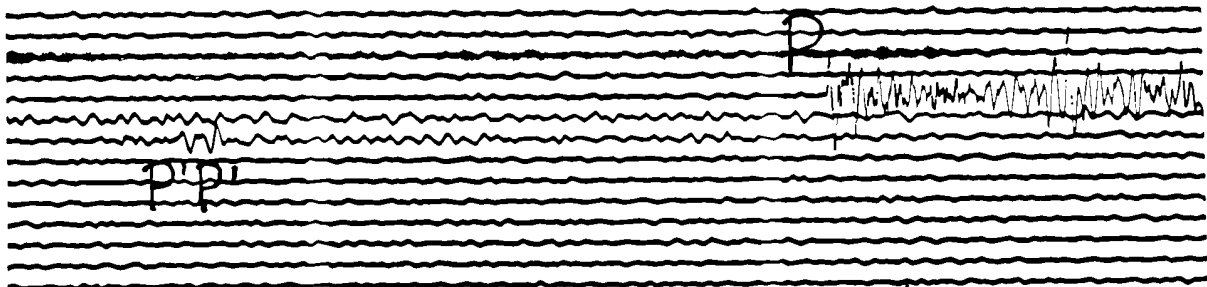
**Figure 3** Short-period records of direct longitudinal core phases from three earthquakes. Time scale: 1 mm/s if plotted 1:1.



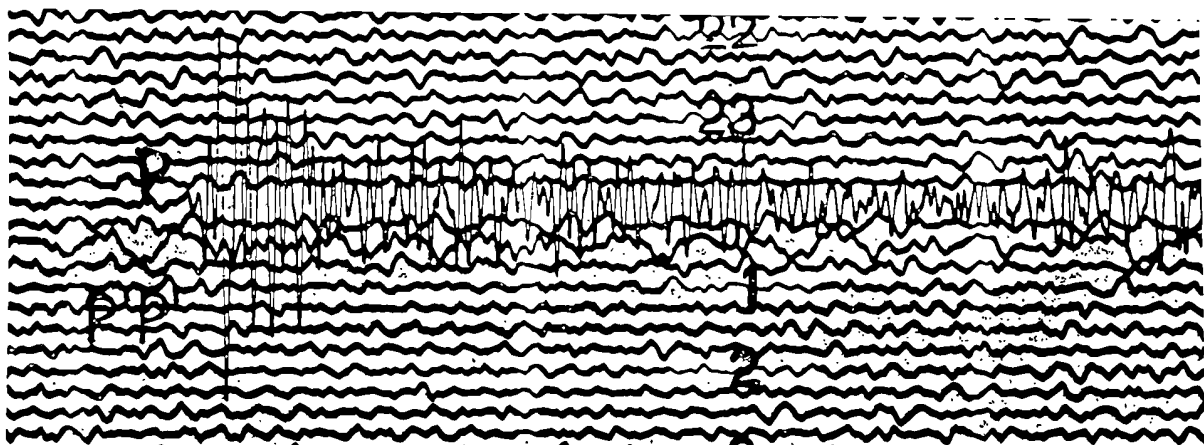
24.08.1995 Mariana Islands: H = 01:55:34.4, D(PLL) = ° ?, h = 588 km,  $m_b = 6.0$



13.07.1994 Banda Sea: H = 11:45:23.4, D(PLL) = ° ?, h = 159 km,  $m_b = 6.5$

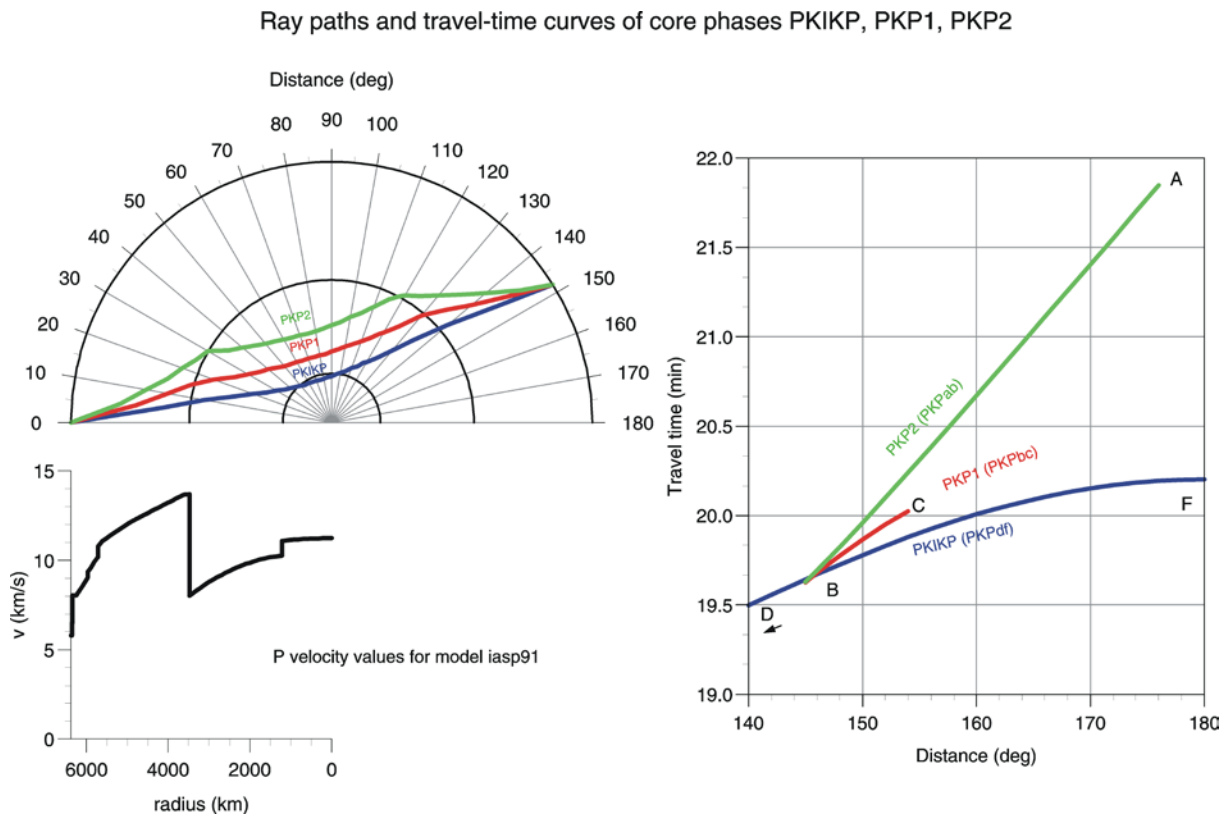


08.08.1994 Myanmar: H = 21:08:31.7, D(PLL) = ° ?, h = 122 km,  $m_b = 6.1$

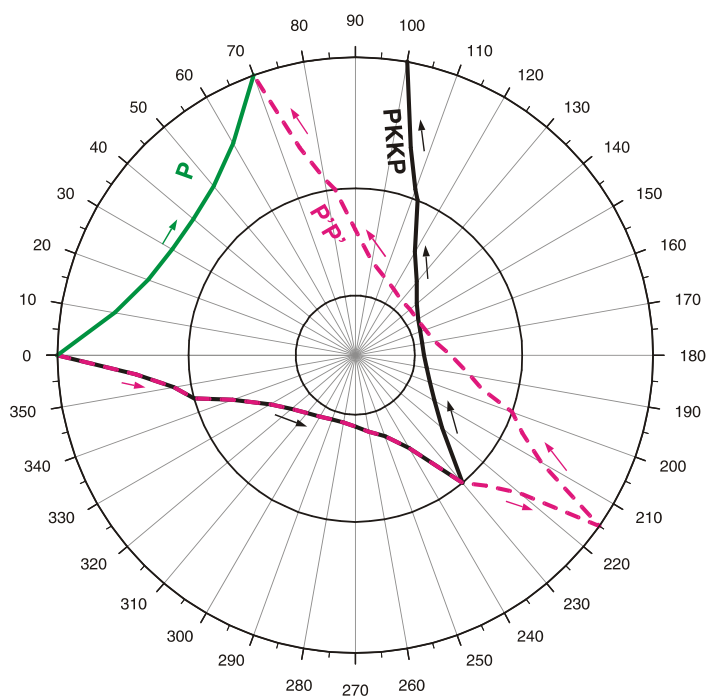


18.02.1996 North of Ascension Island: H = 23:49:27.8, D(PLL) = ° ?, h = 10 km,  $m_b = 6.3$

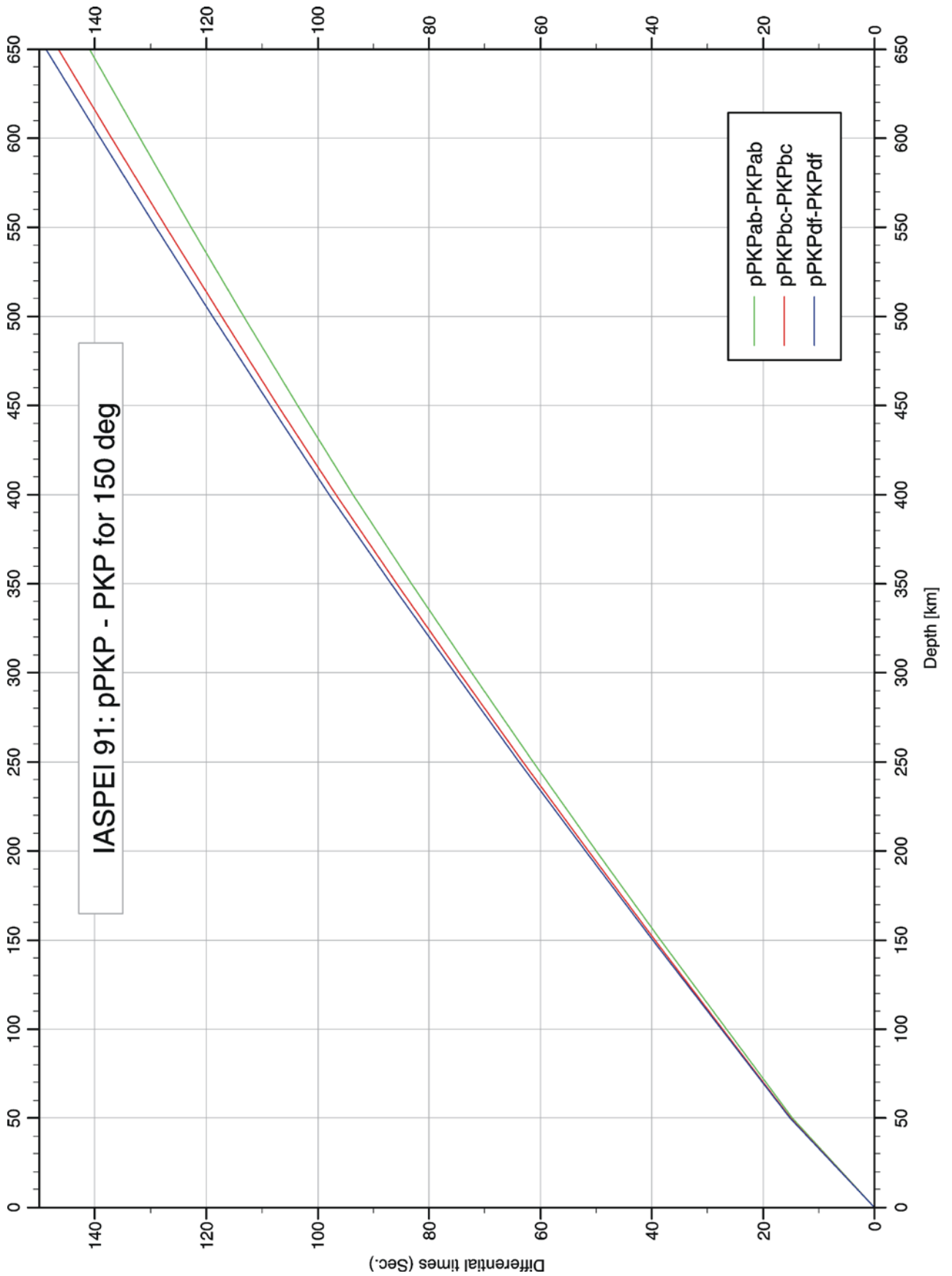
**Figure 4** Records of station CLL of later reflected longitudinal core phases (PKKP and P'P').



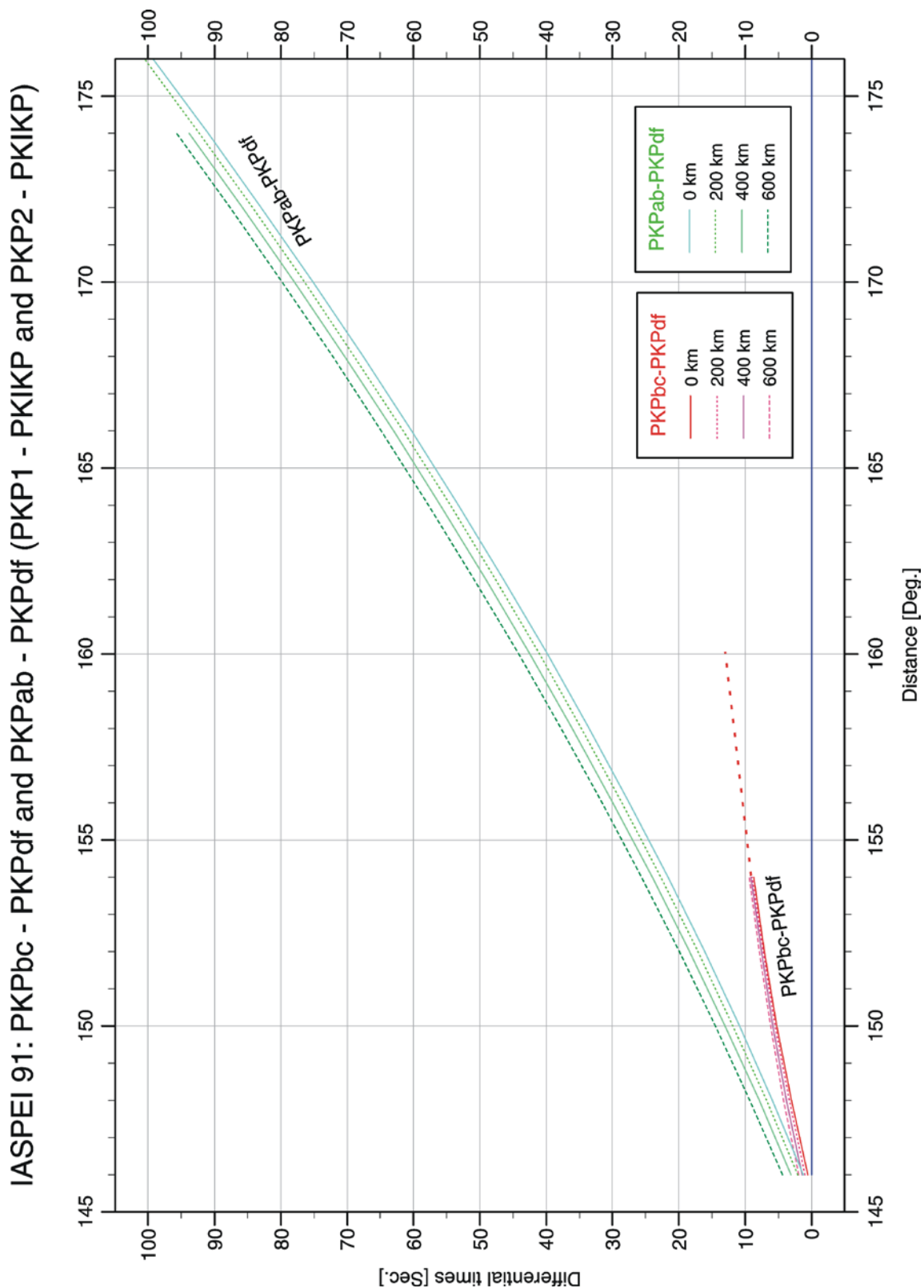
**Figure 5** Ray paths and travel-time curves of direct longitudinal core phases for  $D > 140^\circ$  according to the velocity model IASP91 (Kennett, 1991).



**Figure 6** Ray paths of the reflected core phases P'P' (or PKPPKP) and PKKP, respectively.



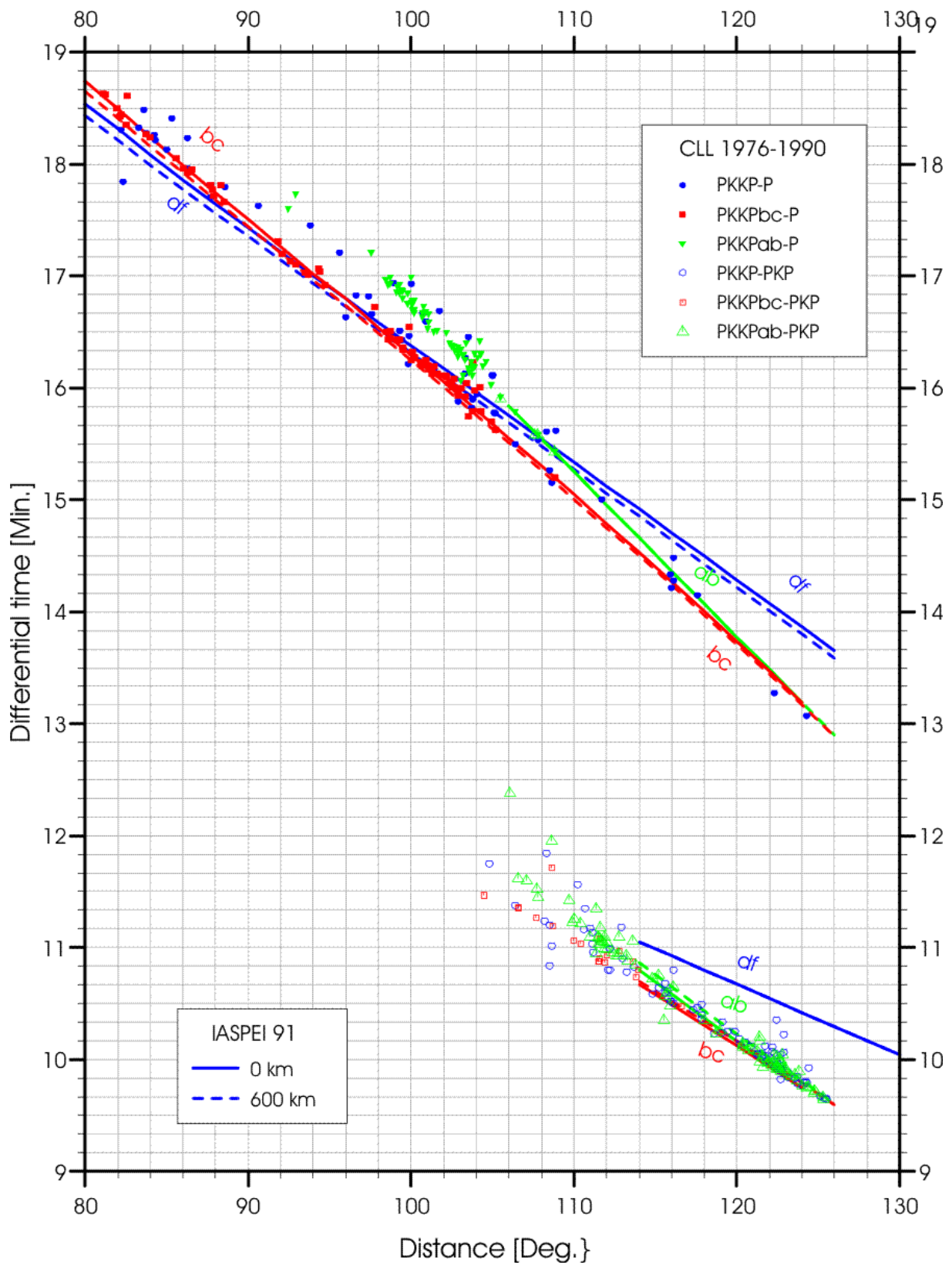
**Figure 7** Travel-time differences between the PKP arrivals for the branches ab, bc and df, respectively, and their related depth phases at an epicentral distance of  $D = 150^\circ$ .



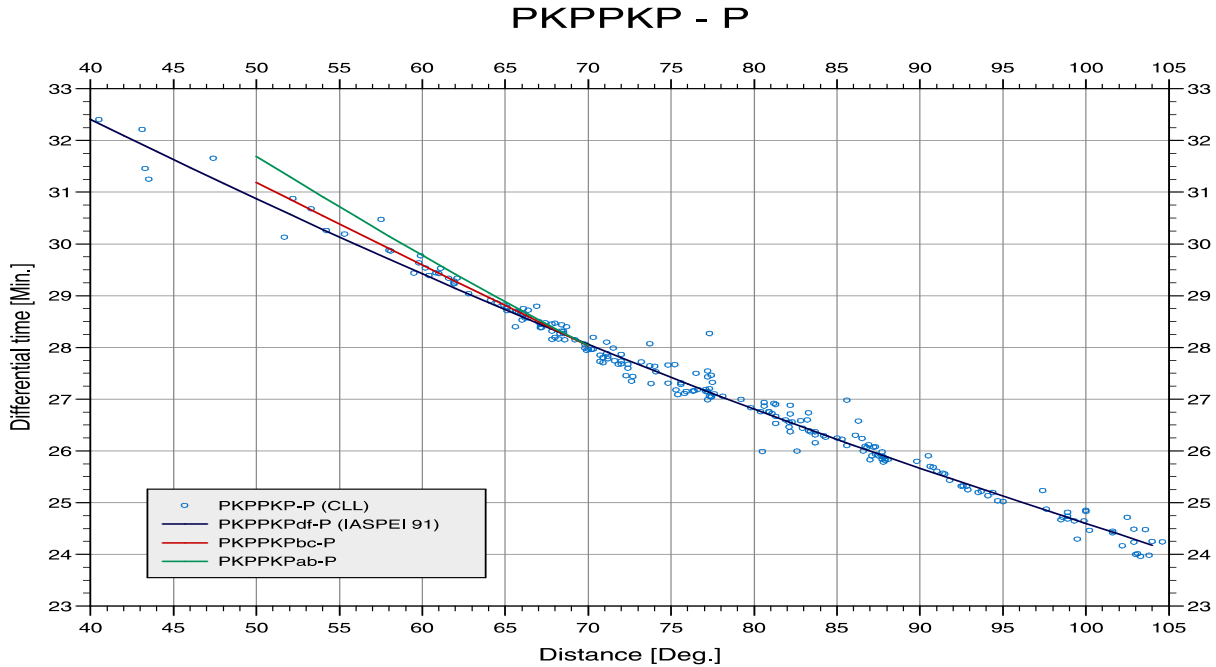
**Figure 8** Travel-time differences between the PKPdf first arrival and the later arrivals PKPbc and PKPab, respectively, for different source depths. The dotted continuation of the branch

PKPbc-PKPdif relates to the approximate differential arrival time of the weak phase PKPdif (PKP diffracted around the inner-core boundary).

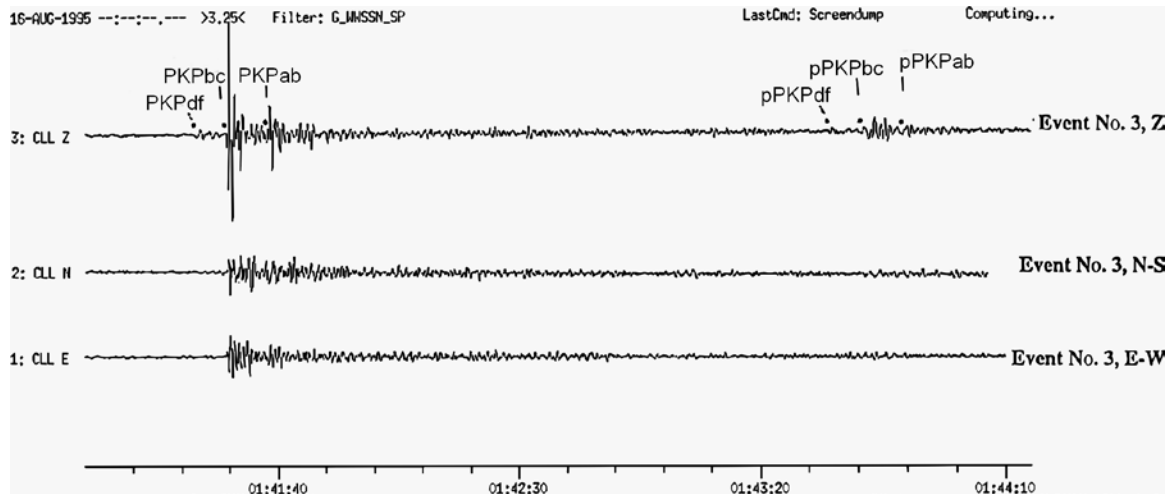
### PKKP - P and PKKP - PKP



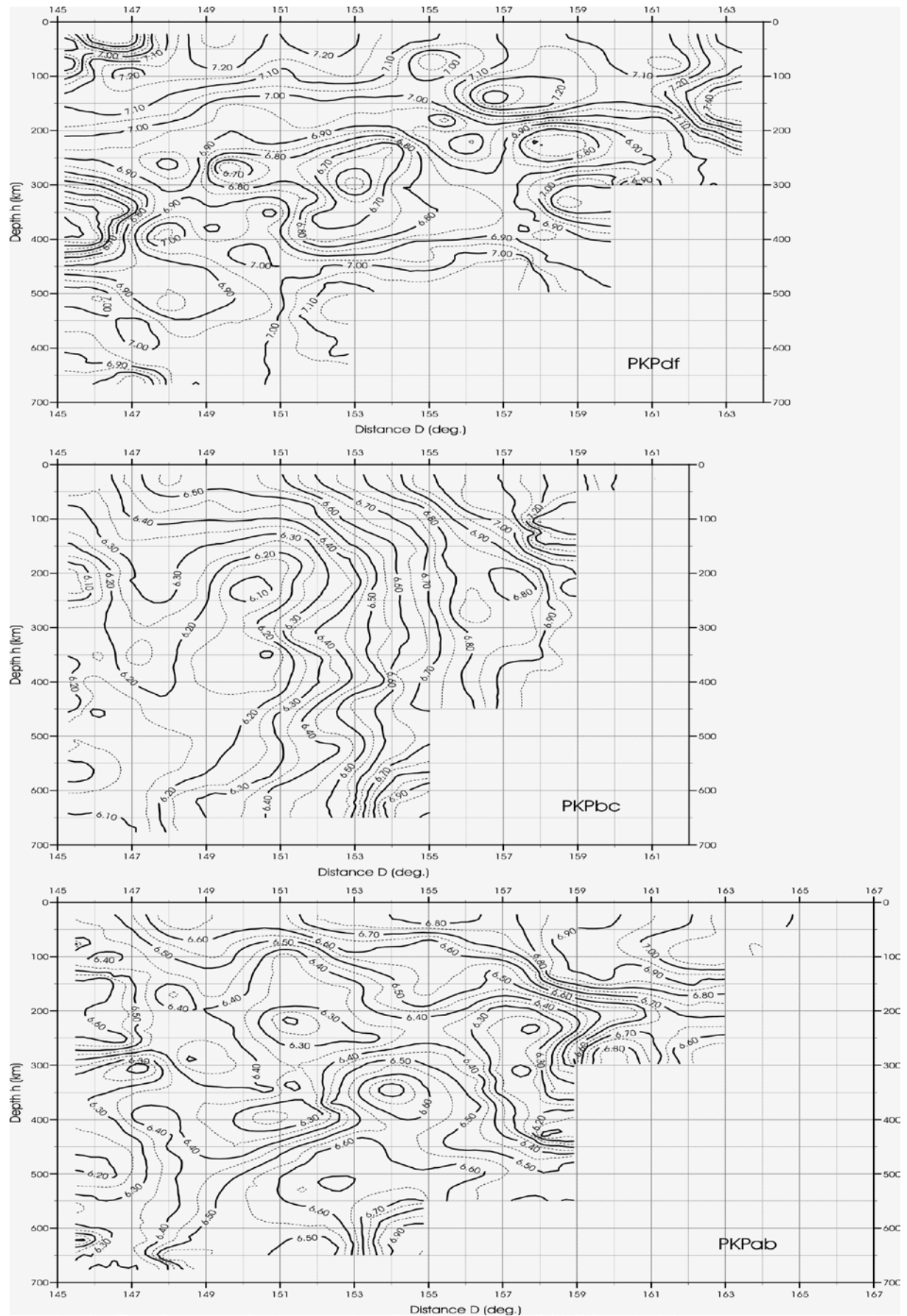
**Figure 9** Comprison between theoretical (IASP 91) and observed travel-time differences for PKKP-P and PKKP-PKP at station CLL as a function of epicentral distance and source depth.



**Figure 10** Comparison between theoretical (IASP91) and observed travel-time differences PKPPKP - P at station CLL depending on epicentral distance.



**Figure 11** Reproduction of the records of event No. 3 with onsets (dots) marked and names of identified phases given on the vertical component. Only the onset of the depth phase PKPbc can be picked without any doubt. However, when taking the time differences between the three direct phases from the beginning of the record into account one recognizes prior and after pPKPbc changes in the waveforms just at the same time differences as for the primary phases. In the records of earthquakes No. 1 and No. 2 in Figure 3 only the depth phases pPKPbc and pPKPab can be picked at about 119 s and 60 s after PKPbc and PKPab, respectively.





**Figure 12** Calibration functions according to S. Wendt for the determination of mb(PKP) for PKIKP = PKPdf, PKP1 = PKPbc and PKP2 = PKPab (see Bormann and Wendt, 1999).

## References

- Jeffreys, H., and Bullen, K. E. (1940, 1948, 1958, 1967, and 1970). Seismological Tables. British Association for the Advancement of Science, *Gray Milne Trust*, London, 50 pp.
- Kennett, B. L. N. (Ed.) (1991). IASPEI 1991 Seismological Tables. Research School of Earth Sciences, Australian National University, 167 pp.
- Kennett, B. L. N., and Engdahl, E. R. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, 105, 429-465.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, 122, 108-124.
- Bormann, P., and Wendt, S. (1999). Identification and analysis of longitudinal core phases: Requirements and guidelines. In: Bormann (1999), 346-366.

| Topic            | Seismic source location                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
|------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Authors</b>   | <p><b>Jens Havskov</b>, University of Bergen, Department of Earth Science, Allégaten 41, N-5007 Bergen, Norway, Fax: +47 55 583660, E-mail: <a href="mailto:Jens.Havskov@geo.uib.no">Jens.Havskov@geo.uib.no</a></p> <p><b>Peter Bormann</b>, (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> <p><b>Johannes Schweitzer</b>, NORSAR, P.O.Box 53, N-2027 Kjeller, Norway, Fax: +47 63818719, E-mail: <a href="mailto:johannes.schweitzer@norsar.no">johannes.schweitzer@norsar.no</a></p> |
| <b>Version 2</b> | September, 2011; DOI: 10.2312/GFZ. NMSOP-2_IS_11.1                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |

|                                                              | page |
|--------------------------------------------------------------|------|
| <b>1 Introduction</b>                                        | 1    |
| <b>2 Single station location</b>                             | 2    |
| <b>3 Multiple station location</b>                           | 5    |
| 3.1 Manual location                                          | 5    |
| 3.2 The Wadati diagram                                       | 6    |
| 3.3 Computer location                                        | 7    |
| 3.3.1 Grid search                                            | 8    |
| 3.3.2 Location by iterative methods                          | 10   |
| 3.3.3 Example of location in a homogeneous model             | 12   |
| 3.3.4 Advanced methods                                       | 13   |
| <b>4 Location errors</b>                                     | 14   |
| 4.1 Error quantification and statistics                      | 14   |
| 4.2 Example of error calculation                             | 16   |
| <b>5 Relative location methods</b>                           | 18   |
| 5.1 Master event technique                                   | 18   |
| 5.2 Joint hypocenter location                                | 18   |
| <b>6 Practical consideration in seismic source locations</b> | 19   |
| 6.1 Seismic phases                                           | 19   |
| 6.2 Initial location                                         | 21   |
| 6.3 Hypocentral depth                                        | 22   |
| 6.4 Outliers and weighting schemes                           | 24   |
| 6.5 Ellipticity of the Earth                                 | 25   |
| 6.6 Importance of the velocity model                         | 26   |
| <b>7 Internal and external (real) accuracy of locations</b>  | 28   |
| <b>Acknowledgements</b>                                      | 31   |
| <b>References</b>                                            | 31   |

## 1 Introduction

The exact location of any source, radiating seismic energy, is one of most important tasks in practical seismology and from time to time most seismologists have been involved in this task. The intention here is to describe the most common location methods without going into the mathematical details, which have been described in numerous textbooks and scientific papers but to give some practical advice on seismic source location. Chapter 9 gives more information about the usage of seismic arrays for seismic source location.

In the case of an earthquake, the source location is defined by its hypocenter ( $x_0, y_0, z_0$ ) and the origin time  $t_0$ . The hypocenter is commonly considered to be the physical location of the starting (or initiation) point of the rupture process, usually given in longitude ( $x_0$ ), latitude ( $y_0$ ), and depth below the surface ( $z_0$  [km]). For simplicity, the hypocenter will be labeled  $x_0, y_0, z_0$  with the understanding that it can be either measured in geographical or Cartesian coordinates, i.e., in [deg] or [km], respectively. The origin time is the start time of the earthquake rupture. The epicenter is the projection of the hypocenter on the Earth's surface ( $x_0, y_0$ ). When the earthquake is large, the physical dimension of the rupture zone can be several hundred square kilometers and the hypocenter can in principle be located anywhere on the rupture surface with consequences also for the position of the related epicenter on the Earth's surface.

The hypocenter and origin time are determined by arrival times, incidence direction (backazimuth) and/or horizontal velocity of seismic phases radiated by the first break of the earthquake rupture. This is true when using any P or S phases since their propagation velocity is always larger than the rupture velocity. Therefore, P- or S-wave energy emitted from the end of a long rupture will always arrive later than energy radiated from its beginning. Standard earthquake catalogs (such as from the International Seismological Center, ISC or the National Earthquake Information Center (NEIC) of the USGS) report origin times OT and locations based primarily on arrival times of high-frequency first P-type arrivals. These OT/location parameters can be quite different from the centroid time and centroid location obtained by moment-tensor inversion of long-period waves, which may be, for really great earthquakes, even more than a minute later than OT and more than hundred kilometers off the initial hypocenter. The reason is that these centroid solutions represent the average time and location for the entire seismic moment released by the earthquake, assuming that the latter can be considered as a point source in space with a prescribed symmetric triangular moment-rate function. Yet, even this assumption does not always hold, especially for great multiple-rupture events (e.g., Chapter 3, Fig. 3.7; Kikuchi and Fukao, 1987; Tsai et al., 2005).

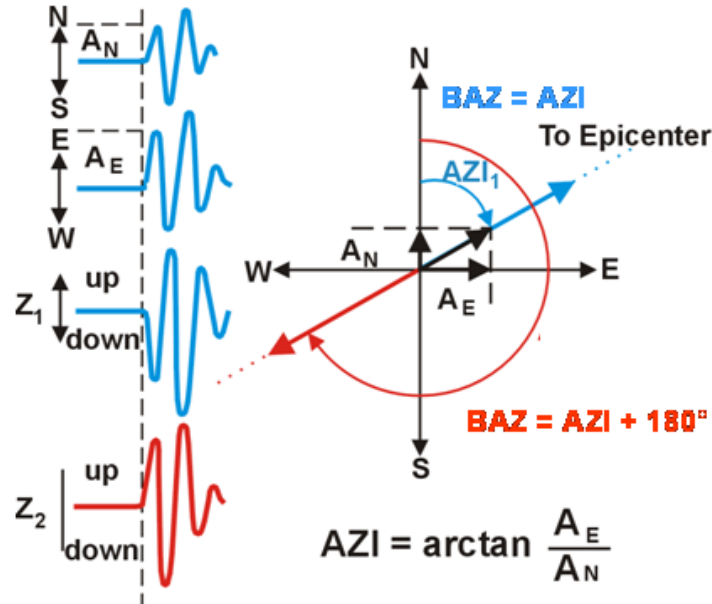
## 2 Single station location

In general, epicenters are determined using many seismic phase observations from different seismic stations. However, it is also possible to locate a seismic source using a single 3-component station. Since the vector of P-wave motion is polarized in the vertical plane of propagation (see Chapter 2) it can be decomposed into a vertical and a radial component of motion. This allows calculating the backazimuth **BAZ** to the epicenter (see Figure 1). The backazimuth is defined as the direction angle from the station towards the epicenter, measured clockwise from North. The radial component of P can be reconstructed from the records of the 2 horizontal seismometer components **N**(orth)-**S**(outh) and **E**(ast)-**W**(est) and the azimuth **AZI** of particle motion can be calculated via the amplitude ratio  $A_E/A_N$  of the first P-phase arrival:

$$AZI = \arctan A_E/A_N. \quad (1)$$

However, eq. (1) leaves an ambiguity of  $180^\circ$  since the first motion (FM) polarity in the vertical component can be up or down. To get the correct **BAZ** this has to be accounted for. Only if the first motion on the vertical component of the P phase is downward, the related horizontal particle motion shows towards the epicenter and then, when **AZI** is measured in the first quadrant, holds **BAZ** = **AZI**. The opposite is the case when the first motion in Z is

upward, due to an outward directed motion at the source which corresponds by definition to a compressional first motion arriving at the station. In this case the radial component of the P phase motion is directed away from the hypocenter and thus  $\text{BAZ} = \text{AZI} + 180^\circ$  (see Figure 1). For details see Table 1 in exercise EX 11.2.



**Figure 1** Example of P-wave first motions on 3-component records (left) from which the AZI (and thereby the BAZ) and the incidence angle  $i$  can be derived according to Eqs. (1) and (2).

The amplitude  $A_Z$  of the Z component can, together with the amplitude  $A_R = (A_E^2 + A_N^2)^{1/2}$  on the radial component, also be used to calculate the *apparent angle of incidence*  $i_{\text{app}} = \arctan(A_R / A_Z)$  of a P wave. However, according already to Wiechert (1907) the true incidence angle  $i_{\text{true}}$  of a P wave is a function of the mean  $v_P$ -to- $v_S$  ratio in the crust below the station

$$i_{\text{true}} = \arcsin \left[ \frac{v_P}{v_S} \times \sin(0.5 \times i_{\text{app}}) \right], \quad (2)$$

with the difference accounting for the amplitude distortion due to the reflection of P and SV-converted wave energy at the free surface. Knowing the incidence angle  $i_0$  and the local seismic velocity  $v_0$  below the observing station, we can calculate the apparent horizontal propagation velocity  $v_{\text{app}}$  of this seismic phase with

$$v_{\text{app}} = v_0 / \sin i_0. \quad (3)$$

One should note, however, that the here described measured  $i_{\text{app}}$  and thus  $v_{\text{app}}$  values at a given station depend on the dominant frequency of the seismic record (see Chapter 2, Fig. 2.6). Therefore, these values are representative for the uppermost layers of the Earth corresponding to the dominant wavelength of the observed phase and not for the incidence angle and apparent horizontal velocity directly underneath the Earth's surface.

In the case of slowly emerging first P swings (e.g., Figure 6 and 8 in IS 11.4) or when analyzing narrow-band high frequency data (e.g., Chapter 4, Figs. 4.9 and 4.10, and Chapter 16, Fig. 16.25) it might not be possible to read correctly the amplitude and polarity of the first break (see Chapter 4, Figs. 4.9 and 4.10). However, the amplitude ratio between the components should remain constant, not only for the first swing of the P phase but also for the following oscillations of the same phase, provided that they are not yet distorted by superposition with laterally refracted (scattered) wave energy, as often the case in high-frequency records. If this condition is fulfilled, we can, with digital data, use the predicted coherence method (Roberts et al., 1989) to automatically calculate backazimuth as well as the angle of incidence. This is much more reliable and faster than using manually read first motion amplitudes for calculating the backazimuth from 3-component records of single stations and has become a routine practice (e.g., Saari, 1991).

In the case of seismic arrays, apparent velocity and backazimuth can directly be measured by observing the propagation of the seismic wavefront with array methods, independently from the local seismic velocities below the station (see Chapter 9). As we show later, backazimuth observations are useful in better constraining epicenter locations and in associating observations to a seismic event. Knowing the incidence angle  $i_o$  at the station and implicitly the ray parameter  $p = \sin i_o / v_o$  of an onset at the station helps to identify the seismic phase and to calculate the epicentral distance via the, from Earth models theoretically known, distance-dependent slowness (or ray parameter) of different seismic phases.

With a single station we have now the direction to the seismic source. The distance can also be obtained from the difference in arrival time of two phases, usually P and S. If we assume a constant velocity, and origin time  $t_0$ , the P- and S-arrival times can then be written as

$$t_p = t_0 + D/v_p \quad t_s = t_0 + D/v_s \quad (4)$$

where  $t_p$  and  $t_s$  are the P- and S-arrival times respectively,  $v_p$  and  $v_s$  are the P and S velocities respectively and  $D$  is the epicentral distance for surface sources; or the hypocentral distance  $d$  for deeper sources. By eliminating  $t_0$  from Equation (4), the distance can be calculated as

$$D = (t_s - t_p) \frac{v_p \cdot v_s}{v_p - v_s} \quad (5)$$

with  $D$  in km and  $t_s - t_p$  in seconds. But Equation (5) is applicable only for the travel-time difference between  $S_g$  and  $P_g$ , i.e., the direct crustal phases of S and P, respectively, provided that the ratio  $v_p/v_s$  does not change within the crust.  $P_g$  and  $S_g$  are first onsets of the P- and S-wave groups of local events only for distances up to about 100 – 250 km, depending on crustal thickness and source depth within the crust. Beyond these distances the phases  $P_n$  and  $S_n$ , being either head waves critically refracted at the Mohorovičić discontinuity or waves diving as body waves in the uppermost part of the upper mantle, become the first onsets (see Fig. 2.32 in Chapter 2 and Fig. 11.40 in Chapter 11). The “cross-over” distance  $x_{co}$  between  $P_n$  and  $P_g$  (or  $P_b$ ) can be approximately calculated for a (near) surface focus from the relationship

$$x_{co} = 2 z_m \{ (v_m - \bar{v}_p) (v_m + \bar{v}_p) \}^{-1/2}, \quad (6)$$

with  $\bar{v}_p$  – average crustal P velocity,  $v_m$  – sub-Moho P velocity, and  $z_m$  – crustal thickness. Inserting the rough average values of  $\bar{v}_p = 6$  km/s and  $v_m = 8$  km/s we get, as a “rule of thumb”,  $x_{co} \approx 5 z_m$ . At smaller distances we can be rather sure that the observed first arrival is Pg. Note, however, that this “rule of thumb” is valid for surface focus only. As demonstrated with Fig. 2.40 in Chapter 2, the crossover distance is only about half as large for near Moho earthquakes and also the dip of the Moho and the direction of observation (up- or downdip) does play a role.

Examples for calculating the epicentral distance  $D$  and the origin time  $OT$  of near seismic events by means of a set of local travel-time curves for Pn, Pg, Sn and Sg (or Lg) are given in exercise EX 11.1. The lower crustal phases Pb and Sb are here less important since they may form only over a very short epicentral distance range the first arrivals of the P and S onsets. In the absence of local travel-time curves for the area under consideration one can use Equation (5) for deriving a “**rule of thumb**” for approximate distance determinations from travel-time differences Sg-Pg. For an ideal Poisson solid  $v_s = v_p/\sqrt{3}$ . This is a good approximation for the average conditions in the crust. With this follows from Equation (5):  $D = (t_{Sg} - t_{Pg}) \times 8.0$  for “normal, medium age” crustal conditions with average  $\bar{v}_p = 5.9$  km/s, or  $D = (t_{Sg} - t_{Pg}) \times 9.0$  for old Precambrian continental shields with rather large average  $\bar{v}_p = 6.6$  km/s. However, if known, the locally correct  $v_p/v_s$  ratio should be used to improve this “**rule of thumb**”. If the distance is calculated from the travel-time difference between Sn and Pn, another good rule of thumb is  $D = (t_{Sn} - t_{Pn}) \times 10$ . It may be applicable up to about 1000 km epicentral distance.

For epicentral distances between about  $20^\circ < \Delta < 100^\circ$  the relationship  $\Delta^\circ = \{(t_s - t_p)_{\min} - 2\} \times 10$  still yields reasonably good epicentral distance estimates with errors  $< 3^\circ$ , however, beyond  $D = 10^\circ$  the use of readily available global travel-time tables such as IASP91 (Kennett and Engdahl, 1991; Kennett, 1991), SP6 (Morelli and Dziewonski, 1993), or AK135 (Kennett et al., 1995; now standard at the NEIC and the ISC) is strongly recommended for more accurate theoretical travel-time and epicentral distance calculations.

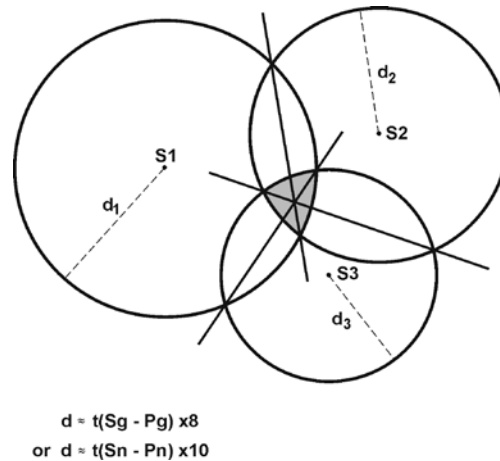
With both backazimuth and epicentral distance, the epicenter of a seismic event can be obtained by measuring off the distance along the backazimuth of approach. Finally, knowing the epicentral distance, we can calculate the P-wave travel time and thereby get the origin time using the P-arrival time (see EX 11.2 for location of teleseismic events by means of 3-component records).

It is important to mention at this point that the described single station location procedures cannot resolve the depth of a seismic event, unless the station analyst is able to surely identify depth phases in the record and determine the source depth from their time difference to the related primary phases. Otherwise, single station location is a pure epicenter and not hypocenter determination method.

### 3 Multiple station location

#### 3.1 Manual location

When at least 3 stations are available, a simple manual location can be made from drawing circles (the circle method) with the center at the station locations and the radii equal to the hypocentral distances calculated from the S-P times (see Figure 2).



**Figure 2** Location by the “circle and chord” method. The stations are located in S1, S2 and S3. The epicenter is found within the shaded area where the circles overlap. The best estimate is the crossing of the chords, which connect the crossing points of circle pairs.

These circles will rarely cross in one point. This may be due to errors in the onset-time readings, in the assumed travel-time or related velocity model and/or that we have wrongly assumed a surface focus. In fact,  $t_s - t_p$  is the travel-time difference for the hypocentral distance  $d$ , which is for seismic sources with  $z > 0$  km generally larger than the epicentral distance  $\Delta$  (or  $D$ ). Therefore, the circles drawn around the stations with radius  $d$  will normally not be crossing at a single point at the epicenter but rather “overshooting”. The amount of overshoot depends on source depth but also on not accounted lower than average velocity inhomogeneities. One should therefore fix the epicenter either in the “center of gravity” of the overlapping area (shaded area in Figure 2) or draw “chords”, i.e., straight lines passing through the crossing points between two neighboring circles. These chord lines intersect in the epicenter (see Figure 1 in EX 11.1). In Section 7 of this exercise and in the animation linked with it, it is illustrated how the source depth can then be determined by iteratively minimizing the overshoot. Still other methods exist (e.g., Båth, 1979) to deal with this depth problem (e.g., the hyperbola method which uses P-wave first arrivals only and assumes a constant P-wave velocity). A detailed description of the method can be found in Pujol (2004).

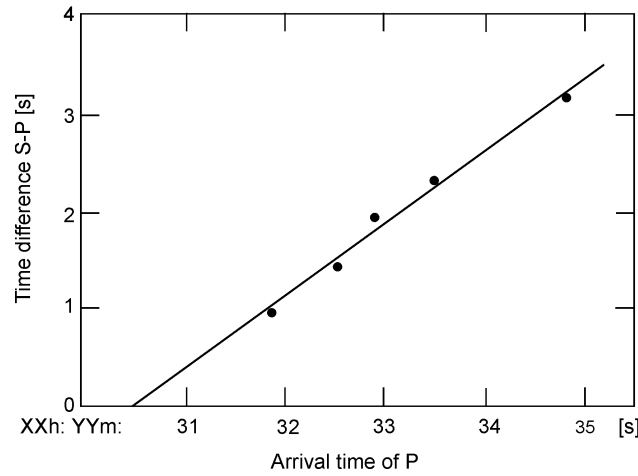
During the last years Earthquake Early Warning Systems (EEWS) became more in focus and for such monitoring systems quick and reliable earthquake location methods are needed. Therefore, all graphical location algorithms became again more interesting. One approach to solve the problem is using the order of first arrival times to locate the event (Nicholson et al., 2004; Rosenberger, 2009) or applying derivatives of the hyperbola method (e.g., Font et al., 2004; Horiuchi et al., 2005; Satriano et al., 2008).

### 3.2 The Wadati diagram

With data from several stations available from a local or regional seismic source, the origin time can be determined by a very simple technique called Wadati diagram (Wadati, 1933). Using Equation (5) and eliminating  $\Delta$ , the S-P travel-time difference can be calculated as

$$t_s - t_p = (v_p/v_s - 1) \times (t_p - t_0) \quad (7)$$

The observed S-P times are plotted against the absolute P time. Since  $t_s - t_p$  goes to zero at the hypocenter, a straight line fit on the Wadati diagram (Figure 3) gives the origin time at the intercept with the P-arrival axis and from the slope  $(v_p/v_s - 1)$  of the curve, we get the ratio  $v_p/v_s$ . Note that it is thus possible to get a determination of both the origin time and a mean  $v_p/v_s$  ratio without any prior knowledge of the crustal structure, the only assumption is that  $v_p/v_s$  is constant and that the P and S phases are of the same type like Pg and Sg, Pb and Sb or Pn and Sn. Such an independent determination of these parameters can be very useful when using other location methods.



**Figure 3** An arbitrary example of a Wadati diagram. The intercept of the best fitting line through the data with the x-axis gives the origin time OT. In the given case, the slope of the line is 0.72 so the  $v_p/v_s$  ratio is 1.72. The misfit of the data with this straight line indicates deviations from a constant  $v_p/v_s$  ratio and/or data reading errors.

The Wadati diagram is also very useful in making independent checks of the observed arrival times. Any points not fitting reasonably well with the linear relationship might be badly identified, either by not being of the same phase type or by misreading.

### 3.3 Computer location

Manual location methods provide insight into location problems; however, in practice we use computer methods. In the following, the most common ways of calculating hypocenter and origin time by computer will be discussed.

The calculated arrival time  $t_i^c$  at station  $i$  can be written as

$$t_i^c = T(x_i, y_i, z_i, x_0, y_0, z_0) + t_0 \quad (8)$$



where  $T$  is the travel time as a function of the location of the station  $(x_i, y_i, z_i)$  and the hypocenter. This equation has 4 unknowns, so in principle 4 arrival-time observations from at least 3 stations are needed in order to determine the hypocenter and origin time. If we have  $n$  observations, there will be  $n$  equations of the above type and the system is over determined and has to be solved in such a way that the misfit or residual  $r_i$  at each station is minimized.  $r_i$  is defined as the difference between the observed and calculated travel times which is the same as the difference between the observed and calculated arrival times

$$r_i = t_i^o - t_i^c \quad (9)$$

where  $t_i^o$  is the observed arrival time. In principle, the problem seems quite simple. However, since the travel-time function  $T$  is a nonlinear function of the model parameters, it is not possible to solve Equation (8) with any analytical methods. So even though  $T$  can be quite simply calculated, particularly when using a 1-D Earth model or pre-calculated travel-time tables, the non-linearity of  $T$  greatly complicates the task of inverting for the best hypocentral parameters. The non-linearity is evident even in a simple 2-D epicenter determination where the travel time  $t_i$  from the point  $(x, y)$  to a station  $(x_i, y_i)$  can be calculated as

$$t_i = \frac{\sqrt{(x - x_i)^2 + (y - y_i)^2}}{v}, \quad (10)$$

where  $v$  is the velocity. It is obvious that  $t_i$  does not scale linearly with either  $x$  or  $y$  so it is not possible to use any set of linear equations to solve the problem and standard linear methods cannot be used. This means that given a set of arrival times, there is no simple way of finding the best solution. In the following, some of the methods of solving this problem will be discussed.

### 3.3.1 Grid search

Since it is so simple to calculate the travel times of all seismic phases to any point in the model, given enough computer power, a very simple method is to perform a grid search over all possible locations and origin times and compute the arrival time at each station (e.g., Sambridge and Kennett, 1986 and 2001; Lomax, 2005). The hypocentral location and origin time would then be the point with the best agreement between the observed and calculated times. This means that some measure of best agreement is needed, particularly if many observations are used. The most common approach is to use the least squares solution, which is to find the minimum of the sum of the squared residuals  $e$  from the  $n$  observations:

$$e = \sum_{i=1}^n (r_i)^2 \quad (11)$$

The root mean squared residual RMS, is defined as  $\sqrt{e/n}$ . RMS is given in almost all location programs and commonly used as a guide to location precision. If the residuals are of similar size, the RMS gives the approximate average residual. As will be seen later, RMS only gives an indication of the fit of the data, and a low RMS does not automatically mean an accurate hypocenter determination. Generally, the precision of the computational solution, which is

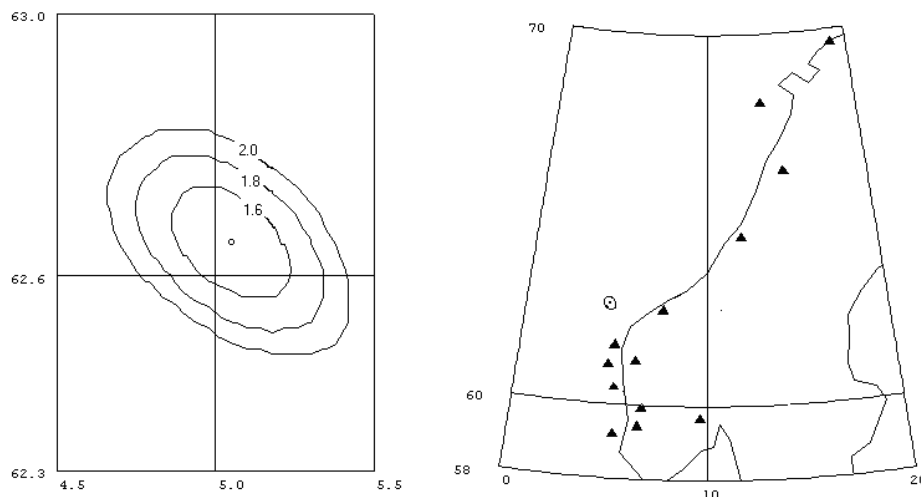
based on various model assumptions, should not be mistaken as real accuracy of the location and origin time. This point will be discussed later under Section 7.

The average squared residual  $e/n$  is called the variance of the data. Formally,  $n$  should here be the number of degrees of freedom  $ndf$ , which is the number of observations minus the number of parameters in fit (here 4). Since  $n$  usually is large, it can be considered equal to the number of degrees of freedom. This also means that  $\mathbf{RMS}^2$  is approximately the same as the variance. The least squares approach is the most common measure of misfit since it leads to simple forms of the equations in the minimization problems (see later). It also works quite well if the residuals are caused by uncorrelated Gaussian noise. However in real problems this is often not the case. A particularly nasty problem is the existence of outliers, i.e., individual large residuals. A residual of 4 will contribute 16 times more to the misfit  $e$ , than a residual of 1. Using the sum of the absolute residuals as a norm for the misfit can partly solve this problem:

$$e1 = \sum_{i=1}^n |r_i|. \quad (12)$$

This is called the L1 norm and is considered more robust when there are large outliers in the data. It is not much used in standard location programs since the absolute sign creates complications in the equations. This is of course not the case for grid search. Therefore, most location programs will have some scheme for weighting out or truncating large residuals (see later), which can partly solve the problem.

Once the misfits (e.g., RMS) have been calculated at all grid points, one could assign the point with the lowest RMS as the ‘solution’. For well-behaved data, this would obviously be the case, but with real data, there might be several points, even far apart, with similar RMS and the next step is therefore to estimate the probable uncertainties of the solution. The simplest way to get an indication of the uncertainty, is to contour the RMS as a function of  $x$  and  $y$  (2-D case) in the vicinity of the point with the lowest RMS (see Figure 4).



**Figure 4** Left: RMS contours (in seconds) from a grid search location of an earthquake off western Norway (left). The grid size is 2 km. The circle in the middle indicates the point with the lowest RMS (1.4 s). Right: The location of the earthquake and the stations used. Note the elongated geometry of the station distribution. Its effect on the error distribution will be

discussed in Section 4.1 below. The RMS ellipse from the figure on the left is shown as a small ellipse in the figure at right. Latitudes are degrees North and longitudes degrees East.

Clearly, if RMS is growing rapidly when moving away from the minimum, a better solution has been obtained than if RMS grows slowly. If RMS is contoured in the whole search area, other minima of similar size might be found indicating not only large errors but also a serious ambiguity in the solution. Also note in Figure 4 that networks with irregular aperture have reduced distance control in the direction of their smallest aperture but good azimuth control in the direction of their largest aperture.

An important point in all grid-search routines is the method of how to search through the possible model space. In particular for events observed at teleseismic distances the model space can be very large. Sambridge and Kennett (2001) published a fast neighborhood algorithm to use for global grid search.

### 3.3.2 Location by iterative methods

Despite increasing computer power, seismic source locations are done mainly by other methods than grid search. These methods are based on linearizing the inversion problem. The first step is to make a guess of hypocenter and origin time ( $x_0, y_0, z_0, t_0$ ). In its simplest form, e.g., in the case of events near or within a station network, this can be done by using a location near the station with the earliest arrival time and using that arrival time as  $t_0$ . Other methods also exist (see below). In order to linearize the problem, it is now assumed that the true hypocenter is close enough to the guessed value so that travel-time residuals at the trial hypocenter are a linear function of the correction we have to make in hypocentral distance.

The calculated arrival times at station  $i$ ,  $t_i^c$  from the trial location are, as given in Equation (8),  $t_i^c = T(x_0, y_0, z_0, x_i, y_i, z_i) + t_0$  and the travel-time residuals  $r_i$  are  $r_i = t_i^o - t_i^c$ . We now assume that these residuals are due to the error in the trial solution and the corrections needed to make them zero are  $\Delta x$ ,  $\Delta y$ ,  $\Delta z$ , and  $\Delta t$ . If the corrections are small, we can calculate the corresponding corrections in travel times by approximating the travel-time function by a Taylor series and using only the first term. The residual can now be written:

$$r_i = (\partial T / \partial x_i) * \Delta x + (\partial T / \partial y_i) * \Delta y + (\partial T / \partial z_i) * \Delta z + \Delta t \quad (13)$$

In matrix form we can write this as

$$\mathbf{r} = \mathbf{G} * \mathbf{X}, \quad (14)$$

where  $\mathbf{r}$  is the residual vector,  $\mathbf{G}$  the matrix of partial derivatives (with 1.0 in the last column corresponding to the source time correction term) and  $\mathbf{X}$  is the unknown correction vector in location and origin time.

This is a set of linear equations with 4 unknowns (corrections to hypocenter and origin time), and there is one equation for each observed phase time. Normally there would be many more equations than unknowns (e.g., 4 stations with 3 phases each would give 12 equations). The best solution to Equation (13) or Equation (14) is usually obtained with standard least squares techniques. The original trial solution is then corrected with the results of Equation (13) or

Equation (14) and this new solution can then be used as trial solution for a next iteration. This iteration process can be continued until a predefined breakpoint is reached. Breakpoint conditions can be either a minimum residuum  $r$ , or a last iteration giving smaller hypocentral parameter changes than a predefined limit, or just the total number of iterations. This inversion method was first invented and applied by Geiger (1910) and is known as the ‘Geiger method’. The iterative process usually converges rapidly unless the data are badly configured or the initial guess is very far away from the mathematically best solution (see later). However, it also happens that the solution converges to a local minimum and this would be hard to detect in the output unless the residuals are very large. A test with a grid search program may tell if the minimum is local, or tests could be made with several initial locations or different velocity models.

So far we have only dealt with observations in terms of arrival times. Many 3-component stations and arrays now routinely report the backazimuth  $\phi$  of an arrival. It is then possible to locate events with only one station and P and S times (see Figure 1). However, the depth must be fixed. If one or several backazimuth observations are available, they can be used together with the arrival time observations in the inversion algorithm. The additional equations for the backazimuth residuals are

$$r_i^\phi = (\partial\phi/\partial x_i) * \Delta x + (\partial\phi/\partial y_i) * \Delta y \quad (15)$$

Equations of this type are then added to the Equations (13) or (14). The  $\Delta x$  and  $\Delta y$  in Equation (15) are the same as for Equation (13), however the residuals are now in degrees. At this point it is worth to mention that backazimuth observations constrain the epicenter of an event but are independent of source depth and source time.

In order to make an overall RMS, the degrees must be ‘converted to seconds’ in terms of scaling. For example, in the location program HYPOCENTER (Lienert et al., 1988; Lienert, 1991; Lienert and Havskov, 1995), a 10 deg backazimuth residual was optionally made equivalent to 1 s travel time residual. Using e.g., 20 deg as equivalent to 1 s would lower the weight of the backazimuth observations. Schweitzer (2001) used in the location program HYPOSAT a different approach. In this program the measured (or assumed) observation errors of the input parameters are used to individually weight all different lines of the equation system (13) or (14) before inverting it. Thereby, more uncertain observations will contribute much less to the solution than well-constrained ones and all equations become non-dimensional.

Arrays (see Chapter 9) or single stations (see Equation (3)) cannot only measure the backazimuth of a seismic phase but also its ray parameter (or apparent velocity), which is a function of epicentral distance and source depth. Consequently, the equation systems (13) or (14) to be solved for locating an event, can also be extended by utilizing such observed ray parameters  $p$  (or apparent velocities) as defining data. In this case we can write

$$r_i^p = (\partial p/\partial x_i) * \Delta x + (\partial p/\partial y_i) * \Delta y + (\partial p/\partial z_i) * \Delta z \quad (16)$$

Also equation (16) is independent of the source time but depends on the source depth. The partial derivatives are often very small but in some cases, in particular if an event is observed with only one seismic array, observed ray parameters give additional constraint for the event location.

Equations (13), (14) and (16) are written without discussing whether the applied velocity model is defined for a layered, flat or a spherical Earth model because they are exactly the same. Using a flat-Earth transformation (e.g., Müller, 1977) any radially symmetric Earth model can be transformed into a flat model. Travel times and partial derivatives are often calculated by interpolating in tables and in principle it is possible to use any Earth model including 2-D and 3-D models to calculate theoretical travel times. In practice, 1-D models are mostly used, since 2-D and 3-D models are normally not well enough known and the travel-time computations are much more time consuming. For local seismology, it is a common practice to specify a 1-D crustal model and calculate arrival times for each ray while for global models, an interpolation in travel-time tables such as IASP91 is the most common. As Kennett and Engdahl (1991) pointed out, the preferred and much more precise method for obtaining travel times from the IASP91 model or any other 1-D global Earth model (see DS 2.1) is to apply the tau-p algorithm developed by Buland and Chapman (1983). To calculate your own travel-time tables for local or global 1-D Earth models, the computer program LAUFZE can be downloaded for free either directly from the program list in the NMSOP-2 cover page folder *Download Programs and Files* or from NORSAR's anonymous ftp-server <ftp://ftp.norsar.no/pub/outgoing/johannes/lauf/>, a description of the program is annexed in PD 11.2. It allows calculating travel times for many different seismic phases and an arbitrary horizontally layered model with any combination of layers with constant velocities, gradients, or first-order discontinuities.

### 3.3.3 Example of location in a homogeneous model

The simplest location case is a source in a homogeneous medium. The arrival times can be calculated as

$$T_i = \frac{\sqrt{(x - x_i)^2 + (y - y_i)^2 + (z - z_i)^2}}{v} + t_0, \quad (17)$$

where  $v$  is the velocity. The partial derivatives can be estimated from Equation (17) and e.g., for  $x$ , the derivative is

$$\frac{\partial T_i}{\partial x} = \frac{(x - x_i)}{v} * \frac{1}{\sqrt{(x - x_i)^2 + (y - y_i)^2 + z^2}}. \quad (18)$$

Similar expressions can be derived for  $y$  and  $z$ .

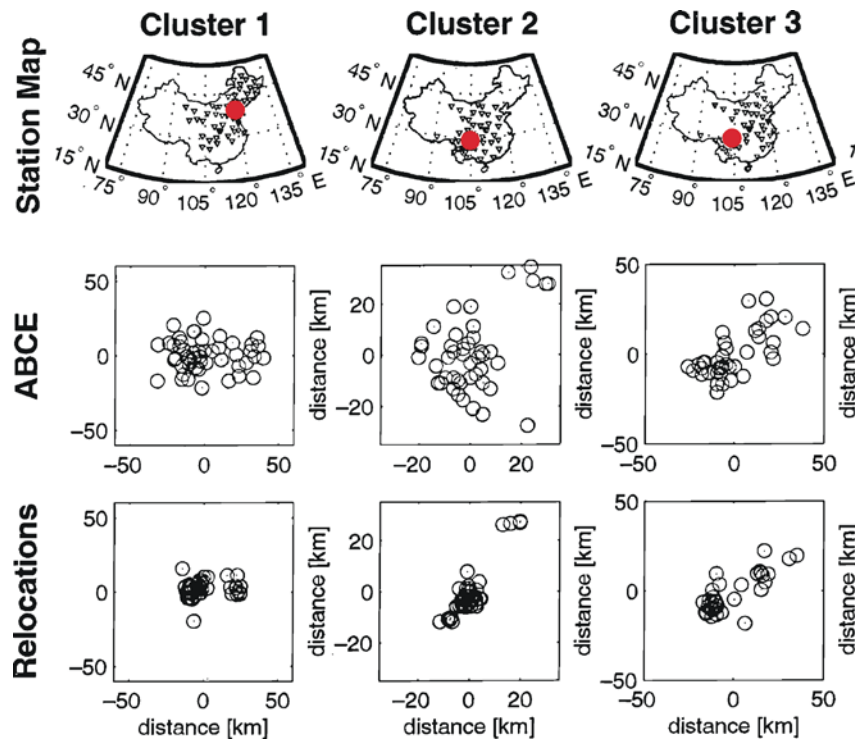
**Table 1** Inversion of error free data. Hypocenter is the correct location, Start is the initial location, and the location is shown for the two following iterations. The rows give the hypocentral parameters ( $x$ ,  $y$ ,  $z$  and  $t_0$ ), the misfit  $e$  according to Equation (11) and the RMS value of the residuals.

|                       | Hypocenter | Start | 1. Iteration | 2. Iteration |
|-----------------------|------------|-------|--------------|--------------|
| x [km]                | 0.0        | 3.0   | -0.5         | 0.0          |
| y [km]                | 0.0        | 4.0   | -0.6         | 0.0          |
| z [km]                | 10.0       | 20.0  | 10.1         | 10.0         |
| $T_0$ [s]             | 0.0        | 2.0   | 0.2          | 0.0          |
| $e$ [s <sup>2</sup> ] |            | 94.2  | 0.6          | 0.0          |
| RMS [s]               |            | 3.1   | 0.25         | 0.0          |

Table 1 gives an example of locating an earthquake with 10 stations in a model with constant velocity (from Stein, 1991). The stations are between 11 and 50 km apart from the hypocenter. The earthquake has an origin time of 0 s at the point (0, 0, 10) km. The initial location is at (3, 4, 20) km at 2 s. The exact travel times were calculated using a velocity of 5 km/s and the iterations were done as indicated above. At the initial guess, the sum of the squared residuals was 92.4 s<sup>2</sup>, after the first iteration it was reduced to 0.6 s<sup>2</sup> and already at the second iteration; the ‘correct’ solution was obtained. This is hardly surprising, since the data had no errors. We shall later see how this works in the presence of errors.

### 3.3.4 Advanced methods

The problem of locating seismic events has during the last decade experienced a lot of attention and new procedures have been developed such as the double-difference location algorithm (Waldhauser and Ellsworth, 2000), a novel global differential evolution algorithm (Ružek and Kvasnička (2001), a probabilistic approach to location in 3-D and layered models by Lomax et al. (2000) as well as advanced grid search procedures to be applied in highly heterogeneous media (Lomax et al., 2001). Recent advances in travel-time calculations for three-dimensional structures complement these methods (e.g., Thurber and Kissling, 2000). Figure 5 shows how much the *precision* (or relative inter-event accuracy) of the locations within earthquake clusters can be improved by applying the above mentioned double-difference location algorithm. Several of these developments are summarized in a monograph edited by Thurber and Rabinowitz (2000), which also includes advances in global seismic event location procedures (Thurber and Engdahl, 2000).



**Figure 5** Examples of improving the ABCE locations for earthquake clusters (red dots) from regional networks of seismic stations (triangles) in China by relocating the events with the double-difference location algorithm (courtesy of Paul G. Richards).

A special volume about event location in context with the special requirements for monitoring the CTBT was published by Ringdal and Kennett (2001). During the IASPEI General Assembly in Santiago de Chile (2005) a workshop had been organized about the question how the ISC location procedures can be improved. Many contributions to this workshop were later published in a special volume (Schweitzer and Storchak, 2006).

## 4 Location errors

### 4.1 Error quantification and statistics

Since seismic events are located with arrival times that contain observational errors and the travel times are calculated under the wrong assumption that the exact model is known, all hypocenter determinations will have errors. Contouring the grid search RMS (Figure 4) gives an indication of the uncertainty of the epicenter. Likewise it would be possible to make 3-D contours to get an indication of the 3-D uncertainty. The question is now how to quantify this measure. The RMS of the final solution is very often used as a criterion for ‘goodness of fit’. Although it can be an indication, RMS depends on the number of stations and does not in itself give any indication of errors and RMS is not reported by, e.g., NEIC and ISC.

From Figure 4 it is seen that the contours of equal RMS are not circles. We can calculate contours within which there is a 67 % probability (or any other desired probability) of finding the epicenter (see below). We call this the error ellipse. This is the way hypocenter errors normally are represented. It is therefore not sufficient to give one number for the hypocenter error since it varies spatially. Standard catalogs from NEIC and ISC give the errors in latitude,

longitude and depth, which can also be very misleading unless the error ellipse has the minor and major axis NS or EW. In the example in Figure 4, this is not the case. Thus the only proper way to report error is to give the full specification of the error ellipsoid.

Before going into a slightly more formal discussion of errors, let us try to get a feeling for which elements affect the shape and size of the epicentral error ellipse. If we have no arrival time errors, there are no epicenter errors so the magnitude of the error (size of error ellipse) must be related to the arrival time uncertainties. If we assume that all arrival time reading errors are equal, only the size and not the shape of the error ellipse can be affected. So what would we expect to give the shape of the error ellipse? Figure 4 is an example of an elongated network with the epicenter off to one side. It is clear that in the NE direction, there is a good control of the epicenter since S-P times control the distances in this direction due to the elongation of the network. In the NW direction, the control is poor because of the small aperture of the network in this direction. We would therefore expect an error ellipse with the major axis NW as observed. Another way of understanding, why the error is larger in NW than it is in NE direction, is to look at Equation (12). The partial derivatives  $\partial T/\partial x$  will be much smaller than  $\partial T/\partial y$  so the  $\Delta y$ -terms will have a larger weight than the  $\Delta x$ -terms in the equations (strictly speaking the partial derivatives with respect to NW and NE). Consequently, errors in arrival times will affect  $\Delta x$  more than  $\Delta y$ . Note that if backazimuth observations were available for any of the stations far in the North or the South of the event, this would drastically reduce the error estimate in the EW direction since  $\partial \phi/\partial x$  is large while  $\partial \phi/\partial y$  is nearly zero.

Changing the geometry of the stations would give another shape of the error ellipse. It is thus possible for any network to predict the shape and orientation of the error ellipses, and given an arrival error, also the size of the ellipse for any desired epicenter location. This can, e.g., be used to predict how a change in network configuration would affect earthquake locations at a given site.

In all these discussions, it has been assumed that the errors have Gaussian distribution and that there are no systematic errors like clock errors. It is also assumed that there are no errors in the theoretical travel times, backazimuths, or ray parameter calculations due to unknown structures. This is of course not true in real life; however error calculations become too difficult if we do not assume a simple error distribution and that all stations have the same arrival time error. This means, however, that such error calculations give us only an idea about the “internal” precision of the locations within the model framework on which they are based. They should not be mistaken as being true location errors (see Section 7).

The previous discussion gave a qualitative description of the errors. We will now show how to calculate the actual hypocentral errors from the errors in the arrival times and the network configuration. The most common approach to seismic source location is based on the least squares inversion and a Gaussian distribution of the arrival time errors, in which case the statistics is well understood and we can use the Chi-Square probability density distribution to calculate errors. For a particular location,  $\chi^2$  can be calculated as:

$$\chi^2 = \frac{1}{\sigma^2} \sum_{i=1}^n r_i^2, \quad (19)$$



where  $\sigma$  is the assumed same standard deviation of any one of the residuals and  $n$  is the number of observations. We can now look at the standard statistical tables (extract in Table 2) to find the expected value of  $\chi^2$  within a given probability. As can be seen from the table, within 5% probability,  $\chi^2$  is approximately the number of degrees of freedom ( $ndf$ ), which in our case is  $n-4$ .

**Table 2** The percentage points of the  $\chi^2$  distribution for different numbers of degrees of freedom ( $ndf$ )

| $ndf$ | $\chi^2$ (95%) | $\chi^2$ (50%) | $\chi^2$ (5%) |
|-------|----------------|----------------|---------------|
| 5     | 1.1            | 4.4            | 11.1          |
| 10    | 3.9            | 9.3            | 18.3          |
| 20    | 10.9           | 19.3           | 31.4          |
| 50    | 34.8           | 49.3           | 67.5          |
| 100   | 77.9           | 99.3           | 124.3         |

If e.g., an event is located with 24 stations ( $ndf=20$ ), there is only a 5% chance that  $\chi^2$  will exceed 31.4. The value of  $\chi^2$  will grow as we move away from the best fitting epicenter and in the example above, the contour within which  $\chi^2$  is less than 31.4 will show the error ellipse within which there is a 95 % chance of finding the epicenter. In practice, errors are mostly reported within 67 % probability, which mean with  $\pm 1$  standard deviation.

The errors in the hypocenter and origin time can also formally be defined with the variance – covariance matrix  $\sigma_X^2$  of the hypocentral parameters. This matrix is defined as

$$\sigma_X^2 = \begin{Bmatrix} \sigma_{xx}^2 & \sigma_{xy}^2 & \sigma_{xz}^2 & \sigma_{xt}^2 \\ \sigma_{yx}^2 & \sigma_{yy}^2 & \sigma_{yz}^2 & \sigma_{yt}^2 \\ \sigma_{zx}^2 & \sigma_{zy}^2 & \sigma_{zz}^2 & \sigma_{zt}^2 \\ \sigma_{tx}^2 & \sigma_{ty}^2 & \sigma_{tz}^2 & \sigma_{tt}^2 \end{Bmatrix}. \quad (20)$$

The diagonal elements are variances of the location parameters  $x$ ,  $y$ ,  $z$  and  $t_0$  while the off diagonal elements give the coupling between the errors in the different hypocentral parameters. For more details, see e.g., Stein (1991). The nice property about  $\sigma_X^2$  is that it is simple to calculate:

$$\sigma_X^2 = \sigma^2 * (G^T G)^{-1}, \quad (21)$$

where  $\sigma^2$  is the variance of the arrival times multiplied by the identity matrix and  $G^T$  is  $G$  transposed. The standard deviations of the hypocentral parameters are thus given by the square root of the diagonal elements and these are the usual errors reported. So how can we use the off diagonal elements? Since  $\sigma_X^2$  is a symmetric matrix, a diagonal matrix in a coordinate system, which is rotated relatively to the reference system, can represent it. We now only have the errors in the hypocentral parameters, and the error ellipse simply have semi axes  $\sigma_{xx}$ ,  $\sigma_{yy}$ , and  $\sigma_{zz}$ . The main interpretation of the off diagonal elements is thus that they define the orientation and shape of the error ellipse. A complete definition therefore requires 6 elements. Eqs. (20) and (21) also show, as stated intuitively earlier, that the shape and

orientation of the error ellipse depends only on the geometry of the network and the crustal structure whereas the standard deviation of the observations is a scaling factor as long they are equal for all observations.

The critical variable in the error analysis is therefore the arrival-time variance  $\sigma^2$ . This value is usually larger than it would be expected from timing and picking errors alone, however it might vary from case to case. Setting a fixed value for a given data set could result in unrealistic error calculations. Most location programs will therefore estimate  $\sigma$  from the residuals of the best fitting hypocenter:

$$\sigma^2 = \frac{1}{ndf} \sum_{i=1}^n r_i^2. \quad (22)$$

Division by  $ndf$  rather than by  $n$  compensates for the improvement in fit resulting from the use of the arrival times from the data. However, this only partly works and some programs allow setting an a priori value which is used only if the number of observations is small. For small networks this can be a critical parameter.

Recently, some studies (e.g., Di Giovambattista and Barba, 1997; Parolai et al., 2001) showed, both for regional and local seismic networks, that the error estimates ERH (in horizontal) and ERZ (in vertical direction), as given by routine location programs (e.g., in HYPOELLIPSE (Lahr, 1989 and 2003)) cannot be considered as a conservative estimate of the true location error and might lead investigators to unjustified tectonic conclusions (see also Figures 11 and 12).

## 4.2 Example of error calculation

We can use the previous error free example (see Table 1) and add some errors (from Stein, 1991). We add Gaussian errors with a mean of zero and a standard deviation of 0.1 s to the arrival times. Now the data are inconsistent and cannot fit exactly. As it can be seen from the results in Table 3, the inversion now requires 3 iterations (2 before) before the locations stop changing. The final location is not exactly the location used to generate the arrival times and the deviation from the correct solution is 0.2, 0.4, and 2.2 km for x, y, and z respectively, and 0.2 s for the origin time. This gives an indication of the location errors.

It is now interesting to compare what is obtained with the formal error calculation. Table 4 gives the variance – covariance matrix. Taking the square root of the diagonal elements we get the standard deviations of x, y, z and  $t_0$  as 0.3, 0.3 and 1.1 km and 0.1 s, respectively. This is close to the ‘true’ error so the solution is quite acceptable. Also note that the RMS is close to the standard error.

**Table 3** Inversion of arrival times with a 0.1 s standard error. As in Table 1, hypocenter is the correct location, Start is the initial location, and the locations are shown after the three following iterations. e is the misfit according to Equation (11).

|        | Hypocenter | Start | 1. Iteration | 2. Iteration | 3. Iteration |
|--------|------------|-------|--------------|--------------|--------------|
| x [km] | 0.0        | 3.0   | -0.2         | 0.2          | 0.2          |
| y [km] | 0.0        | 4.0   | -0.9         | -0.4         | -0.4         |

|                     |      |      |      |      |      |
|---------------------|------|------|------|------|------|
| z [km]              | 10.0 | 20.0 | 12.2 | 12.2 | 12.2 |
| t <sub>0</sub> [s]  | 0.0  | 2.0  | 0.0  | -0.2 | -0.2 |
| e [s <sup>2</sup> ] |      | 93.7 | 0.33 | 0.04 | 0.04 |
| RMS [s]             |      | 3.1  | 0.25 | 0.06 | 0.06 |

**Table 4** Variance – covariance matrix for the example in Table 3.

|   | x    | y     | z     | t     |
|---|------|-------|-------|-------|
| x | 0.06 | 0.01  | 0.01  | 0.00  |
| y | 0.01 | 0.08  | -0.13 | 0.01  |
| z | 0.01 | -0.13 | 1.16  | -0.08 |
| t | 0.00 | 0.01  | -0.08 | 0.0   |

The variance – covariance matrix shows some interesting features. As seen from the diagonal elements of the variance – covariance matrix, the error is much larger in the depth estimate than in x and y. This clearly reflects that the depth is less well constrained than the epicenter which is quite common unless there are stations very close to the epicenter and thus  $|(\mathbf{d}-\Delta)| / \Delta \gg 1$ . For simplicity, we have calculated the standard deviations from the diagonal terms, however since the off diagonal terms are not zero, the true errors are larger. In this example it can be shown that the semi-major and semi-minor axis of the error ellipse have lengths of 0.29 and 0.24 km respectively, and the semi-major axis trends N22°E, so the difference from the original diagonal terms is small.

The z,t or (t,z) term, the covariance between depth and origin time, is negative, indicating a negative trade-off between the focal depth and the origin time; an earlier source time can be compensated by a larger source depth and vice versa. This is commonly observed in practice and is more prone to happen if only first P-phase arrivals are used such that there is no strong control of the source depth by P times measured at different distances.

Error calculation is a fine art, there are endless variations on how it is done and different location programs will usually give different results.

## 5 Relative location methods

### 5.1 Master event technique

The relative location between events within a certain region can often be made with a much greater accuracy than the absolute location of any of the events. This is the case when velocity variations outside the local region are the major cause of the travel-time residuals such that residuals measured at distant stations will be very similar for all of the local events. Usually, the events in the local area are relocated relative to one particularly well-located event, which is then called the **master event**.

Since Gutenberg and Richter (1937), hypocenters of closely located events have been estimated relatively to each other by choosing one of them as a master event. Later, algorithms were developed to solve the problem for larger sets of events in one common

inversion scheme (e.g., Douglas, 1967; Evernden, 1969; Dewey, 1972; Dewy and Algermissen, 1974). Today, many program packages can be found that locate large sets of events more or less closely spaced together, based on different solution strategies (e.g., Kissling et al., 1994; Pujol, 2000; Waldhauser and Ellsworth, 2000) depending on the availability of data types and research interests. It should be mentioned that the Master Event Technique can only be used when the distance to the stations is much larger than the distance between the events.

Most location programs can be used for a master event location. For this travel-time anomalies outside the source region are assumed to cause all individual station residuals after the location of the master event. By using these station residuals as station corrections, the location of the remaining events will be made relative to the master event since all relative changes in arrival times are now entirely due to changes in location within the source region. It is obvious that only stations and phases for which observations are available for the master event can be used for the remaining events. Ideally, the same combination of stations and phases should be used for all events.

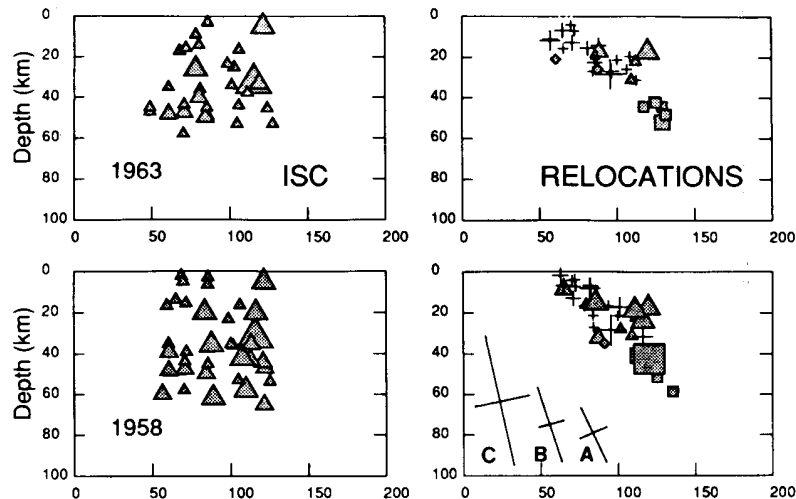
## 5.2 Joint hypocenter location

In the Master Event Technique, it was assumed that true structure dependent residuals could be obtained absolutely correct from the master event, however, other errors could be present in the readings for the master event itself. A better way is to determine the most precise station residuals using the whole data set. This is what Joint Hypocenter Determination (JHD) is about (Douglas, 1967, Pujol, 2000; 2003). Instead of determining one hypocenter and origin time, we will jointly determine  $m$  hypocenters and origin times, and  $n$  station corrections. This is done by adding the station residuals  $\Delta t_i^s$  to Equation (13) and writing the equations for all  $m$  events (index  $j$ ):

$$r_{ij} = (\partial T / \partial x_{ij}) * \Delta x + (\partial T / \partial y_{ij}) * \Delta y + (\partial T / \partial z_{ij}) * \Delta z + \Delta t_i^s + \Delta t_j. \quad (23)$$

The first to propose the JHD method was Douglas (1967). Since the matrix  $\mathbf{G}$  of Equation (14) is now much larger than the  $4 \times 4$  matrix for a single event location, efficient inversion schemes must be used. If we use e.g., 20 stations with 2 phases each for 10 events, there will be  $20 * 10 * 2 = 400$  equations and 80 unknowns (10 hypocenters and origin times, and 20 station residuals).

The relative locations obtained by the Master Event Technique or the JHD are usually more reliable than individually estimated relative locations. However, only if we have the absolute location of one of the events (e.g., a known explosion), we will be able to convert the relative locations of a Master Event algorithm into absolute locations, whereas for the JHD “absolute” locations are obtained for all events if the assumed velocity model is correct. Accurate relative locations are useful to study, e.g., the structure of a subduction zone or the geometry of an aftershocks area, which might indicate the orientation and geometry of the fault. Pujol (2000) has given a very detailed outline of the method and its application to data from local seismic networks. Figure 6 shows an example for increased location accuracy after applying JHD.



**Figure 6** Comparison of earthquake locations using the normal procedure at ISC (left) and JHD relocations (right). The events are located in the Kurile subduction zone along the rupture zones of large thrust events in 1963 and 1958. The vertical cross sections shown traverse the thrust zone from left to right. Note that the JHD solutions reduce the scatter and make it possible to define a dipping plane (from Schwartz et al., 1989).

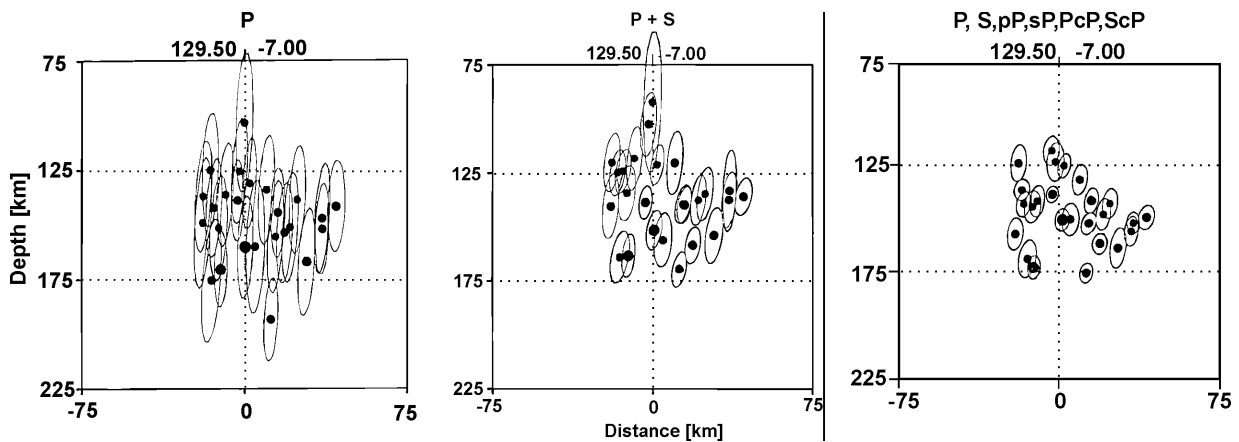
## 6 Practical consideration in seismic source locations

This section is intended to give some practical hints on seismic source location. Different location programs, in particular for observations at local and regional distances, have been developed and are in use since many years (e.g., HYPO71 (Lee and Lahr, 1975; Lee et al., 2003), HYPOINVERSE (Klein, 1978; 1985, 2002, 2003), HYPOELLIPSE (Lahr, 1982; 2003)). This Section does not refer to any particular location program, but most of the parameters discussed can be used with the programs HYPOCENTER (Lienert et al., 1988; Lienert, 1991; Lienert and Havskov, 1995) or HYPOSAT (Schweitzer, 2001, see also PD 11.1), which can locate seismic sources at local, regional and teleseismic distances.

### 6.1 Seismic phases

For the IASPEI recommended seismic phase-naming conventions see IS 2.1 or Storchak et al. (2003). The most unambiguous seismic phase to pick is usually the first P phase. Such P phases are the main phase observations used for most teleseismic locations. For seismic events at local and regional distances, usually S phases are also used. Using phases with different seismic velocities and/or slownesses has the effect of better constraining the distances and there is then less trade-off between depth and origin time or epicenter location and origin time if the epicenter is outside the network. The focal depth is best controlled (with no trade-off between depth and origin time) when seismic phases are included in the location procedure, which have a different sign of the partial derivative  $\partial T/\partial z$  in Equation (13) such as for very locally observed direct up-going Pg (positive) and Pn (negative) (see Section 6.3 Hypocentral depth and Figure 9). In general, it is thus an advantage to use as many different phases as possible under the assumption that they are correctly identified. Schöffel and Das (1999) gave a striking example (see Figure 7). But one very wrong phase can throw off an

otherwise well constrained solution. This highlights the crucial importance of the capability of the observatory personnel to recognize and report such phases during their routine seismogram analysis.



**Figure 7** Examples of significant improvement of hypocenter location for teleseismic events by including secondary phases. Left: hypocenter locations using only P phases; middle: by including S phases; right: by including also depth phases and core reflections with a different sign of  $\partial T/\partial z$  (modified from Schöffel and Das, *J. Geophys. Res.*, Vol. 104, No. B6, page 13, Figure 2; © 1999, by permission of American Geophysical Union).

Engdahl et al. (1998) used the entire ISC database to relocate more than 100,000 seismic events. They used not only a new scheme to associate correctly secondary phases, they also systematically searched for pwP onsets in the case of subduction-zone events to get better depth estimates, and they used a more modern global Earth model (AK135) to avoid the known problems with the Jeffreys-Bullen tables (Jeffreys and Bullen, 1940; 1948; 1958; 1967; 1970). With all these changes the authors reached a far more consistent distribution of hypocenters (in particular for subduction zones) and sharper picture of global seismicity. From data year 2006 on, the ISC is using model AK135 instead of the Jeffreys-Bullen tables in its bulletin productions.

The majority of location programs for local earthquakes use only first arrivals (e.g., HYPO71, Lee and Lahr, 1975). This is good enough for many cases. In some distance ranges, Pn is the first arrival, and it usually has small amplitudes. This means that the corresponding Sn phase, which is then automatically used by the program, might have also very small amplitudes and is not recognized, while actually the phase read is Sg or Lg instead. Since the program automatically assumes a first arrival, a wrong travel-time curve is used for the observed phase, resulting in a systematic location error. This error is amplified by the fact that the S phase, due to its low velocity, has a larger influence on the location than the P phase. It is therefore important to use location programs where all crustal phases can be specified.

Schweitzer (2001) developed an enhanced routine to locate both local/regional and teleseismic events, called HYPOSAT. The program runs with global Earth models and user defined horizontally layered local or regional models with layers of arbitrary velocity gradients. It provides the best possible hypocenter estimates of seismic sources by using travel-time differences between the various observed phases besides the usual input parameters such as arrival times of first and later onsets (complemented by backazimuth and ray parameters in the case of array data or polarization analyses). If S observations are also

available, preliminary origin times are estimated by using the Wadati approach (see Figure 3 in Section 3.2). An initial epicenter with a priori uncertainties can be estimated by calculating the intersection of all backazimuth observations. By relocating events with real data Schweitzer could show that HYPOSAT solutions have the smallest errors when, besides the absolute onset times the travel-time differences of all available primary and secondary phase readings are also taken into account. The most advanced version of HYPOSAT can be downloaded for free either directly from the program list in the NMSOP-2 cover page folder *Download Programs & Files* or from NORSAR's anonymous ftp-server <ftp://ftp.norsar.no/pub/outgoing/johannes/hyposat/> and a detailed program description is given in PD 11.1.

## 6.2 Initial location

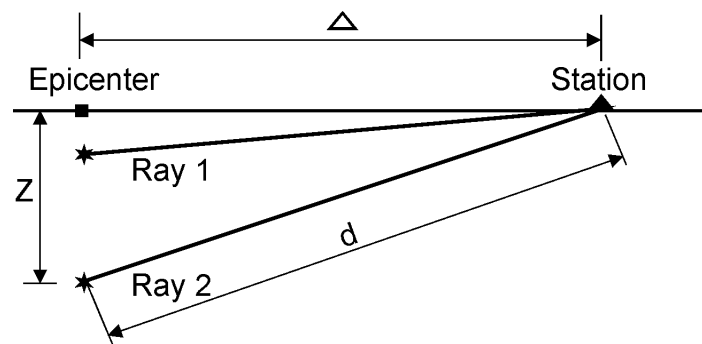
Iterative location programs often start with an initial location at a point near the station recording the earliest arrival. This is good enough for most cases, particularly when the station coverage is good and the epicenter is near or within the network. However, this can also lead to problems when using least squares techniques, which converge slowly or sometimes not at all for events outside the limits of a regional network (Buland, 1976). Another possibility is that the solution converges to a local minimum, which might be far from the correct solution. For small-elongated networks, two potential solutions may exist at equal distances from the long axis. An initial location close to the station with the earliest arrival can then bias the final solution to the corresponding side of such a network. Although this bias usually is on the correct side, any systematic error in the earliest-arrival station's time can have a disproportionately strong effect on the final location. Thus in many cases, it is desirable to use a better start location than the nearest station. There are several possibilities:

- a) in many cases the analyst knows by experience the approximate location and can then manually give an initial location; most programs have this option;
- b) similar phases at different stations can be used to determine the apparent velocity and backazimuth of a plane wave using linear regression on the arrival times (e.g., using the network as a seismic array, see Chapter 9). With apparent velocity and/or S-P times, an estimate of the initial location can be made. This method is particularly useful when locating events far away from the network (regionally or globally);
- c) backazimuth information is frequently available from 3-component stations or seismic arrays and can be used as under b);
- d) if backazimuth observations are available from different stations, an initial epicenter can be determined by calculating the intersection of all backazimuth observations;
- e) S-P and the circle method can be used with pairs of stations to get an initial location;
- f) the Wadati approach can be used to determine an initial source time.

The initial depth is usually a fixed parameter and set to the most likely depth for the region. For local earthquakes usually a depth range of 10-20 km is used, while for distant events, the initial depth was often set to 33 km, which is the crustal thickness in the Jeffreys-Bullen spherical Earth model. If depth phases, e.g., pP are available for distant events, these phases can be used to set or fix the depth (see next section).

## 6.3 Hypocentral depth

The hypocentral depth is the most difficult parameter to determine due to the fact that the travel-time derivative with respect to depth changes very slowly as function of depth (see Figure 8) unless the station is very close to the epicenter. In other words, the depth can be moved up and down without changing the travel time much. Figure 8 shows a shallow (ray 1) and a deeper event (ray 2). It is clear that the travel-time derivative with respect to depth is nearly zero for ray 1 but not for ray 2. In this example, it would thus be possible to get a depth estimate for the deeper event but not for the shallower one. Unfortunately, at larger distances from the source, most rays are more like ray 1 than ray 2 and locations are therefore often made with a fixed ‘normal’ start depth. Only after a reliable epicenter is obtained will the program try to iterate for the depth. Another possibility is to locate the event with several initial depths and then use the depth that gives the best fit to the data. Although one depth will give a best fit to all data, the depth estimate might still be very uncertain and the error estimate must be checked.



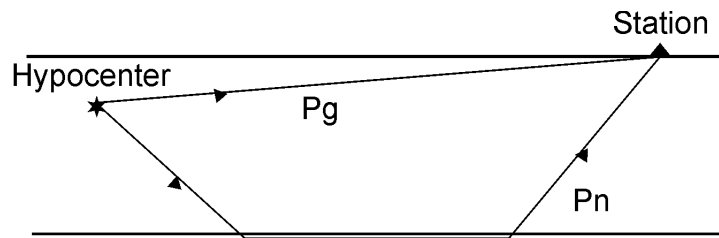
**Figure 8** The depth – distance trade off in the determination of focal depth.

For teleseismic events, the best way to improve the depth determination is to include readings from the so-called depth phases (e.g., Gutenberg and Richter, 1936 and 1937; Engdahl et al., 1998) such as pP, pwP (reflection from the ocean free surface), sP, sS or similar but also reflections from the Earth's core like PcP, ScP or ScS (see Figure 7). The travel-time differences (i.e., depth phase-direct phase) as pP-P, sP-P, sS-S and pS-S are quite constant over a large range of epicentral distances for a given depth so that the depth can be determined nearly independently of the epicentral distance. Another way of getting a reliable depth estimate for teleseismic locations is to have both near and far stations available. In particular, event observations from local and regional stations together with PKP observations have been used together for this purpose. However, this is unfortunately not possible for many source regions and small magnitude events.

For local events, a rule of thumb is that at least several near stations should not be further away than 2 times the depth in order to get a reliable estimate (Figure 8). This is very often not possible, particularly for regional events. At a distance of more than  $2 \times \text{depth}$ , the depth-depending partial derivative changes very little with depth if the first arriving phase is the more or less horizontally propagating Pg. But at distances where the critically refracted (so-called head-waves) Pb or Pn arrive, there is again some sensitivity to depth due to the more steeply down going rays of Pb or Pn (Figure 9) and because of the different sign of the partial derivatives of their travel times with depth, which is negative for Pn and positive for Pg. Thus, if stations are available at distances where both direct and refracted rays are first arrivals, reasonably reliable depth solutions might be obtained. An even better result one gets when both Pg and Pn arrivals can reliably be picked at the same station. The accuracy of [the](#)

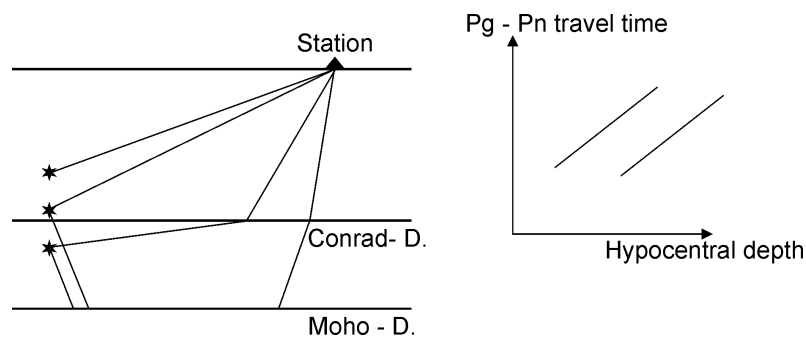


hypocenter location is then comparable to using P and pP for teleseismic events. Yet, correct identification of secondary P phases is often difficult and a wrong identification might make matters worse.



**Figure 9** Example of both Pg and Pn rays in a simplified, single layer crustal model.

The depth estimate using a layered crustal model remains problematic even with a mix of phases. In checking catalogs with local earthquakes, it will often be noted that there is a clustering of hypocenters at layer boundaries. This is caused by the discontinuities in the travel-time curves of the direct phase Pg as a function of depth at layer boundaries (see Figure 10 as an example). The Pg-travel time suddenly decreases when the hypocenter crosses a boundary (here Moho) since a larger part of the ray suddenly is in a higher velocity layer, while the Pn-travel time continuously decreases as the depth increases as long as the event is still within the crust. This gives rise to the discontinuities in the Pg-Pn travel-time curve. So one Pn-Pg travel-time difference is not enough to ensure a reliable depth estimate, several such phase arrivals must be available.



**Figure 10** Simplified two-layer crustal model of ray paths of Pg and Pn phases (left). On the right-hand panel the travel-time difference Pg-Pn as a function of depth is qualitatively sketched. It is assumed that the distance is large enough for Pn to be the first arrival.

Even when several Pg, Pb (very seldom) and Pn phases are available, depth estimates still remain a problem at regional distances due to the uncertainty in the crustal models. Since the depth estimates critically depend on accurate travel-time calculations for these phases, small model uncertainties can quickly throw off the depth estimate.

Many location programs give the RMS of the travel-time residuals in a grid around the calculated hypocenter. In addition to the error estimates, this gives an idea about the accuracy and thus a local minimum might be found. A more direct way of estimating the quality of the depth estimate is to calculate the RMS as a function of depth in order to check if a local

minimum has been reached. This is particularly relevant for crustal earthquakes at shallow depth and can also be used as a complementary tool for discriminating better between quarry blasts and earthquakes.

## 6.4 Outliers and weighting schemes

The largest residuals have a disproportionately large influence on the fit of the arrival times due to the commonly used least squares fit. Most location programs will have some kind of residual weighting scheme in which observations with large residuals are given lower or even no weight. Bisquare weighting is often used for teleseismic events (Anderson, 1982). The residual weighting works very well if the residuals are not extreme since the residual weighting can only be used after a few iterations when the residuals are already close to the final ones. Individual large residuals can often lead to completely wrong solutions, even when 90% of the data are good; residual weighting will not help in these cases. Some programs will try to scan the data for gross errors (like minute errors) before starting the iterative procedure. If an event has large residuals, try to look for obvious outliers. A Wadati diagram can often help in spotting bad readings for local events (see Figure 3 in Section 6.2).

The arrival-time observations by default will always have different weights in the inversion. A simple case is that S waves may have larger weights than P waves due to their lower velocities. An extreme case is the T wave (a guided wave in the ocean), which with its low velocity (1.5 km/s) can completely dominate the solution. Considering that the accuracy of the picks is probably best for the P waves, it should be natural that P arrivals have more importance than S arrivals in the location. However, the default parameter setting in most location programs is to leave the original weights unless the user actively changes them. It is normally possible to give *a priori* for all S phases a lower weight and in addition, all phases can be given individual weights, including being totally weighted out.

When working with local seismic sources, the nearest stations will usually provide the most accurate information due to the clarity of the phases. In addition, uncertainty in the local model has less influence on the results at short distances than at large distances; this is particularly true for the depth estimate. It is therefore desirable to put more weight on data from near stations than on those from distant stations and this is usually done by using a distance weighting function of

$$w_d = \frac{x_{far} - \Delta}{x_{far} - x_{near}}, \quad (24)$$

where  $\Delta$  is the epicentral distance,  $x_{near}$  is the distance to which full weight is used and  $x_{far}$  is the distance where the weight is set to zero (or reduced). The constants  $x_{near}$  and  $x_{far}$  are adjusted to fit the size of the network;  $x_{near}$  should be about the diameter of the network, and  $x_{far}$  about twice  $x_{near}$ . For a dense network,  $x_{near}$  and  $x_{far}$  might be made even smaller for more accurate solutions.

## 6.5 Ellipticity of the Earth

Until now we only assumed that the model used for calculating distances or travel times is either a flat model for local or regional events or a standard spherical model of the Earth for teleseismic events. However, the Earth is neither a sphere nor a flat disk but an ellipsoid symmetrical to its rotation axis. It was Gutenberg and Richter (1933) who first pointed out that the difference between a sphere and an ellipsoid must be taken into account when calculating epicentral distances and consequently also the travel times of seismic phases. Therefore, they proposed the usage of geocentric coordinates instead of geographic coordinates to calculate distances and angles on the Earth. Because of the axially symmetrical figure of the Earth, the geocentric longitude is identical to the geographic longitude. To convert a geographic latitude  $lat_g$  into a geocentric latitude  $lat_c$  one can use the following formula:

$$lat_c = \arctan((1 - (6378.136 - 6356.751) / 6378.136)^2 * \tan lat_g). \quad (25)$$

With this formula all station latitudes have to be converted before an event location and after the inversion, the resulting geocentric event latitude has to be converted back by applying the inverse equation

$$lat_g = \arctan(\tan lat_c / (1 - (6378.136 - 6356.751) / 6378.136)^2). \quad (26)$$

With this procedure all angle calculations related to an event location are done for a sphere. The calculated distances are measured in degrees and to convert them into km, one has to use the local Earth radius  $R_{loc}$ :

$$R_{loc} = \sqrt{(6378.136 * \cos lat_c)^2 + (6356.751 * \sin lat_c)^2}. \quad (27)$$

This value has then to be applied for converting a distance  $D$  measured in degrees into a distance measured in km, or vice versa:

$$D[km] = \frac{2\pi * R_{loc}}{360} * D[deg] \quad \text{or} \quad D[deg] = \frac{360}{2\pi * R_{loc}} * D[km] \quad (28)$$

All standard Earth models are defined for a spherically symmetrical Earth with a mean radius of 6371 km. Therefore the standard tables also contain travel times calculated for a sphere. Bullen (1937, 1938, 1939) was the first to calculate latitude-depending travel-time corrections (ellipticity corrections) to be used together with travel-time tables for a spherical Earth. Later work on this topic was done by Dziewonski and Gilbert (1976) and Dornboos (1988b). Kennett and Gudmundsson (1996) published a set of ellipticity corrections for a large number of seismic phases calculated applying the spherical Earth model AK135.

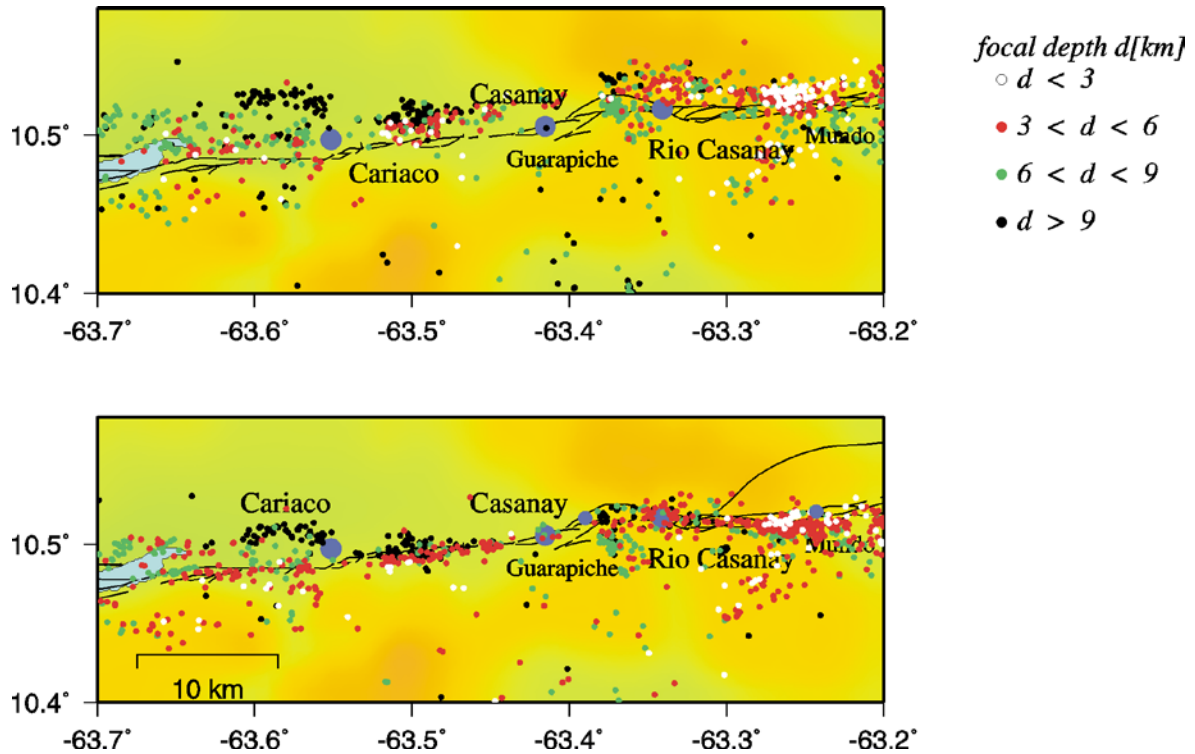
In conclusion: to get the theoretical travel time for an event in teleseismic or regional distance, one has first to calculate the geocentric epicentral distance, then use travel-time tables as calculated for a spherical Earth model, and finally apply the latitude (event and station!) dependent ellipticity correction. Most location routines automatically apply the described methods and formulas but it is important to check this in detail and eventually to change a location program.

## 6.6 Importance of the velocity model

In this context the importance of the model assumptions underlying the location procedure has to be emphasized. The chosen 1-D regional velocity model has a strong influence on the location of seismic sources (e.g., Kennett, 1992) and many studies have shown (e.g., Kissling, 1988) that the accuracy of locating hypocenters can even be more improved by using station corrections in addition to a well-constrained global 1-D velocity. However, Spallarossa et al. (2001) showed that in strongly heterogeneous local areas even a 1-D model with station corrections does not significantly improve the accuracy of the location parameters. High-precision location in such cases can only be achieved by using a 3-D model. This is particularly true for locating earthquakes in volcanic areas (see Lomax et al., 2001).

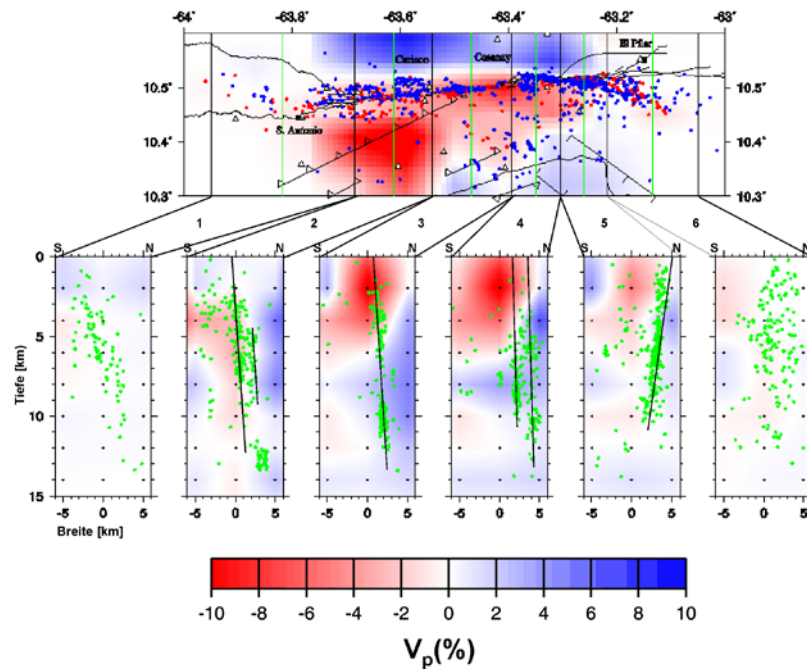
Smith and Ekström (1996) investigated the improvement of teleseismic event locations by using a 3-D Earth model. They came to the conclusion that it “... offers improvement in event locations over all three 1-D models with, or without, station corrections.” For the explosion events, the average mislocation distance is reduced by approximately 40%; for the earthquakes, the improvements are smaller. Corrections for crustal thickness beneath source and receiver are found to be of similar magnitude to the mantle corrections, but use of station corrections together with the 3-D mantle model provide the best locations. Also Chen and Willemann (2001) carried out a global test of seismic event locations using 3-D Earth models. Although a tighter clustering of earthquakes in subduction zones was achieved by using a 3-D model rather than using depth from the ISC Bulletin based on 1-D model calculations, they concluded that the clustering was not as tight as for depths computed by Engdahl et al. (1998) who used depth phases as well as direct phases. Thus, even using the best available global 3-D models can not compensate for the non-use of depth phases and core reflections in teleseismic hypocenter location (see Figure 7).

An example case for the improved location of local earthquakes is given in Figures 11 and 12. The upper panel in Figure 11 shows the initial epicenter locations of aftershocks of the Cariaco earthquake ( $M_s = 6.8$ ) on July 9, 1997 in NE Venezuela based on an averaged 1-D crustal velocity model. The mean location error (i.e., the calculated *precision* with respect to the assumed model) was about 900 m. On average, the aftershocks epicenters occurred about 2 to 3 km north of the surface fault trace.



**Figure 11** Epicentral distribution of aftershocks of the Cariaco earthquake ( $M_s=6.8$ ) on July 9, 1997 in NE Venezuela. Top: results from HYPO71 based on a one-dimensional velocity-depth distribution. Bottom: Relocation of the aftershocks on the basis of a 3-D model derived from a tomographic study of the aftershock region (courtesy of M. Baumbach () H. Grosser and A. Rietbrock).

A detailed tomographic study revealed lateral velocity contrasts of up to 20% with higher velocities towards the north of the El Pilar fault (Figure 12). Relocating the events with the 3-D velocity model the epicenters were systematically shifted southward by about 2 km and now their majority aligned rather well with fault traces mapped before the earthquake as well as with newly ruptured fault traces (Figure 11, lower panel). Also in the cross sections the data scatter was clearly reduced so that closely spaced outcropping surface faults could be traced down to a depth of more than 10 km. These results point to the fact that in the presence of lateral velocity inhomogeneities epicenter locations are systematically displaced in the direction of higher velocities. We will look into this problem more closely in Section 7.



**Figure 12** 3-D distribution of the P-wave velocity in the focal region of the 1997 Cariaco earthquake as derived from a tomographic study. The horizontal section shows the velocity distribution in the layer between 2 km and 4 km depth. Red and blue dots mark the epicenters of the aftershocks. The red ones were chosen because of their suitability for the tomography. The six vertical cross sections show depth distributions of the aftershocks (green dots) together with the deviations of the P-wave velocity from the average reference model. The depth range and the lateral changes of fault dip are obvious (courtesy of M. Baumbach<sup>(†)</sup>, H. Grosse and A. Rietbrock).

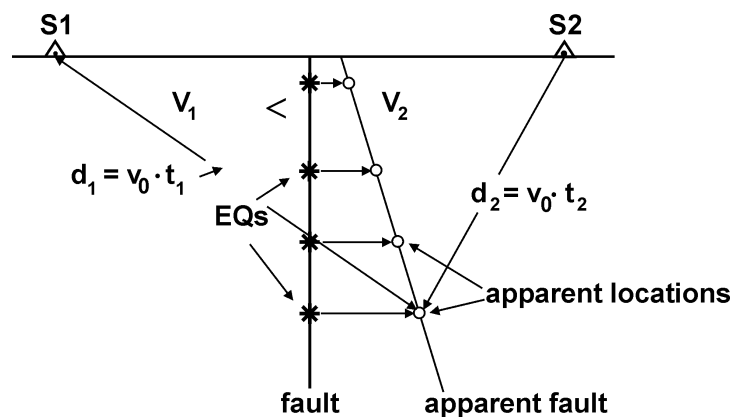
## 7 Internal and external (real) accuracy of locations

For decades the international data centers have located earthquakes world-wide by means of the 1-D Jeffreys-Bullen velocity model of the Earth and Jeffreys-Bullen travel-time tables (Jeffreys and Bullen, 1940; 1948; 1958; 1967; 1970) without external control of the accuracy of such solutions by independently checking them with similarly strong events of exactly known position and origin time. Therefore, the question has remained open for a long time as to whether these calculated location errors were real or just the minimized average errors for the best fitting solutions to the observed data based on model assumptions with respect to the validity of the velocity model, the non-correlation of the various parameters to be determined and the Gaussian distribution of both the model errors and the data reading errors. If the latter is the case then the calculated errors are no measure of the real accuracy of the calculated location and origin time but rather a measure of the internal precision of fitting best the data to the used model.

In order to investigate this in more detail, Bormann (1972a and b) looked into the travel-time errors reported by the international data centers for the German seismological observatory Moxa (MOX) for earthquakes in different regions of the world. As an example, he got for the same data set of events from the Kurile Islands the mean residual  $\bar{\delta}t_p = + 0.16$  s and a standard deviation  $\sigma = \pm 0.65$  s when referring the MOX onset-time readings to the locations

published by the U.S. Coastal and Geodetic Survey (USCGS, World Data Center A, WDC A) and  $\bar{\delta}t_p = +0.35$  s with  $\sigma = \pm 1.1$  s when referring to the locations published by the Academy of Sciences of the Soviet Union (ANUSSR, World Data Center B, WDC B) which used the same Jeffreys-Bullen travel-time tables as the USCGS. Thus, the travel-time (or onset-time reading) errors calculated by the data centers for seismic stations are not real errors of these stations or their readings but depend on the number and distribution of stations used by these centers in their location procedure. And these were rather different for WDC A and WDC B. While the USCGS used the data of a worldwide station network, ANUSSR based its locations on the station network of the former Soviet Union and East European countries and these “looked at” events outside Eurasia from a much narrower azimuth and distance range. But this is equivalent to the discussion related to Figure 4. Although the mean station residuals calculated by these two data centers for the considered region were not significantly different and not far from zero, their systematic mislocations with respect to “ground truth” locations such as those of underground nuclear explosions (Herrin and Taggart, 1968; Bormann, 1972a; Engdahl et al., 1998) reached in subduction zones with large lateral velocity inhomogeneities several 10 km and were larger for the network with a limited azimuthal coverage. Therefore, the ISC waits for one to two years before running its final location procedure. This allows collecting as many seismic phase readings as possible from worldwide distributed seismological observatories and thus assures the best geographic coverage for each seismic event.

What is the reason for this systematic mislocation, which usually remains unrecognized unless one locates strong independently controlled sources of exactly known source parameters and origin time? Figure 13 shows some hypothetical earthquakes at different depth on a vertically dipping fault. It separates two half-spaces with different wave propagation velocity  $v_2 > v_1$ . This is a realistic model for parts of the San Andreas Fault. The lateral velocity difference across the fault may be as large as 5 to 7 %. S1 and S2 may be two stations at the same hypocentral distances from the events. But because of  $v_2 > v_1$  the onset time  $t_2$  at S2 is earlier (travel-time shorter) than for  $t_1$  at S1. Running the location procedure with the common residual minimization on the assumption of a laterally homogeneous velocity model will result in hypocentral distances  $d_2(h) < d_1(h)$ . Since the difference increases with depth, the hypocenters are not only offset from the real fault but seem to mark even a slightly inclined fault, which is not the case.



**Figure 13** Illustration of the systematic mislocation of earthquakes along a fault with strong lateral velocity contrast.  $v_0$  is the assumed model velocity with  $v_2 > v_0 > v_1$ .

From this hypothetical example we learn that locations based on 1-D velocity models in the presence of 2-D or 3-D velocity inhomogeneities will be systematically shifted in the direction of increasing velocities (or velocity gradients), the more so, the less the station distribution controls the event from all azimuths. This is precisely the cause for the above mentioned larger systematic mislocations of WDC B as compared to WDC A. While the latter localizes events using data from a global network, the former used solely data from the former Soviet and East European territory, i.e., stations which view the Aleutian Islands from only a narrow azimuth range. The direction of systematic mislocation of both centers to the NW agrees with the NW directed subduction of the Pacific plate underneath the Aleutians. According to Jacob (1972) this cold lithospheric plate has 7 to 10% higher P-wave velocities than the surrounding mantle. A study by Lienert (1997) also addresses this problem of assessing the reliability of seismic event locations by using known nuclear tests.

The above discussion illustrates that we have to differentiate between the *precision* and the *accuracy* of locations (for definition of these terms see also the Glossary) and highlights the utmost importance of the general availability and easy accessibility of so-called seismic reference events with very well known hypocenter coordinates and origin time. They are needed for tomographic studies, the derivation of improved local, regional and global velocity models, the calibration and quality assessment of seismic network and array location procedures and other related issues. Such events may be chemical or nuclear underground explosions, which are commonly known exactly, or earthquakes that have been precisely located by available dense local networks. Such reference events can be considered as independent “ground truth” (GT) data. They are classified according to the accuracy with which the event coordinates are known in terms of integer values of kilometer (GT0, GT1, GT2, GT5, GT10). GT0 to GT2 data are available only for chemical and underground nuclear explosions, GT5 and larger mostly for earthquakes.

For many years the IASPEI Commission on Seismic Observation and Interpretation (CoSOI) had entrusted a Working Group with the elaboration of guidelines for the identification, collection and quality assessment as well as the initial compilation of a world-wide listing of such events. Meanwhile, the International Seismological Centre (ISC) has established a website data base of reference earthquakes and explosions for which hypocenter information is known with high confidence and which were associated with seismic signals that have been recorded at regional and/or teleseismic distances (see <http://www.isc.ac.uk/GT/index.html> ). On this website one finds the actual global IASPEI Reference Event List and map. As of summer 2011 it comprised the following numbers of GT events: GT0 – 650, GT1 – 367, GT2 – 123, GT5 – 6195 (out of which about 6000 are earthquakes). Through this website one can also submit new information about reference events. IASPEI encourages the scientific community to submit to the ISC such reference event information using the IASPEI developed guidelines for identifying candidate events and assessing their quality.

How to identify and collect such ground truth events is described in detail in IS 8.6, and IS 8.5 gives more details about the criteria for assessing the accuracy and reliability of locations depending on the geometry and azimuthal distribution of reporting stations, specifically about the GT5 criteria at 90% and 95% confidence level for local networks as defined by Engdahl et al. (2001), Bondar et al. (2004), and Bondár and McLaughlin (2009), respectively. These criteria are now operationally applied in the authoritative location scheme of the European-Mediterranean Seismological Centre (EMSC; <http://www.emsc-csem.org>) and its real time information services. In this scheme locations are considered to be accurate when they fulfill the GT5 criteria and additionally as reliable if they can be reproduced to within 15 km when



applying different location procedures and velocity models to the same data set. To expect more accurate routine earthquake locations based on regional and common local network data is hardly realistic for earthquakes with magnitudes below 6 and questionable anyway for earthquakes with magnitudes above 7 and thus linear rupture dimensions ranging from several 10 to several 100 kilometers.

## Acknowledgments

Some of the preceding text has followed a similar description in Shearer (1999). Several figures and ideas have also been taken from Stein (1991), Lay and Wallace (1995), Schöffel and Das (1999). Thanks go to M. Baumbach, H. Grosser and A. Rietbrock for making Figures 11 and 12 available and to R. E. Engdahl for critical proof reading and valuable suggestions which helped to improve the text and to complement the references.

## References

- Anderson, K. R. (1982). Robust earthquake location using M-estimates, *Phys. Earth. Plan. Inter.*, **30**, 119-130.
- Båth, M. (1979). Introduction to seismology. *Birkhauser Verlag*, Basel, 428 pp.
- Bondár I. and McLaughlin, K. L. (2009). A new ground truth data set for seismic studies. *Seismol. Res. Lett.*, **80**(3), 465-472
- Bondár, I., Myers, S. C., Engdahl, E. R., and Bergman, E. A. (2004). Epicenter accuracy based on seismic network criteria. *Geophys. J. Intern.*, **156**, 483–496; doi10.1111/j.1365-246X.2004.02070.x.
- Bormann, P. (1972a). A study of travel-time residuals with respect to the location of teleseismic events from body-wave records at Moxa station. *Gerl. Beitr. Geophys.*, **81**, 117-124.
- Bormann, P. (1972b). A study of relative frequencies of body-wave onsets in seismic registrations of the station Moxa. In: *Seismological Bulletin 1967, Station Moxa (MOX)*. Akademie-Verlag, Berlin 1972, 379-397
- Buland, R. (1976). The mechanics of locating earthquakes, *Bull. Seism. Soc. Am.*, **66**, 173-187.
- Buland, R., and Chapman, C. (1983). The computation of seismic travel times. *Bull. Seism. Soc. Am.*, **73**, 1271-1302.
- Bullen, K. E. (1937, 1938, 1939). The ellipticity correction to travel-times of P and S earthquake waves. *Montl. Not. R. Astr. Soc., Geophys. Suppl.*, **4**, 143-157, 158-164, 317-331, 332-337, 469-471.
- Chen, Q. F., and Willemann, R. J. (2001). Global test of seismic event locations using three dimensional Earth models. *Bull. Seism. Soc. Am.*, **91**, 6, 1704-1716.
- Dewey, J. W. (1972). Seismicity and tectonics in western Venezuela. *Bull. Seism. Soc. Amer.*, **62**, 1711-1751.
- Dewey, J. W., and Algermissen, S. T. (1974). Seismicity of the Middle America arc-trench system near Managua, Nicaragua. *Bull. Seism. Soc. Amer.*, **64**, 1033-1048.

- Di Giovambattista, R., and Barba, S. (1997). An estimate of hypocenter location accuracy in a large network, possible implication for tectonic studies in Italy. *Geophys. J. Int.*, **129**, 124-132.
- Doornbos, D. J. (Ed.) (1988a). Seismological algorithms, Computational methods and computer programs. *Academic Press*, New York, xvii + 469 pp.
- Dornboos, D. J. (1988b). Asphericity and ellipticity corrections. In: Dornboos (1988a), 75-85.
- Douglas, A. (1967). Joint hypocenter determination. *Nature*, **215**, 47-48.
- Dziewonski, A. M., and Gilbert, F. (1976). The effect of small, aspherical perturbations on travel times and a re-examination of the corrections for ellipticity. *Geophys. J. R. astr. Soc.* **44**, 7-17.
- Engdahl, E. R., Van der Hilst, R. D., and Buland, R. P. (1998). Global teleseismic earthquake relocation with improved travel times and procedures for depth determination. *Bull. Seism. Soc. Am.*, **88**, 722-743.
- Engdahl, E. R., Bondar, I., Bergmann, E., and Firbas, P. (2001). IASPEI Working group on reference events: 1999-2001 activity report.
- Evernden, J. F. (1969). Identification of earthquakes and explosions by use of teleseismic data. *J. Geophys. Res.*, **74**, 3828-3856.
- Font, Y., Kao, H., Lallemand, S., Liu, C.-S. , and Chiao, L.Y. (2004). Hypocentre determination offshore of eastern Taiwan using the Maximum Intersection method. *Geoph. J. Int.*, **158**, 655-675.
- Geiger, L. (1910). Herdbestimmung bei Erdbeben aus den Ankunftszeiten. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*, Mathematisch-Physikalische Klasse, 331-349. (In 1912 translated to English by Peebles, F. W. L., and Corey, A. H.: Geiger, L. (1912). Probability method for the determination of earthquake epicenters from the arrival time only. *Bulletin St. Louis University*, **8**, 60-71).
- Gutenberg, B., and Richter, C. F. (1933). Advantages of using geocentric latitude in calculating distances. *Gerl. Beitr. Geophysik*, **40**, 380-389.
- Gutenberg, B., and Richter, C. F. (1936). Materials for the study of deep-focus earthquakes. *Bull. Seism. Soc. Am.*, **26**, 341 – 390 (edition in French: Données relatives a l'étude des tremblements de terre a foyer profond. *Publications du Bureau Central Séismologique International*, Série A, Travaux Scientifiques **15**, 1 - 70, 1937).
- Gutenberg, B., and Richter, C. F. (1937). Materials for the study of deep-focus earthquakes (second paper). *Bull. Seism. Soc. Am.*, **27**, 157 – 183 (partly edited in French: Données relatives a l'étude des tremblements de terre a foyer profond. *Publications du Bureau Central Séismologique International*, Série A, Travaux Scientifiques **15**, 1 - 70, 1937).
- Herrin, E. and Taggart, J. (1968). Source bias in epicenter determinations. *Bull. Seism. Soc. Am.*, **58**, 1791–1796.
- Horiuchi, S., Negishi, H., Abe, K., Kamimura, A., and Fujinawa, Y. (2005). An automatic processing system for broadcasting earthquake alarms. *Bull. Seism. Soc. Amer.*, **95**, 708-718.
- Jacob, K. H. (1972). Global tectonic implications of anomalous seismic P travel-times from the nuclear explosion Longshot. *J. Geophys. Res.*, **77**, 14.
- Jeffreys, H., and Bullen, K. E. (1940, 1948, 1958, 1967, and 1970). Seismological Tables. British Association for the Advancement of Science, *Gray Milne Trust*, London, 50 pp.

- Kennett, B. L. N. (1992). Locating oceanic earthquakes - the influence of regional models and location criteria. *Geophys. J. Int.*, **108**, 848-854.
- Kennett, B. L. N. (Ed.) (1991). IASPEI 1991 Seismological Tables. Research School of Earth Sciences, Australian National University, 167 pp.
- Kennett, B. L. N., and Engdahl, E. R. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, **105**, 429-465.
- Kennett, B. L. N., and Gudmundsson, O. (1996). Ellipticity corrections for seismic phases, *Geophys. J. Int.*, **127**, 40-48.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, **122**, 108-124.
- Kikuchi, M., and Fukao, Y. (1987). Inversion of long-period P-waves from great earthquakes along subduction zones. *Tectonophysics*, **144**, 231-247.
- Kissling, E. (1988). Geotomography with local earthquake data. *Rev. Geophys.*, **26**, 659-698.
- Kissling, E., Ellsworth, W. L., Eberhart-Philips, D., and Kradolfer, U. (1994). Initial reference models in local and regional earthquake tomography. *J. Geophys. Res.*, **99**, 19635-19646.
- Klein, F. (2002). User's Guide to HYPOINVERSE-2000, a Fortran program to solve for earthquake locations and magnitudes, *USGS, Open File Report 02-171*, 123pp.
- Klein, F. W. (1978). Hypocenter location program HYPOINVERSE. *U.S. Geol. Surv. Open-File Report*. **78-694**.
- Klein, F. W. (1985). HYPOINVERSE, a program for VAX and professional 350 computers to solve the earthquake locations. *U.S. Geological Survey Open-File Report 85-515*, 53 pp.
- Klein, F. W. (2003). The HYPOINVERSE2000 earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003), 1619-1620.
- Lahr, J. C. (1989). HYPOELLIPSE/Version 2.0: A computer program for determining local earthquakes hypocentral parameters, magnitude, and first motion pattern. *U.S. Geological Survey Open-File Report 89-116*, 92 pp.
- Lahr, J. C. (2003). The HYPOELLIPSE earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003), 1617-1618.
- Lay, T., and Wallace, T. C. (1995). Modern global seismology. ISBN 0-12-732870-X, *Academic Press*, 521 pp.
- Lee, W. H. K., and Lahr, J. C. (1975). HYPO71 (revised): A computer program for determining hypocenter, magnitude and first motion pattern of local earthquakes. *U.S. Geological Survey Open-File Report 75-311*, 116 pp.
- Lee, W. H. K., Lahr, J. C., and Valdes, C. M. (2003). The HYPO71 earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003), 1641-1642.
- Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C., (Eds.), (2003). "International Handbook of Earthquake and Engineering Seismology, Part B", *Academic Press, Amsterdam*, 1012 pp and 2 CD-ROMs.
- Lienert, B. R. E. (1991). Report on modifications made to Hypocenter. *Institute of Solid Earth Physics, University of Bergen*.
- Lienert, B. R. E. (1997). Assessment of earthquake location accuracy and confidence region estimate using known nuclear tests. *Bull. Seism. Soc. Am.*, **87**, 5, 1150-1157.

- Lienert, B. R. E., and Havskov, J. (1995). A computer program for locating earthquakes both locally and globally. *Seism. Res. Lett.*, **66**, 26-36.
- Lienert, B. R. E., Berg, E., and Frazer, L. N. (1988). HYPOCENTER: An earthquake location method using centered, scaled, and adaptively least squares. *Bull. Seism. Soc. Am.*, **76**, 771-783.
- Lomax, A. (2005). A reanalysis of the hypocentral location and related observations for the great 1906 California earthquake. *Bull. Seism. Soc. Amer.*, **95**, 861-877.
- Lomax, A., Virieux, J., Volant, P., and Berge, C. (2000). Probabilistic earthquake location in 3D and layered models: Introduction of a Metropolis-Gibbs method and comparison with linear locations. In: Thurber and C. H., and Rabinowitz, N. (2000). *Advances in Seismic Event Location*. Kluwer, Amsterdam, 101-134.
- Lomax, A., Zollo, A., Capuano, P., and Virieux, J. (2001). Precise, absolute earthquake locations under Somma-Vesuvius volcano using a new three-dimensional velocity model. *Geophys. J. Int.* **146**, 313-331.
- Morelli, A., and Dziewonski, A. M. (1993). Body wave traveltimes and a spherically symmetric P- and S-wave velocity model, *Geophys. J. Int.*, **112**, 178-194.
- Müller, G. (1977). Earth flattening approximation for body waves derived from geometric ray theory - improvements, corrections and range of applicability. *J. Geophys.*, **44**, 429-436.
- Nicholson, T., Gudmundsson, Ó., and Sambridge, M. (2004). Constraints on earthquake epicentres independent of seismic velocity models. *Geophys. J. Int.*, **156**, 648-654.
- Parolai, S., Trojani, L., Monachesi, G., Frapiccini, M., Cattaneo, M., and Augliera, P. (2001). Hypocenter location accuracy and seismicity distribution in the Central Apennines (Italy). *J. Seism.*, **5**, 243-261.
- Pujol, J. (2000). Joint event location – the JHD technique and applications to data from local networks. In: Thurber and Rabinowitz (2000), 163-204.
- Pujol, J. (2003). Software for joint hypocentral determination using local events. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003), 1621-1623.
- Pujol, J. (2004). Earthquake location tutorial: graphical approach and approximate epicentral location techniques. *Seism. Res. Lett.*, **75**, 63-74.
- Ringdal, F., and Kennett, B. L. N. (Eds.) (2001). Special issue: Monitoring the Comprehensive Nuclear-Test-Ban Treaty: Source location. *Pure appl. Geophys.*, **158**, (1/2), 1-419.
- Roberts, R. G., Christofferson, A., and Cassidy, F. (1989). Real time event detection, phase identification and source location using single station 3 component seismic data and a small PC. *Geophys. J.*, **97**, 471-480.
- Rosenberger, A. (2009). Arrival-time order location revisited. *Bull. Seism. Soc. Amer.*, **99**, 2027-2034.
- Ružek, B., and Kvasnička, M. (2001). Differential evolution algorithm in earthquake hypocenter location. *Pure appl. geophys.*, **158**, 667-693.
- Saari, J. (1991). Automated phase picker and source location algorithm for local distances using a single three-component seismic station. *Tectonophysics*, **189**, 307-315.
- Sambridge, M. S., and Kennett, B. L. N. (1986). A novel method of hypocentre location. *Geophys. J. R. astr. Soc.*, **87**, 679-697.
- Sambridge, M. S., and Kennett, B. L. N. (2001). Seismic event location: nonlinear inversion using a neighbourhood algorithm. *Pure appl. Geophys.*, **158**, 241-257.

- Satriano, C., Lomax, A., and Zollo, A. (2008). Real-time evolutionary earthquake location for seismic early warning. *Bull. Seism. Soc. Amer.*, **98**, 1482-1494.
- Schöffel, H.-J. and Das, S. (1999). Fine details of the Wadati-Benioff zone under Indonesia and its geodynamic implications. *J. Geophys. Res.*, **104**, 13101-13114.
- Schwartz, S. Y., Dewey, J. W., and Lay, T. (1989). Influence of fault plane heterogeneity on the seismic behavior in southern Kurile Islands arc. *J. Geophys. Res.*, **94**, 5637-5649.
- Schweitzer, J. (2001). HYPOSAT – An enhanced routine to locate seismic events. *Pure and Appl. Geophys.*, **158**, 277-289.
- Schweitzer, J., and Storchak, D. A. (2006). Modernizing the ISC location procedures (editorial). *Phys. Earth Planet. Int.*, **158**, 1-3.
- Shearer, P. M. (1999). Introduction to seismology. Cambridge University Press, 260 pp.
- Smith, G. P., and Ekström, G. (1996). Improving teleseismic event locations using a three-dimensional Earth model. *Bull. Seism. Soc. Am.*, **86**, 3, 788-796.
- Spallarossa, D., Ferretti, G., Augliera, P., Bindi, D., and Cattaneo, M. (2001). Reliability of earthquake location procedures in heterogeneous areas: synthetic tests in the South Western Alps, Italy. *Phys. Earth Planet. Int.*, **123**, 247-266.
- Stein, S. (1991). Introduction to seismology, earthquakes and Earth structure. *Lecture notes*, Department of Geological sciences, Northwestern University, USA.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2003). The IASPEI Standard Seismic Phase List. *Seism. Res. Lett.*, **74**, 761-772.
- Thurber, C. H., and Engdahl, E. R. (2000). Advances in global seismic event location. In: in Thurber and Rabinowitz (2000), 3-22.
- Thurber, C. H., and Kissling, E. (2000). Advances in travel-time calculations for three-dimensional structures. In: Thurber and Rabinowitz (2000), 71-100.
- Thurber, C. H., and Rabinowitz, N. (Eds.) (2000). Advances in seismic event location. *Kluwer Academic Publishers*, 266 pp.
- Tsai, V. C., Nettles, M., Ekström, G., and Dziewonski, A. (2005). Multiple CMT source analysis of the 2004 Sumatra earthquake. *Geophysical Research Letters*, **32**; L17304, doi: 10.1029/2005GL023813.
- Wadati, K. (1933). On the travel time of earthquake waves. Part. II. *Geophys. Mag. (Tokyo)* **7**, 101-111.
- Waldhauser, F., and Ellsworth, W. L. (2000). A double-difference earthquake location algorithm: method and application to the northern Hayward fault, California. *Bull. Seism. Soc. Am.*, **90**, 1353-1368.
- Wiechert, E. (1907). Über Erdbebenwellen. Theoretisches über die Ausbreitung der Erdbebenwellen. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen, Mathematisch-physikalische Klasse*, 413-529.

| Topic   | Reports and bulletins                                                                                                                                                                                          |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Gernot Hartmann</b> , Bundesanstalt für Geowissenschaften und Rohstoffe, Stilleweg 2, 30655 Hannover, Phone: +49 (511) 643 3227; E-mail: <a href="mailto:Gernot.Hartmann@bgr.de">Gernot.Hartmann@bgr.de</a> |
| Version | March 2011; DOI: 10.2312/GFZ.NMSOP-2_IS_11.2                                                                                                                                                                   |

## 1 General information

The results of seismogram analyses are published in reports, event lists, bulletins and data catalogues (called "products"), which are the basis of data exchange between seismological institutions, data centers and for informing the public. The parameter data, primarily source and/or phase parameter values, should be stored in a digital database and presented in a clear and uniform manner.

The predefinition of an appropriate format depends on the requirements for the major usage of the product, for example whether a product is intended for further applications on a computer or for human readability. However, earthquake catalogues are published in numerous different formats. Therefore, a description of the individual used format should be given as a reference or directly attached to the product. Some formats have been developed to a comprehensive level, which is widely accepted in the seismological community. The IASPEI Seismic Format (ISF) was adopted as a standard format for exchanging seismic parameter data by the IASPEI Commission on Seismic Observation and Interpretation. The ISC as well as the NEIC and an increasing number of national data centers publish their catalogues in the ISF format, which also complies with the IMS1.0 format that was developed for the verification of the Comprehensive Nuclear Test Ban Treaty (CTBT). The ISF format description is available on the ISC-WEB site (<http://www.isc.ac.uk/standards/isf>).

A relatively new development is the format QuakeML. It was developed with special emphasis on real-time parameter exchange and in view of the need to frequently extend the format by data fields not contained in the standard itself. This is achieved by using XML as basis for QuakeML. More information on QuakeML can be found on the website <http://www.quakeml.org>

The products contain source and/or phase parameters. Products which include source parameters represent the seismicity within a given time period for a pre-defined area and above a reliable magnitude threshold. These limitations have to be taken into account, in order to provide a high quality product which claims to be of a high degree of completeness and accuracy.

Comprehensive products based on networks of stations distributed world-wide are published by the World Data Center (WDC) for Seismology (operated by NEIC, Denver, Colorado, USA) (<http://neic.usgs.gov>), the International Seismological Centre (ISC) (Newbury, England) (<http://www.isc.ac.uk>), and the International Data Centre of the Comprehensive Nuclear-Test-Ban Treaty Organisation (Vienna, Austria) (<http://www.ctbto.org>). The precision of these products with respect to epicenter location is  $\Delta D < 1^\circ$ .

In general, the products of national data centers and observatories provide data for events within an area that is well covered by the station network used for the data analysis. How complete these products are depends on the spacing between the stations, which also has a

major influence on the magnitude threshold. An example of such products is given by the German local bulletin of the Federal Institute for Geosciences and Natural Resources (BGR) at the following internet address: <http://www.bgr.bund.de/erdbebenkatalog-deutschland>. The precision of this product is  $\Delta D < 0.1^\circ$ .

Publication of teleseismic epicenter data from a regional network is useful only if the precision of the data is known or reliable calibration values for the correction of systematic location errors are available. Such a calibration for the GRF array/GRSN (see Section 8.7.5 and Fig. 8.17 in Chapter 8) is used by the BGR for determinations of epicenters world-wide in its teleseismic bulletin. The precision in the distance range  $D = 13^\circ\text{--}100^\circ$  is  $\Delta D < 3^\circ$  (see [http://www.bgr.bund.de/DE/Themen/Erdbeben-Gefahrungsanalysen/Seismologie/Seismologie/Erdbebenauswertung/Erdbebenkataloge/Teleseism\\_Bulletins/teles\\_bulletins\\_node.html](http://www.bgr.bund.de/DE/Themen/Erdbeben-Gefahrungsanalysen/Seismologie/Seismologie/Erdbebenauswertung/Erdbebenkataloge/Teleseism_Bulletins/teles_bulletins_node.html)).

However, if only phase parameters are available, it is also important for these to be reported to the international data centers, because these values are indispensable to improve the accuracy and comprehensiveness of their products. The ISC, for example, compiles a data catalogue based on reported phase parameter values received from a number of observatories and data centers world-wide (<http://www.isc.ac.uk/doc/analysis/2007p01/datacol.html>).

The information provided by a product also depends on how long after the event the product is to be published. For example:

## 2 Fast determination of epicenters of strong earthquakes

Information is provided for a single event immediately after it is detected and recognized as a strong earthquake or an earthquake which could cause substantial damage. The parameter values are obtained by automatic processing or manual analysis. They are published within minutes or hours after the event and are distributed mainly by e-mail or made available on the WWW. The NEIC publishes a near-real-time earthquake bulletin with about 20 recent earthquakes, which is updated every 5 minutes (<ftp://hazards.cr.usgs.gov/cnss/quake>). More detailed information is provided for the latest earthquakes in the world on the website [http://earthquake.usgs.gov/earthquakes/recenteqsww/Quakes/quakes\\_all.php](http://earthquake.usgs.gov/earthquakes/recenteqsww/Quakes/quakes_all.php). This website contains information for earthquakes of the last 7 days. A global, near-real-time earthquake bulletin is also produced by the GFZ Potsdam via the website <http://geofon.gfz-potsdam.de/eqinfo> where automatically produced earthquake information is published within minutes for significant events.

For local and/or regional purposes in Central Europe, fast epicenter determinations are also provided by the European Mediterranean Seismological Centre (EMSC) (<http://www.emsc-csem.org/Earthquake/index.php>). Epicenter determinations, which are received from national data centers, are automatically associated to the corresponding earthquake and published on the website for latest data contributions (<http://www.emsc-csem.org/Earthquake/seismologist.php>).

## 3 Preliminary products

Information is provided for all routinely analyzed events at regular time intervals, typically daily, weekly or monthly. This information may be subject to modification if phase readings from additional stations or arrival times of later phases are identified at a later stage of the

analysis or new data is received from contributing observatories. NEIC publishes, for example, their preliminary products on a weekly and monthly basis (<ftp://hazards.cr.usgs.gov/weekly> and <ftp://hazards.cr.usgs.gov/edr> , respectively). The BGR produces a preliminary event list of local, regional and world-wide seismic events with a time delay of 1 to 3 days. It is published on the BGR-website ([http://www.bgr.bund.de/DE/Themen/Erdbeben-Gefahrungsanalysen/Seismologie/Seismologie/Erdbebenauswertung/Detaillierte\\_Auswertung/Daten\\_GRS\\_GRF/grf\\_grsn\\_node.html](http://www.bgr.bund.de/DE/Themen/Erdbeben-Gefahrungsanalysen/Seismologie/Seismologie/Erdbebenauswertung/Detaillierte_Auswertung/Daten_GRS_GRF/grf_grsn_node.html)). Revised German products are the monthly distributed German local bulletin and the regional and teleseismic bulletin as described in section 1. All German products are based on the GERESS and GRF arrays, GRSN, and GEOFON networks. Additionally, seismic stations of neighboring countries, which provide online data access, are used for a precise determination of earthquakes in the border areas of the German station network. The results of local seismological services and observatories with their dense local station networks are also considered, in order to provide the German local bulletin with the best possible earthquake localizations.

## 4 Final products

The most complete and precise data on seismic events is published when all of the available data has been analysed. These products may be published up to several months after the events. ISC, for example, distributes their final products (<http://www.isc.ac.uk/doc/products/index.html>) about two years after the earthquakes occurred. The final ISC bulletin can be downloaded from the ISC website (<ftp://ftp.isc.ac.uk/pub/isf>) or can be requested at the ISC on compact disc. This bulletin comprises the most comprehensive information on global earthquakes including the final results of various national data centers and forms the base for scientific studies at universities and research facilities.

National as well as international data centers make their earthquake catalogues and bulletins available to the public on their WEB sites. In most cases it is also possible to search online for earthquake information in the data bases of the data centers. Requests can be individually configured for selected spatio-temporal criteria. This service is, for example, provided at the ISC website <http://www.isc.ac.uk/search/index.html>.

### Additional references

ORFEUS software library, <http://orfeus.knmi.nl>  
IRIS, <http://www.iris.washington.edu>  
RENASS, <http://renass.u-strasbg.fr/>  
SED, <http://www.seismo.ethz.ch/index>  
GEOFON: <http://geofon.gfz-potsdam.de>

### Acknowledgement

Thanks of the editor go to Joachim Saul of the GFZ German Research Centre for Geosciences, Potsdam, who has reviewed the draft and proposed some valuable amendments.



| Topic   | Animation of seismic ray propagation and seismogram formation                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p><b>Siegfried Wendt</b>, Geophysical Observatory Collm, University of Leipzig, 04779 Wermsdorf, Germany;<br/>E-mail: <a href="mailto:wendt@rz.uni-leipzig.de">wendt@rz.uni-leipzig.de</a></p> <p><b>Ute Starke</b>, Computer Center, University of Leipzig, Augustusplatz 10-11, 04109 Leipzig; E-mail: <a href="mailto:starke@rz.uni-leipzig.de">starke@rz.uni-leipzig.de</a></p> <p><b>Peter Bormann</b> (formerly GeoForschungsZentrum Potsdam, Telegrafenberg, 14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> |
| Version | October, 2002; DOI: 10.2312/GFZ.NMSOP-2_IS_11.3                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |

## 1 Introduction

Presented are 10 movies with animations of seismic ray propagation and the formation of seismic recordings in the distance range from  $0.1^\circ$  to  $167^\circ$ . Below we comment on each of these movies, which complement record sections presented in the Chapters 2 and 11 of this Manual. The animations can be opened by right mouse click on the file name **printed in bold blue letters** and then follow the instructions given below in section 2. The animation files can also be downloaded via the summary listing *Download Programs & Files* (see Overview on the NMSOP-2 cover page).

The development of such movies began at the University of Leipzig more than 20 years ago. Dr. Bernd Tittel, now retired from the University's Geophysical Observatory Collm, was an expert in identifying late and very late core phases. Therefore, the idea was born to produce an animation of the propagation of the seismic rays for such very late phases through a simple standard 1-D Earth model. The related seismic recordings of station Collm (CLL) were later added as standing pictures at the end of the movie. More than a decade ago one of us (S. Wendt) revived these earlier efforts by extending these animations to crustal, mantle and core phases in the local, regional and teleseismic distance range and by "writing" the sequence of records at several stations of local or regional networks as soon as the respective seismic phase rays arrive at these stations.

All ray paths and travel-time curves were calculated with the program TauP of H. P. Crotwell (1998) on the basis of the IASP91 Earth model (Kennett and Engdahl, 1991).

## 2 Software requirements and handling

The animations run on a Windows-PC with QuickTime Player 4 to 7. Quick Time Player can be downloaded from [www.apple.com](http://www.apple.com). After its installation, activate the event record file you wish to see in *Download Programs & Files*. Move with the cursor in front of the bold blue file name you wish to open. A right-mouse click opens a window which offers also the choices "copy hyperlink" and "open hyperlink". When you confirm with a left mouse click "open hyperlink" then - after a while - the cover page of the respective movie file appears in a frame. Start the movie by pressing the start-button  in the middle of the lower frame ridge. If this button is not visible then you have to reduce the frame size in the Quick Time Player menu line which appears above the movie cover frame. Click "Display Size" in order to adopt the movie frame to the size of your screen and then start the movie. The moving marker above

the lower command line will start running and the movie will show up. If you wish to hold-on the movie or control its speed at will then just catch the moving marker with your cursor and move it as you wish.

If lines, letters and numbers in the movie appear broken or blurred, the screen resolution has to be increased before starting QuickTime. Click “Start” in the desktop area, then “Settings”, and then “Display” on the appearing “Control Panel”. When “Display Properties” appears click “Settings” and there you may change the pixel number in the desktop area. Usually, a resolution of 1280x1024 pixels should be sufficient to resolve all details. Then return to the desk top, go back to *Download Programs & Files* and start the movie file again or select another such demonstration.

Yet, sometimes a much easier way works even when the film is already running. Go to “Display Size” and change there from “adapt to screen size” to “normal size”.

### 3 General structure and contents of the movies

Each movie file shows first a cover page in the start-frame, which gives the data, origin time, source co-ordinates and name of the source region of the earthquake. After starting the movie you will see, for all teleseismic earthquakes, the propagation of seismic rays of different phases, coded in different color, through a simplified cross section of the Earth. Shown are only those rays which will reach stations of the German Regional Seismic Network (GRSN). As soon as the different rays arrive at these stations, the onsets and associated waveforms will be “written” according to the original seismograms recorded at these stations from the considered earthquake. The response filters applied to the original velocity broadband records are given together with the record component in the record frames. The station codes and the epicentral distances to the stations are written on the ordinates and the travel times are marked on the abscissas of the record section frames. For all depicted phases, the theoretical travel-time curves according to the IASP91 Earth and travel-time model are also plotted in the respective phase color into the record frames.

Exceptions are the two local earthquake recordings (Files 1 and 2). Here the propagation of the Pg and Sg wavefronts and the formation of the related records at stations of the GRSN in different azimuths from the source are either shown in conjunction with the formation of the local travel-time curve (File 1) or on separate map projections.

### 4 Movie files and peculiarities of the recordings shown

Below the following information is given for each animation:

- file name;
- geographical region of epicenter;
- source parameters date, origin time OT, co-ordinates, source depth h;
- magnitude and data center which has provided these source data;
- the size of each animation file in megabyte (MB);
- the interval of epicentral distances D of the recording stations shown;
- remarks about the depicted seismic phases and recording frames;

- complementary remarks about the peculiarities of the records shown and the lessons learned from this demonstration.

For the definition of seismic phase names and complementary ray diagrams see IS 2.1. The original velocity broadband records have been filtered according to the response characteristics of standard seismographs, e.g., of type Kirnos SKD, SRO-LP, WWSSN-SP and WWSSN-LP. Their response curves are shown in Chapter 11, section 11.3.2 together with complementary record examples.

Note, that in the following the file names are printed in bold black, since they are not hyperlinked with IS 11.3.

**File 1: [wendt\\_vogtland1\\_qt](#)**

NW-Bohemia, Czech Republic, Novy Kostel

17.09.2000 OT=15:14:33.5 50.22N 12.47E h=10km Ml=3.1 (SZGRF)

23.0 MB

D=10–130km

Pg and Sg

**Left:** propagating wave fronts of Pg (blue) and Sg (red); **right:** records of some stations of the GRSN. Traces are sorted according to distance. At the moment of wave-front arrival at a station the onset and related waveform of this arrival is written in the record of the seismic station.

Note that the depicted travel-time curves for an average 1-D local crustal model match well with the onsets at the stations WERN, MOX and CLL but the onsets recorded at the stations GFRO and WET are about 2 to 4 s later than the onset times expected according to this model. This illustrates the need for improved local travel-time curves which may also be azimuth-dependent.

**File 2: [wendt\\_vogtland2\\_qt](#)**

NW-Bohemia, Czech Republic, Novy Kostel

04.09.2000 OT=00:31:45.2 50.21N;12.44E h=10km Ml=3.2 (BGR)

28.3 MB

D = 10–240km

Phases: Pg and Sg

**Left:** developing records at several GRSN-stations; **right:** propagating wave fronts of Pg (blue) and Sg (red).

When a wave front arrives at a station the respective onset and original waveform recorded at this station is written in the seismogram. The amplitude ratio Pg/Sg strongly varies with azimuth due to the different source radiation pattern for P and S waves with respect to the station azimuth (see Figs. 3.25 and 3.26).

**File 3: [wendt\\_slovenia\\_qt](#)**

Slovenia

12.07.2004 OT=13:04:06.3 46.30N; 13.64D h=2km mb=5.0 (NEIC)

190.2 MB

D = 80-910km

Phases: Pn, (Pb?), Pg, Sn, Sb and Sg

Presented are the Z-component (above) and EW-component records (below) of WWSSN-SP filtered records of 8 stations of the German Regional Seismic Network (GRSN) (see diagram upper right) and a schematic crustal cross section (lower right) with the assumed ray-paths of the standard crustal phases Pg, Pb, Pn, Sg, Sb, and Sn. The following should be noted:

- 1.) Up to about 780km distance Pg is here the dominant phase onset in the early parts of the Z-component records;
- 2.) Yet, at the most distant station RGN Pn amplitudes are clearly larger than that of Pg which fall already in the signal-generate noise level;
- 3.) Pb, the critically refracted P wave from the mid-crustal Conrad discontinuity is not recognizable clearly in any record with the exception of the station RGN Z-component record, where it may be associated with slightly earlier than predicted arrival of a wave group with increased amplitude but no sharp onset;
- 4.) The P-wave onsets in the horizontal component records are generally smaller than in the Z component.
- 5.) Sg is the onset of the most prominent wave group with the largest amplitudes in the whole event record train;
- 6.) Sn is a clear earlier shear-wave arrival prior to Sg only in the horizontal component records and there also Sb has occasionally recognizably increased amplitudes, in this case at records of stations GRFO and CLL even with distinct onsets.

However, in the majority of earthquake records Pb and Sb are not well developed with clear onsets and significant SNR above the signal-generated noise between Pn and Sg.

**File 4: [wendt\\_hindukush\\_qt](#)**

Afghanistan-Tajikistan border region

30.06.1994 OT=09:23:21.4 36.3N;71.1E h=227km mb=6.1 (NEIC)

54.0 MB

D=43° - 47°

Phases: P, pP, sP, PcP, PP, pPP, PPP, sPP, ScP

Strong phases P, sP and sPP on KIRNOS displacement BB Z-component records (top traces) of a deep Hindukush earthquake, whereas the reflections at the core-mantle boundary PcP and ScP are more clear in WWSSN-SP-filtered traces (bottom traces). The travel-time curves of PcP and PP intersect near 43° epicentral distance. This overlap distance is depth-dependent. Also note the simple impulsive P-wave onset from this deep source as well as the different amplitudes of the depths phases pP and pPP on the one hand and the much stronger depth phases sP and sPP on the other hand. This is due to the different P- and S-wave radiation pattern in the direction of the short up-going rays p and s, respectively (see Fig. 2.43). Stations situated in another azimuth from the same source may observe a different amplitude ratio of these depth phases.

**File 5: [wendt\\_india\\_qt](#)**

W- India

26.01.2001 OT=03:16:40.7 23.3N;70.3E h=22km Ms=7.9 (NEIC)

63.5 MB

D=51.5°-55.5°

Phases: P, PP, S, ScS, SS, PKPPKPdf, and PKKKKP (P4KP)

**Top:** vertical (Z) component records; and middle: transverse (T) component records; both SRO-LP filtered; **bottom:** zoomed windows with Z-component short-period filtered records (4<sup>th</sup> order band-pass, 0.5 – 1.7 Hz) of the multiple reflected core phases PKPPKPdf (P'P') and

PKKKKP (P4KP), which have travel times of more than 39 min and 46 min, respectively. Note that PKPPKP arrives from the opposite azimuth as compared to the direct P wave and has a negative slowness (i.e., the travel-time to stations at larger D is shorter).

**File 6: [wendt\\_russlchina\\_qt](#)**

E-Russia-NE-China border region

28.06.2002 OT=17:19:30.2 43.8N;130.7E h=564km Mw=7.3 (NEIC)

90.1 MB

D=68.5°-75.5°

Phases: P, pP, PP, pPP, S, sS, ScS, SS, SSS

Vertical (Z)- and transverse (T)- component records with SRO-LP-simulation. This deep earthquake produced strong body waves with simple waveforms, including clear depth phases and strong, transversely polarized S waves. Surface waves are absent. The clear records of very late core phases PKPPKP, SKPPKP, and SKPPKPPKP in WWSSN-SP filtered records have not been depicted in this animation.

**File 7: [wendt\\_peru\\_qt](#)**

N- Peru

02.05.1995 OT=06:06:05.7 3.8S;76.9W h=97km mb=6.5 (NEIC)

27.8 MB

D=90°-93°

Phases: P, PKKPbc, PKPPKPdf (P'P'df), PKPPKPPKPbc (3P'bc)

Shown are four short record windows, each one minute long, for P and pP as well as the multiple core phases PKKPbc, PKPPKPdf, and PKPPKPPKPbc with their respective depth phases. These late and very late core phases have travel-times of about 30, 38, and 59 min, respectively. Despite their long travel paths through the Earth, these late core phases still have an astonishingly good signal-to-noise ratio at most stations. They may easily be misinterpreted at individual stations as P-wave onsets from other independent events. Note the negative slowness of the phases PKKP and P'P'.

**File 8: [wendt\\_newbritain\\_qt](#)**

New Britain region

10.05.1999 OT=20:33:02.1 5.2S;150.9E h=138km mb=6.5 (NEIC)

25.6 MB

D=122°-126°

Phases: Pdiff, PKPpdf, PP, PPP, PS, PPS, SS, SSS, LQ, LR, 4PKPbc

The rarely observed phase 4PKPbc with a travel time of about 79 min is recorded with good SNR in the WWSSN-SP filtered traces (sampling rate of 20Hz). The projection of its path on the surface is  $(2 \times 360 - 124) \text{ deg} = 596 \text{ deg}$ . The long-period diffracted P-wave arrival Pdif (old name Pdiff) is well developed and arrives about 3.5 min ahead of PKPpdf in the SRO-LP filtered Z-component record (see record window in the lower left corner; sampling rate 1 Hz). The SRO-LP filtered Z-component (top left window) and T-component records (middle left window) show several mantle phases, a sharp onset of Love waves (LQ) in T and the onset of the Rayleigh wave LR in Z. The latter shows clear normal dispersion with rather long-period

waves ( $T \approx 1$  min) and large amplitudes at the beginning whereas shorter periods have much smaller amplitudes due to the intermediate source depth of this earthquake.

**File 9: [wendt\\_fiji\\_qt](#)**

Fiji Islands region

18.12.2000 OT=01:19:18.6 21.3S;179.2W h=600km mb=6.4 (NEIC)

35.9 MB

D=147°-152°

Phases: PKP<sub>df</sub>, PKP<sub>bc</sub>, PKP<sub>ab</sub>, PP, PPP, PPS, SS, sSS, sSSS, PKKKKKP (P5KP)

Shown are SRO-LP filtered vertical (Z)- and transverse (T)- component records (upper and middle record window) as well as zoomed record windows in the lower left and middle with WWSSN-SP filtered Z-component records of the beginning (PKP-wave group) and the very late part of the seismogram with the rare phase.

Note the good match of the actual and theoretically expected travel-time onsets according to the IASP91 Earth model for mantle phases in the upper and middle record window. In contrast, this model does not predict well the onsets of the core phases which arrive about 5 to 15 seconds later than expected by this model.

**File 10: [wendt\\_nz\\_qt](#)**

East of North Island, New Zealand

21.08.2001 OT=06:52:06.7 37.0S;179.8W h=33km mb=6.5 (NEIC)

52.0 MB

D=160°-167°

Phases: PKP<sub>df</sub>, PKP<sub>ab</sub>, PP, PP2, PcPPKP, PPP2, SS, SSSS

The SRO-LP-filtered seismograms of this very distant event show remarkably strong PP2 and PPP2 onsets, which arrived at the stations from the opposite backazimuth over the long ray paths ( $360^\circ - D$ ). On the inserted map of Central Europe, which also shows the position of the stations of the GRSN, the passing of these wave fronts can be watched: The phases PKP, PP, PPP, SS and SSSS approach from the NE whereas the phases PcPPKP, PP2 and PPP2 approach from the SW. Also note that the wave fronts of the first arriving waves with rather small (steep) incidence angles have a very high apparent horizontal speed of wave propagation whereas the wave fronts of the later phases with large (shallow) incidence angles travel much slower through the network of seismic stations. Also note the well developed and very long surface-wave train from this shallow (crustal) earthquake. The surface waves arrive for this very distant earthquake more than one hour after the PKP first arrival.

## References

- Crotwell, H. P., Owens, T. J.; Ritsema, J. (1999). The TauP toolkit: Flexible seismic travel-time and ray-path. *Seism. Res. Lett.* **70**, 154–160.
- Kennett, B. L. N., and Engdahl, E. R. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, **105**, 429-465.

| Topic   | Tutorial for consistent phase picking<br>at local to regional distances                                                                                                                                                                                                                                                                                                                                                                                                                                             |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <p><b>Tobias Diehl</b>, Lamont-Doherty Earth Observatory, Columbia University, Palisades, New York, USA; now: Swiss Seismological Service, ETH Zurich, Sonneggstrasse 5, 8092 Zurich, Switzerland, E-mail: tobias.diehl@sed.ethz.ch</p> <p><b>Edi Kissling, ETH Zurich</b>, Institute of Geophysics, Sonneggstr. 5, 8092 Zurich, Switzerland, E-mail: kiss@tomo.ig.erdw.ethz.ch</p> <p><b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany); E-mail: pb65@gmx.net</p> |
| Version | May 2011; DOI: 10.2312/GFZ.NMSOP-2 IS 11.4                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |

|                                                                          | page      |
|--------------------------------------------------------------------------|-----------|
| <b>1 Introduction</b>                                                    | <b>1</b>  |
| <b>2 Routine hand picking</b>                                            | <b>2</b>  |
| 2.1 Phase Picking Part I: Phase timing and its error assessment          | 3         |
| 2.2 Phase Picking Part I: Phase identification and its error assessment  | 6         |
| 2.3 Phase Picking Part I: First motion polarity and its error assessment | 9         |
| 2.4 Size of time window and amplitude scaling                            | 9         |
| 2.5 Aliasing and waveform filtering                                      | 11        |
| 2.6 Additional comments for S-wave picking                               | 13        |
| 2.6.1 Polarization of the wavefield                                      | 13        |
| <b>3 Consistent hand picking procedure</b>                               | <b>16</b> |
| 3.1 Consistent waveform filtering                                        | 16        |
| 3.2 Consistent window size and amplitude scaling                         | 17        |
| 3.3 Consistent phase identification                                      | 20        |
| <b>Acknowledgment</b>                                                    | <b>20</b> |
| <b>References</b>                                                        | <b>20</b> |

## 1 Introduction

Modern digital acquisition systems (broadband sensors, timing with GPS or DCF-systems, sampling rates  $\geq 80$  Hz) and the use of modern analyzing software allow arrival-time determination of seismic signals with a precision of up to a few tenth of milliseconds. To benefit from this advancement in travel-time based methods like earthquake location and seismic tomography, consistent arrival time determination and error assessment (including uncertainty estimates of timing and phase interpretation) is crucial. Uncertainties of the arrival-time determination have to be provided in order to weight the data properly in the inversion procedure, i.e. arrival-time picks of high quality contribute more to the solution than those of low quality. The impact of such consistently picked and weighted data on tomographic models has recently been shown by several studies (e.g. Di Stefano et al. 2006, Diehl et al. 2009, Husen et al. 2009).

Although seismic arrival times represent the fundamental data of many seismological applications, the description of consistent phase picking receives only little attention. Simon (1981) and Kulháněk (1990 and 2002) provide a general overview about seismogram

interpretation from local to teleseismic scales. They focus mainly on basic descriptions of phases observable in common seismograms. The assessment of timing uncertainty and phase interpretation, however, is barely discussed. The fundamentals of digital signal processing and their influence on onset properties are described, e.g., in Seidl and Stammer (1984), Scherbaum (2001), and Scherbaum (2002). Among the few recent guidelines, the first edition of the *New Manual of Seismological Observatory Practice* (NMSOP) provides in Chapter 11 (Bormann et al., 2002) already an introduction to basic picking principles for local, regional and teleseismic seismograms, proposing the quantification of the onset-time reliability, yet a detailed description for consistent quality assessment is missing, similar to the discussion in Scherbaum (2001).

This tutorial is intended to provide a more detailed description on seismic phase picking. Basic principles (including a consistent quality assessment for timing uncertainty and phase interpretation) and common pitfalls of manual picking are described in section 2. In section 3, we present a practical procedure for consistent hand picking. The proposed method is especially intended for picking a reference data set for testing and calibrating of automatic picking algorithms, as described in Chapter 16 of NMSOP-2. The tutorial is focused on crustal and upper mantle phases of local and regional seismic events, recorded at epicentral distances from a few 10 km to several 100 km. But the outlined principles apply likewise to teleseismic records, however, then for different time and frequency resolution and recording bandwidth.

## 2 Routine hand picking

Especially for local and regional earthquake data, phase picking often becomes inconsistent and ambiguous due to rather complex waveform patterns and the close arrival of different phase types within a short time period of the same coda. In general, the shape of a seismic wavelet is affected by the source time function, the radiation pattern, dispersion, attenuation, scattering, interference with other phases, the signal-to-noise ratio (SNR) at the recording site, and the response characteristic of the recording system. The superposition of these components may lead to highly complex waveforms in case of high-frequency local and regional recordings. Moreover, the picking of later phases such as PmP or S-waves becomes even more difficult, since they interfere with the coda of earlier phases. Since the target accuracy of such arrivals is usually within few hundred milliseconds, errors like phase misinterpretations denote blunders and may significantly bias hypocenter locations and velocity models derived from onset-time observations. Furthermore, regional studies usually require the compilation of data from several networks. In such a case, merging routinely measured phase data will add even more inconsistency.

The possible sources for inconsistent hand picking described in the following sections refer mainly to P-wave picking. However, the basic principles of picking and error assessment of phases are also valid for S-wave picking. In addition, section 2.6 includes problems specific to S-wave picking.

In general, a seismic phase is defined by two visual observations:

1. Change primarily in amplitude:

The amplitude exceeds the background noise (for a certain amount of time). This can also be denoted as amplitude based signal-to-noise ratio (*ASNR*). We may define a phase, if its amplitude exceeds the background noise at least by a factor of 1.5 (*ASNR*



$\geq 1.5$ ). Figure 1 represents a typical example for a phase arrival defined by a change of the *ASNR*.

2. Change primarily in frequency:

A change of the dominant frequency indicates the arrival of a seismic phase. We can refer to this observation as frequency based signal-to-noise ratio (*FSNR*). Unlike the *ASNR*, it is often much more difficult to quantify visually. Moreover, the dominant frequency of noise and signal can sometimes be very similar. But especially for broad time windows, the *FSNR* can help to determine the approximate position of a phase. Figure 2 represents a typical example for a phase arrival defined by a change of the *FSNR*.

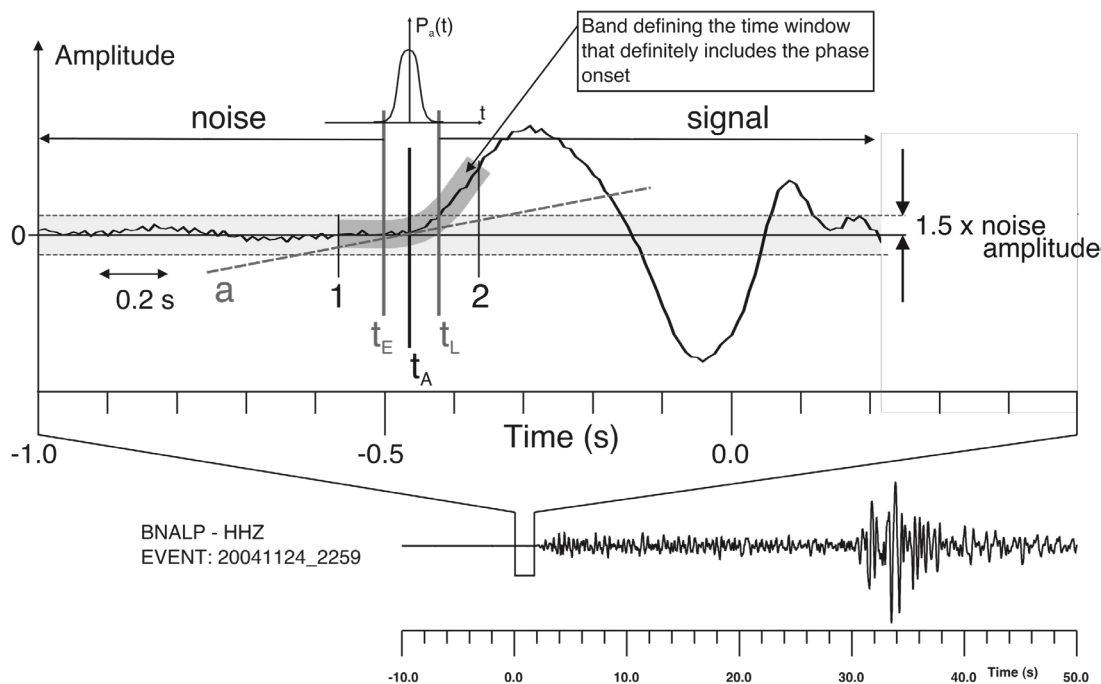
Once a phase is recognised based on *ASNR* or on *FSNR* or - most likely - a combination of both, we have to determine the precise wavelet onset ('arrival time'). However, its position is usually not completely independent from the associated overall observation uncertainty. Finally, we have to identify the phase type and to do so, we have to answer questions like: May we assign the ray path to the phase (Pg, Pn, PmP) with certainty or is the station near a triplication point (see Chapter 2, Figs. 2.29 and 2.40)? Are we sure that it is the first arrival or is it possible, that an earlier phase is hidden in the noise? In addition, the visual examination of *ASNR* and *FSNR* can be rather subjective and strongly depends on the width of the used time window and also on amplitude scaling. Finally, filtering can affect *ASNR* and *FSNR* significantly. In the following sections, we will give examples for these problems and suggest possible solutions in order to minimize inconsistencies during the picking process. Once identified, it should be possible to minimize their contribution by defining certain rules and procedures, which finally add up to a consistent picking workflow and consistent uncertainty assessment for each observation.

## 2.1 Phase Picking Part I: Phase timing and its error assessment

The basic quantities associated with a picked phase are usually the absolute arrival time and the corresponding observation error. However, it is difficult or even impossible to give a general definition of a seismic onset, which could be used for the actual measurement of first arrival time from a sampled band-limited signal in the presence of noise (Seidl and Stammer 1984). Therefore, the visual determination of absolute arrival times often implies a high degree of subjectivity and inconsistency. Consequently, a physical consistent formulation can only be achieved by a probabilistic point of view. Such an approach directly relates the measured arrival time with the corresponding observation error. Considering the onset of a seismic phase as probabilistic function  $P_a(t)$ , the arrival time is expressed as the "most likely" time  $t_A$ , with  $P_a(t_A) = \text{Max}(P_a)$ . On the other hand, the 'earliest' possible time for the phase onset is defined as  $t_E$ , where the likelihood for onset is approaching zero. Thus  $P_a(t_E) \geq 0$ . Similarly, the 'latest' possible time for the phase onset  $t_L$ , is defined as  $P_a(t_L) \geq 0$ .

Figure 1 illustrates the proposed concept for a typical *ASNR*-case in further detail. Although the onset of the phase is rather impulsive and exhibits an almost ideal SNR, it is difficult to determine an arrival time consistent with picks of waveforms from the same seismic source recorded at other stations. The thick grey band between position '1' and '2' defines the time window that for certain includes the onset of the wavelet. The band outlined by two broken lines denotes the possible threshold of the noise amplitude. In practice, we first determine the position of  $t_L$  and  $t_E$ . For a consistent determination of  $t_L$  and  $t_E$ , we have to setup a common

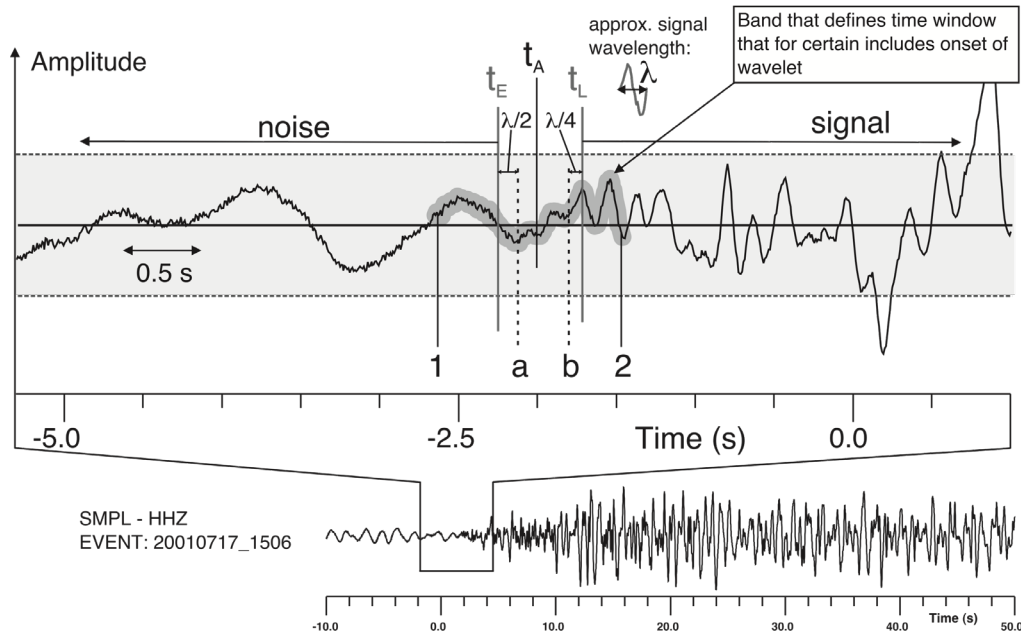
procedure. Since the amplitude exceeds the threshold several times at position ‘2’, the end of the grey band is certainly too late to be picked as  $t_L$ . Therefore, we define the intercept between signal amplitude and the a priori noise threshold (here: 1.5 times the largest noise amplitude in the noise-measurement time window) as  $t_L$ . Usually, the consistent determination of  $t_E$  is more difficult. In Figure 2 we fit a tangent (dashed line ‘a’) to the smoothed slope of the onset. If we shift the tangent from  $t_L$  towards earlier times, the slope decreases. The earliest possible time  $t_E$  corresponds with the first zero slope from  $t_L$  towards earlier time. Therefore, the start of the grey band (position ‘1’) is certainly too early and on the other hand,  $t_A$  would be too late to be picked as  $t_E$ . To ensure  $t_E$  includes the zero slope time in the presence of larger background noise, we could shift it to earlier arrival by approximately half a wavelength of the dominant noise. Subsequently, we pick the arrival of the phase at the most likely position  $t_A$ , within the error interval of  $t_E$  and  $t_L$  (e.g. on the seismogram’s leading edge). For the special case of an ideal delta-pulse,  $t_E$  and  $t_L$  would coincide with  $t_A$ .



**Figure 1** Probabilistic phase picking approach based on change of *ASNR*: the ‘earliest’ possible pick corresponds to  $t_E$ , the ‘latest’ possible pick corresponds to  $t_L$ . The most likely arrival time  $t_A$  is located within this interval. Primarily amplitude is used for determination of  $t_E$  and  $t_L$ . Copy of Fig. 2 in Diehl et al. (2009, p. 544) with © granted by the Geophysical Journal International.

Figure 2 illustrates a possible concept for the *FSNR*-case. Again, the thick grey band between position ‘1’ and ‘2’ defines the time window that for certain includes the onset of the wavelet and the band outlined by two broken lines denotes the possible threshold of the noise amplitude. The change in frequency is obvious for arrival times greater then position ‘b’. Therefore, we define  $t_L$  a quarter signal wavelength ( $\lambda/4$ ) after position ‘b’. For arrival times earlier than position ‘a’, the signal frequency, which dominates the grey-banded wavelet after position b, is certainly no longer recognizable. Therefore, we pick  $t_E$  half a signal wavelength ( $\lambda/2$ ) before position ‘a’.

The consistent determination of  $t_A$  in the ASNR as well as in the FSNR case can be difficult and might include some degree of subjectivity in practice. Although  $t_A$  in Figure 1 and 2 seems to be located halfway between  $t_E$  and  $t_L$ , a symmetrical distribution is not always appropriate and in some cases, a highly asymmetrical distribution is required (especially for broadband records). There is no universal definition for determination of  $t_A$ , but its position usually coincides with a prominent kink or discontinuity in the waveform, separating the noise from the signal. The position of  $t_E$  should include all possible earlier onsets of the picked phase. In that sense the utmost precision of  $t_A$  is less crucial as long as it is included in an appropriate uncertainty interval.

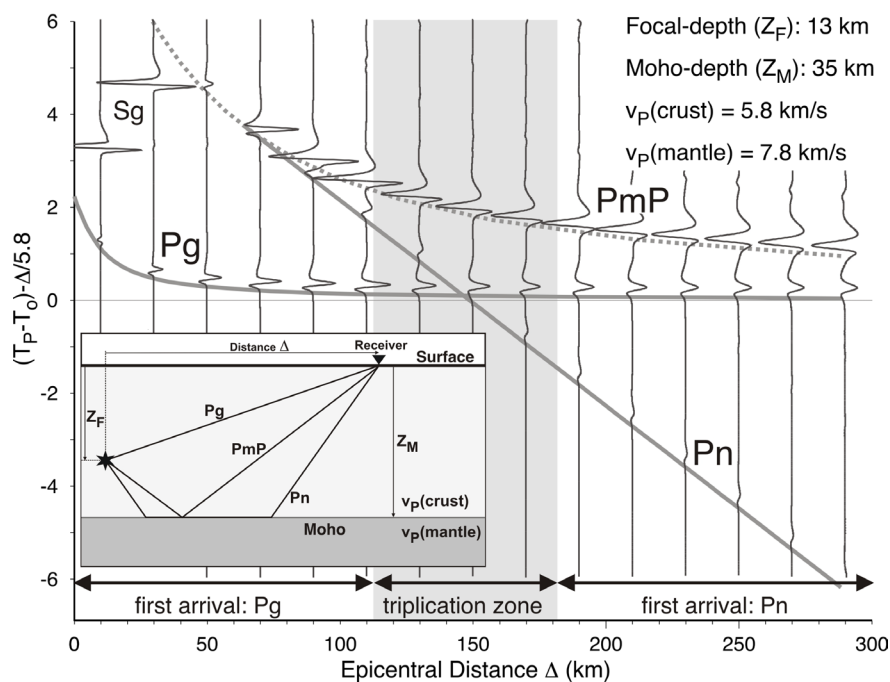


**Figure 2** Probabilistic phase picking approach based on change of  $FSNR$ : the ‘earliest’ possible pick corresponds to  $t_E$ , the ‘latest’ possible pick corresponds to  $t_L$ . The most likely arrival time  $t_A$  is located within this interval. Primarily frequency is used for determination of  $t_E$  and  $t_L$ .

Usually, the observation error is related with a discrete weighting or quality class (e.g. 0-5). However, these classes are often traditionally defined only through qualitative attributes, like waveform shape (impulsive or emergent). Such classification is not longer sufficient for a consistent error assessment. Instead, a quantitative weighting scheme has to be defined, which assigns the weighting class depending only on the measured time error interval  $t_L - t_E$ . In addition, the measurement of  $t_L$ ,  $t_E$ , and  $t_A$  allows us to adjust the weighting class definition even after the picking process. Finally, the availability of uncertainty intervals offers the possibility to assess the performance of automatic picking algorithms in a quantitative sense. Phase picks of automatic algorithms are expected to fall within uncertainty intervals of the corresponding manual picks. For more details on test and calibration of automatic algorithms see Chapter 16.

## 2.2 Phase Picking Part II: Phase identification and its error assessment

Although phase misinterpretation can result in significantly large errors, phase identification is typically not supplied with any observation error or uncertainty attribute at all (unlike arrival time of a phase). As demonstrated by synthetic travel time curves in Figure 3, the first arrival of the local or regional P-wave prior to its related coda is usually either Pg (direct wave as illustrated in the inset of Fig. 3) or Pn (Moho-refracted wave). Whether Pg or Pn is the first arriving wave depends on epicentral distance  $\Delta$ , crustal thickness  $z_M$ , and focal depth  $z_F$ . As a rule of thumb, Pn from a surface focus takes over Pg at  $\Delta \approx 5 z_M$ , for deeper sources, however, at shorter distances. Also Moho tilt and the direction of observation with respect to the direction of Moho dip influence both the slope of the Pn travel-time curve and the “take-over” distance (for both effects see Chapter 2, Fig. 2.40).



**Figure 3** Reflectivity seismograms (vertical components) and synthetic travel time curves (solid and dashed lines) for main phases observed in local and regional earthquakes (one layer over half space). Focal depth is set to 13 km and time axis is reduced by 5.8 km/s. Phase identification is expected to be difficult in the distance range of phase triplication. The position and width of this zone mainly depends on focal depth and Moho topography.

Moreover, the discrimination between Pg and Pn can become rather difficult in the distance range of phase triplication (see Fig. 3 as well as Figs. 2.28b, 2.29 and 2.40 with related texts in Chapter 2). The Moho-reflected PmP phase always arrives after Pg and Pn, although its amplitude can become the dominant phase in the coda. The synthetic reflectivity seismograms (Fuchs and Müller, 1971) in Figure 3 illustrate the theoretically expected amplitude ratios between Pg, Pn, and PmP for a simplified crustal model using a moment tensor as a source.

According to it, especially for large epicentral distances, Pn can be expected to be masked in the presence of noise and Pg or PmP then likely to be misinterpreted as the first arrival.

However, Figure 4 presents a velocity-reduced real record section and the related travel-time curves for Pg, Pn and Pm from a local earthquake near Walenstadt, Switzerland with a focal depth of 13 km. It looks very different from the synthetics in Figure 3, calculated for a source at the same depth in the same simplified crustal model. From this it is obvious that, based on synthetic waveform characteristics alone, phase interpretation is rather difficult. Real records of seismic waves that have been radiated by extended seismic sources and travelled through complex media, resulting in multi-pathing and long codas, may look very different. In Figure 4 this holds especially for records at some stations in the distance range of phase triplication (e.g., station SPAK and SIERE). Also the amplitude ratio between Pg and Pn exhibits strong variations between some stations (e.g., EMV and HEI), probably due to 3D Moho topography and/or the influence of the source radiation pattern, which effects the Pn/Pg amplitude ratio because of the rather different take-off angles of these two waves from the source. Whereas in the case of EMV Pn is likely to be missed and Pg be picked as first arrival phase Pn is an unexpectedly large and sharp onset at station HEI. Moreover, for distances above some 400 km there is a tendency of Pn (which then becomes more and more a diving phase in the uppermost mantle, usually not modelled by any synthetics) to have larger amplitudes than Pg, which is more strongly attenuated and/or scattered in the more complex upper crust. This is illustrated by many record examples in DS 11.1. Similarly, as obvious from Figure 4, the determination of the phase type from generalized waveform features alone can also be rather ambiguous in the range of the crossover distance between Pg and Pn.

Since most applications like hypocenter localisation and travel-time tomography are (still) based on first arrivals only, an assessment for phase identification should avail of some guidance to avoid inconsistencies. The relevance of correct phase interpretation becomes even more evident for S-wave picking (see section 2.6). For use in routine first-arrival studies and subsequent special tomographic studies at local to regional distances, we therefore propose to identify crustal phases according to the rough waveform characteristics outlined below and to decide about their suitability for first arrival studies according to the assessment scheme presented in Table 1.

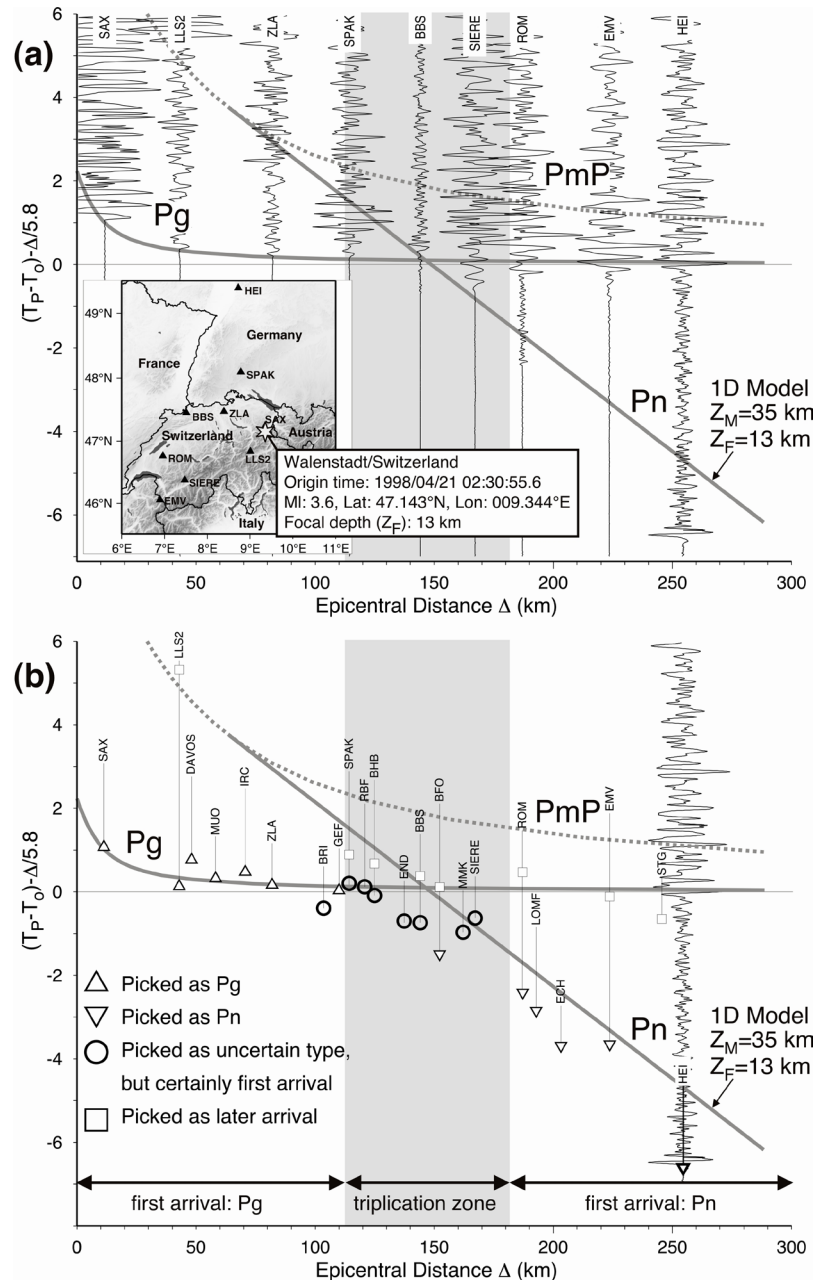
**Table 1** Suitability assessment scheme of crustal phases for routine first arrival studies.

| Phase label | Phase is...                               | Phase used for routine 1 <sup>st</sup> arrival studies |
|-------------|-------------------------------------------|--------------------------------------------------------|
| Pg, Sg      | Direct (crustal)                          | Yes (if first arrival)                                 |
| Pn, Sn      | Moho-refracted                            | Yes (if first arrival)                                 |
| PmP, SmS    | Moho-reflected                            | No                                                     |
| P1, S1      | Unknown type, but certainly first arrival | Yes                                                    |
| P2, S2      | Unknown type, second arrival              | No                                                     |
| P3, S3      | Unknown type, third arrival               | No                                                     |
| P, S        | Unknown type, uncertain if first arrival  | No                                                     |

- **Pg:** Impulsive high-frequency onset. At short distances, absorption is small and dispersion tends to sharpen the wavelet front (Seidl and Stammer 1984). Pg is first arrival close to epicenter.
- **Pn:** Usually with small amplitude at  $\Delta < 400$  km, impulse or emergent onset. The wavelet front may be smoothed for increasing distance since absorption becomes dominant (Seidl and Stammer 1984). Pn is first arrival for larger epicentral distance, usually followed several seconds later by stronger Pg and PmP phase, if Pg is not yet

strongly attenuated at distances  $\Delta < 400$  km. As a rule of thumb, Pn from a surface focus takes over Pg at  $\Delta \approx 5 z_M$ , for deeper sources, however, at shorter distances.

- **PmP**: Has the largest amplitude of all crustal P phases because of total reflection from the Moho at overcritical incidence angles. PmP is never first arrival but follows very closely Pg (with  $\delta t < 2$  s) after Pn has taken over as first arrival.

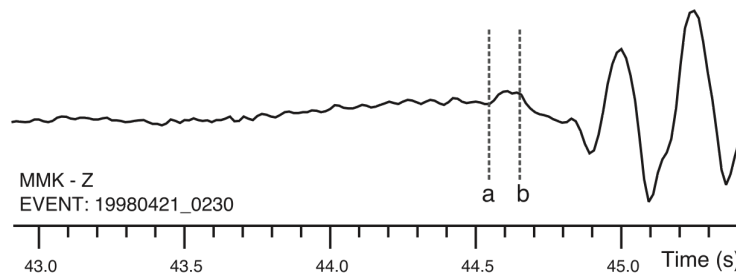


**Figure 4a)** Velocity-reduced record section of a local earthquake near Walenstadt, Switzerland. Amplitudes are normalised to maximum amplitude of each trace. Solid and dotted lines indicate synthetic travel-time curves for Pg, Pn, and PmP.

**b)** Velocity-reduced phase picks, crosschecked against synthetic travel-time curves derived from a simplified crustal model for the area under investigation (see text for discussion). Copy of Fig. 3 in Diehl et al. (2009, p. 545) with © granted by the Geophysical Journal International.

### 2.3 Phase Picking Part III: First-motion polarity and its error assessment

Information about first motion polarity is mainly needed for focal mechanism determination. Usually the polarity is denoted as ‘Up’ or ‘Down’. Figure 1 represents an example for certain polarity identification. On the other hand, it is impossible to determine any polarity for the example of Figure 2. However, besides these two obvious cases, it is sometimes necessary to use an intermediate quality class for the first motion polarity. Time ‘a’ and ‘b’ in Figure 5 represent two possible positions for arrival time picks. In this case, the polarity is depending on the location of the pick. If the pick is assigned to position ‘a’, the polarity would be ‘Up’. If position ‘b’ is used, the polarity has to be ‘Down’. In both cases, we are not sure about the actual polarity, therefore we have to introduce an intermediate quality classes ‘+’ or ‘-’. The error assessment scheme used here is summarized in Table 2.



**Figure 5** First-motion polarity and its error assessment: In this example the first-motion polarity is not independent from the pick position. If the pick is assigned to position ‘a’, the polarity would be ‘Up’. If position ‘b’ is used, the polarity has to be ‘Down’. In both cases, we are not sure about the actual polarity. Therefore we have to use the intermediate quality classes ‘+’ or ‘-’.

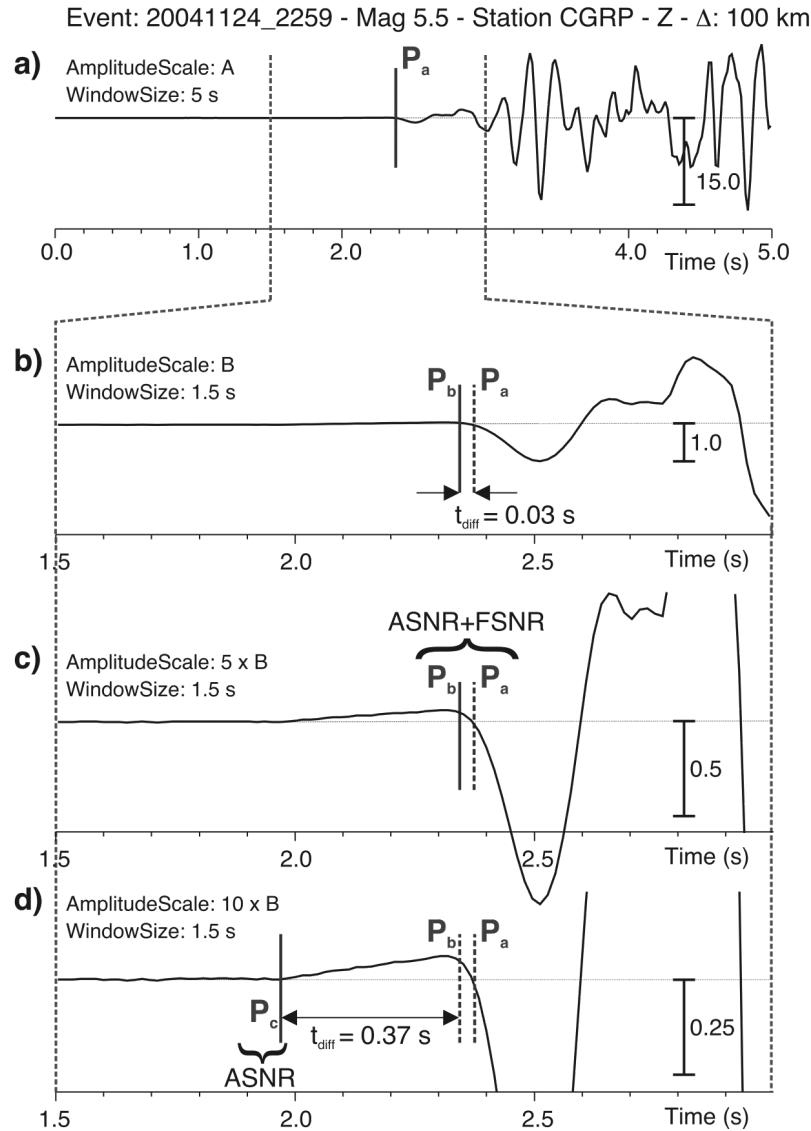
**Table 2** Error assessment used for first-motion polarity.

| Polarity label | Polarity is... | Weight                                |
|----------------|----------------|---------------------------------------|
| U/D            | Up/Down        | Polarity is identified with certainty |
| +/-            | Up/Down        | Polarity is identified but uncertain  |
| N              | None           | Polarity cannot be identified         |

### 2.4 Size of time window and amplitude scaling

The influence of time-window size and amplitude scaling on picking accuracy and phase identification as described e.g. by Douglas et al. (1997) is demonstrated in Figure 6a-d. In the uppermost example, the used time window is 5 s (Fig. 6a). The picked onset is labelled as  $P_a$ . For a narrower time window of 1.5 s, the onset at  $P_a$  seems a bit too late (Fig. 6b). Thus, we would pick the phase at position  $P_b$  in this window. The result is even more different, if the time window size remains the same, but the amplitude is multiplied by a factor of 5 (Fig. 6c) and 10 (Fig. 6d) respectively. An earlier phase is visible, which can be picked as  $P_c$  (Fig. 6d). But not only the absolute timing of the picked phase depends on window size and amplitude scaling, also the assigned uncertainty interval. The error interval might be smaller for the narrower time window and the magnified case.





**Figure 6** Different phase picking, caused by variable use of time window size (a, b) and scaling (c, d). See text for further discussion.

Considering the magnitude of the event, the small, ramp-like phase  $P_c$  in Figure 6 might be associated with the initiation of the fault rupture.  $P_c$  is obviously the first arriving phase, but it is arguable if it can be observed at stations with lower SNR (e.g., at larger epicentral distances). To obtain a consistent set of picks for the same earthquake it may be considered to pick the first prominent phase ( $P_b$  in Figure 6) instead. In contrast to the small precursory phase, the first prominent phase is expected to be observable at the majority of stations. Only precursors related to rupture complexities could be treated this way. Instead,  $P_c$  in Figure 6 might also be interpreted as the evolving Pn, taking over Pg as the first arriving phase. In this case,  $P_c$  has to be picked as the first arriving phase and cannot be ignored. Complexity of the source function usually scales with the rupture area. Therefore, small events ( $M < 4$ ) can be approximated as point sources in the far field and the shape of Pg is expected to be rather simple and impulsive. As mentioned earlier, an evolving Pn phase beyond the take-over distance between Pg and Pn can have the signature of a precursor signal. Similar effects are expected in subduction environments, where slab-phases are commonly observed. Such slab-



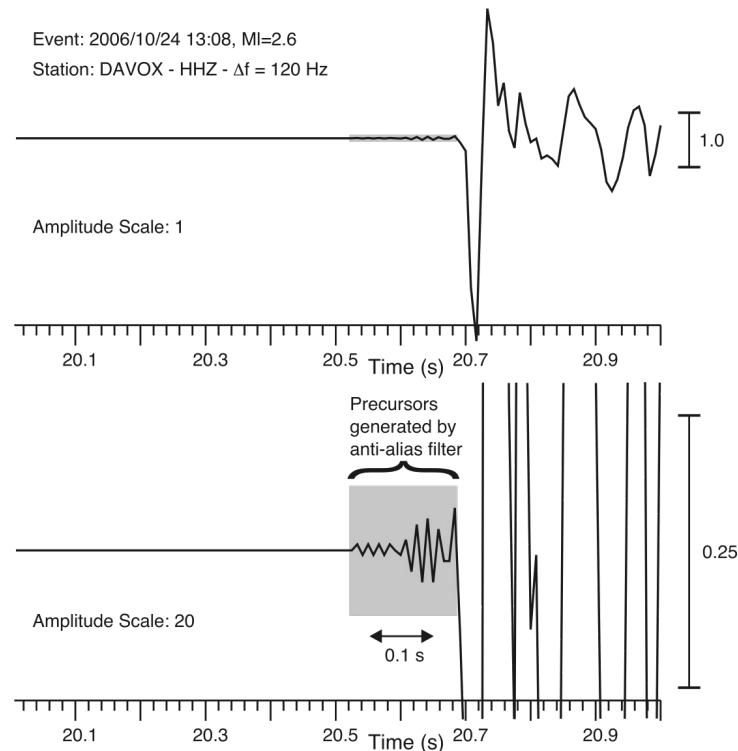
phases travel through the faster lithosphere of the subducting plate and manifest as precursors prior to the direct phase. Finally, it should be noted that interpretation of precursor signals requires the detailed knowledge of the complete instrument response as described in the following section. As demonstrated by Scherbaum (2001), precursor signals are often just artefacts of the instrument's anti-alias filter.

## 2.5 Aliasing and waveform filtering

From signal theory we know, that the correct digital representation of a continuous waveform depends on the sampling frequency. To avoid aliasing, a signal with frequency  $f$  has to be sampled with a sampling frequency  $\Delta f > 2f$ . Thus, the Nyquist frequency is defined as  $f_N = 1/(2\Delta t) = \Delta f/2$ . Any frequencies higher than  $f_N$  are aliased into lower ones. Therefore, we have to ensure that picking accuracy and phase identification will not be affected by sampling rates, which are rather small. As an example we consider a waveform sampled at  $\Delta f = 20$  Hz or  $\Delta t = 0.05$  s, which corresponds to a Nyquist frequency of  $f_N = 10$  Hz. Concerning the observation error of a phase pick, the uncertainty intervals can not be smaller than the sampling interval. Therefore, the minimum picking uncertainty is defined as  $\pm\Delta t$ . To avoid possible inconsistency due to significantly different sampling rates used at different stations, we suggest to use only data with  $\Delta f \geq 40$  Hz for local earthquake studies.

Aliasing occurs when the data are sampled, and once this occurs, the data cannot be 'unaligned'. Therefore, seismic data are usually filtered with an analogue anti-aliasing filter to remove frequencies above the Nyquist frequency before sampling. Furthermore, modern acquisition systems make use of oversampling and decimation techniques (Scherbaum, 2001; NMSOP Chapter 6). Such systems imply digital anti-alias filters, which often denote symmetrical or acausal impulse responses to avoid time shifting of the phases (zero-phase). As a consequence, the onset of very impulsive signals (also signal with frequency close to Nyquist frequency) may be obscured by 'acausal' precursory oscillations and their true onset becomes difficult, if not impossible, to determine. As an example for such precursors, Figure 7 shows the P-wave recording of a local earthquake near Davos, Switzerland (2006/10/24 13:08,  $M_l = 2.6$ ) at the nearby station DAVOX.

In principle, this effect can be minimized by an inverse filtering process, described e.g. in Scherbaum (2001). However, this nontrivial procedure requires knowledge of the original anti-alias filter coefficients. Especially for large and inhomogeneous data sets, compiling the correct information can be rather difficult or even impossible. An alternative way to deal with this problem is to apply a consistent error assessment to it, as it is described above. The significance strongly depends on the amplitude scaling as discussed in 2.4 and low-pass filtering or integration of the signal may also help to reduce the amplitudes of such precursors. Since these precursors are present only for very impulsive and/or high-frequent wavelets, they are expected to be limited to recordings close to the epicenter and/or with low sampling rates and large relative bandwidth.



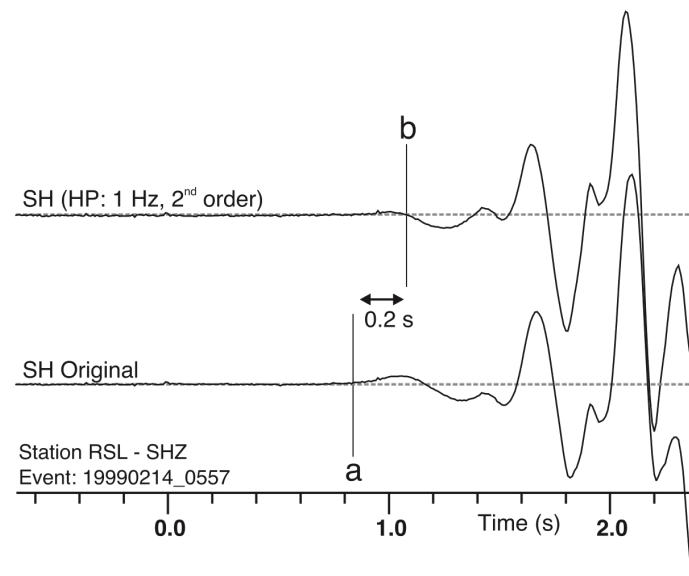
**Figure 7** Precursors caused by acausal anti-alias filtering observed on the broadband registration of a local earthquake near Davos, Switzerland (2006/10/24 13:08,  $M_I = 2.6$ ). Sampling frequency is 120 Hz. The precursors become obvious for enlarged amplitude scale (lower trace).

Filters are usually applied to enhance the SNR in the presence of noise. However, filtering can significantly change the true shape of the arriving wavelet and may also result in time-dependent phase shifts within the signal group (so-called “transient response”). These effects are particularly pronounced in narrow-band records (see section 4.2 in Chapter 4, e.g., Fig. 4.18). Therefore, care has to be taken when applying filters in the picking process. First of all, it should be noted that the original seismic record itself is the output of a filter (transfer function of sensor and digitizer), which distorts the true ground motion input depending on its relative bandwidth. Only a record of infinite bandwidth could reproduce the ground motion undistorted. Commonly used band-pass or high-pass filters not only reduce the amplitude of the noise, but can also result in distortion and phase shift of the signal. The phase shift introduced by a filter depends on the instrument response, the filter response and the frequency content of the signal. The closer the signal’s dominant frequency is to the corner frequency  $f_c$  of the filter, the stronger the expected effect on the signal. Moreover, the steeper the flanks of the filter (higher order), the stronger are the expected effects on frequencies close to  $f_c$ .

Figure 8 demonstrates the effect resulting from the application of an arbitrary filter to a velocity proportional short period record. The corner frequency of the 2<sup>nd</sup> order high-pass filter ( $f = 1$  Hz) is close to the dominant frequency of the first part of the wave group. Application of a 2<sup>nd</sup> order high-pass increases the effective steepness of the velocity proportional pass-band to the 5<sup>th</sup> order. This leads to an increased distortion of signals close to corner frequency, even for  $f > f_c$  (see also Figures 7a and 7b in IS 5.2). As a result, the filter significantly affects the low-frequent part of the signal as demonstrated in Figure 8. Position

‘a’ and ‘b’ represent two possible arrival-time picks on the original and the high-pass filtered record. Time difference between the two positions could exceed 0.2 s, a value much larger than the uncertainty of each individual pick. Moreover, the filter process affects the first-motion polarity. A similar effect is demonstrated in Fig. 4.10 of Chapter 4.

In general, the use of filters has to be consistent for arrival-time picking. If broadband and short-period instruments are present in the same network, a common frequency band should be used to avoid inconsistencies between the two types of instrument responses. The application of a 2<sup>nd</sup> order high-pass with  $f_c = 1\text{ Hz}$  to a broadband record is appropriate to sufficiently simulate a short-period instrument response. Further practical guidelines for the use of filters are provided in section 3.1.



**Figure 8** Effect of 2<sup>nd</sup> order Butterworth high-pass filter (1 Hz) on a short-period waveform. Note the significant change in waveform between the original and the filtered trace. The earliest low-frequency onset is hardly observable on the filtered trace. Position ‘a’ and ‘b’ represent two possible arrival picks for each trace with a time difference that exceeds 0.2 s. See text for further details.

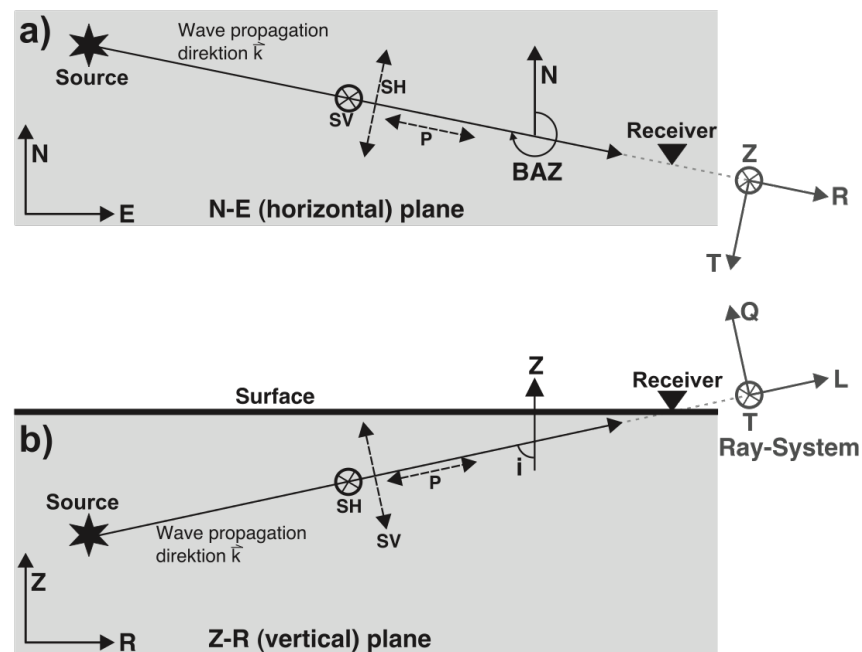
## 2.6 Additional comments for S-wave picking

Most of the previous findings apply likewise to S-wave picking, however, identification and picking of S-waves is usually more difficult, since its onset is often superposed by the P-wave coda. Furthermore, we have to consider S-wave splitting due to possible seismic anisotropy and the presence of Sp-converted precursors due to the presence of shallow strong velocity discontinuities.

### 2.6.1 Polarization of the wavefield

According to linear theory of wave propagation in homogeneous isotropic media P and S waves are fully independent solutions of the equation of motion results with linear particle motion. Figure 9 illustrates the particle motions of P and S waves in the horizontal (a) and vertical (b) planes of wave propagation. The particle motion of the P wave is parallel (or

longitudinal) to the direction of wave propagation vector  $\mathbf{k}$ . In contrast, the displacement field of the S wave is transverse to  $\mathbf{k}$  and can be subdivided into a vertical component (SV) and a horizontal component (SH). The vector  $\mathbf{k}$  can be described in the Cartesian ZNE-system of the seismometer (vertical, north-south, east-west) by two angles: the backazimuth ( $BAZ$ ) and the angle of incidence ( $i$ ). Both angles can be determined from polarization analysis of a three-component recording in a window around the P-wavelet. If station and epicenter coordinates are known,  $BAZ$  can also be derived from the orientation of the great-circle including station and epicenter. If focal depth and epicentral distance are known, we can also determine the theoretical angle of incidence in a given velocity model. However, in real inhomogeneous Earth the true arrival angles  $BAZ$  in the horizontal and  $i$  in the vertical plane depend on frequency and thus wave length (see Fig. 2.6 in Chapter 2 and Fig. 11.23 in Chapter 11).



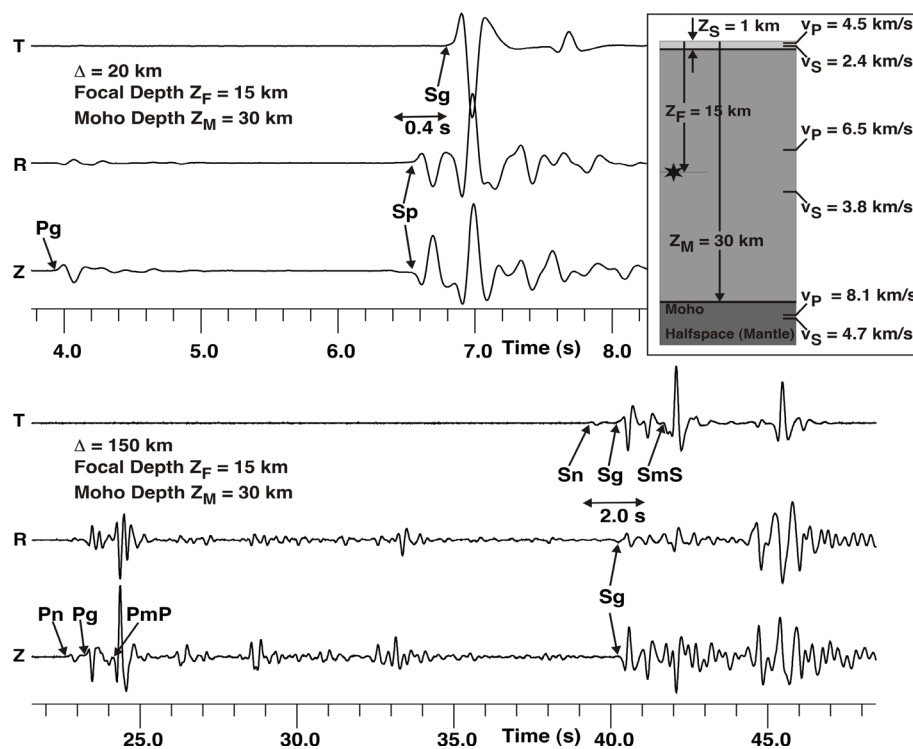
**Figure 9** Particle motion for P and S wave along wave propagation direction a) in the horizontal plane and b) in the vertical plane.

Both angles can be used to rotate the ZNE coordinate system into a local ray-system. Within the ray system P, SH, SV components can be partly separated from each other. The horizontal rotation of ZNE by  $BAZ$  results in the ZRT-system as demonstrated in Figure 9a.

The T-component denotes the transverse component, i.e., perpendicular to the direction of wave propagation in the horizontal plane, R coincides with the radial component, i.e., in the horizontal direction of wave propagation. In the ideal case of a flat-layered isotropic medium, the T-component contains only SH energy. P- and SV-energy are still mixed on Z and R to a variable degree, depending on the incidence angle  $i$  and the source mechanism (see Chapter 3). An additional rotation of ZRT by the incidence angle  $i$  results in a decoupled ray-system LQT. The complete P-wave energy is then recorded on the L-component (longitudinal) and the Q-component contains the SV energy only. The T-component remains the same as in ZRT.

Since both P and SV are polarized in the vertical plane of propagation, SV-wave energy incident on a velocity discontinuity can partly be converted into a P wave during reflection at or transmission through such an interface (or vice versa P into SV). If such an interface is close to the surface and denotes a strong velocity contrast (e.g. basement-sediment interface), Sp precursors are generated, which then arrive shortly prior to the initial SV-wave. These Sp precursors may be misinterpreted as a too early onset of the S-wave. The use of a ray-system can be used to identify such precursors, as demonstrated in Figure 10. Sp precursors are not present on the T-component, since there is no coupling between SH and P. However, the ratio between radiated SH and SV energy strongly depends on the focal mechanism and its orientation in space and thus also on the observation azimuth.

To avoid the misinterpretation of Sp precursors as the true S-wave arrival we recommend the use of a ray system (ZRT or LQT) and pick the S wave preferable on the T-component. Furthermore, the use of a ray-system can be helpful for a better discrimination of S energy generated by the P coda and to identify phases like Sn as demonstrated in Figure 10. In addition, the application of a Wood-Anderson simulation filter (high-pass with  $f_c = 1$  Hz in combination with an integration from velocity to displacement) can be useful to enhance the S-wave onset. The final S-pick, however, should be crosschecked with its position on the velocity proportional record.



**Figure 10** Theoretical seismograms calculate by the reflectivity method of Fuchs and Müller (1971) for a simplified 1D crustal model (upper right inset) using the *refmet* code provided by T. Forbriger. The uppermost layer represents a sedimentary basin of reduced P- and S-wave velocities. The uppermost traces show vertical, radial and transverse component at a distance of 20 km. The Sp precursor phase is clearly visible on Z and R component. The lowermost traces correspond to a distance of 150 km. The amplitude of the first arriving Sn phase is rather weak compared to later arrivals (Sg, SmS) and only visible on the T component. All amplitudes are normalized by the calculated maximum record amplitude for the considered station.

Another form of S-wave “splitting” results from travelling through anisotropic Earth material. In such media part of the SV-wave energy is coupled with some time shift into the T component, producing instead of a linear SV motion a more or less elliptical S-wave motion in the R-T plane. The degree of ellipticity and the amount of time shift between the related S onsets in the R and T component depend on the degree of anisotropy of the medium through which the S wave is travelling. Furthermore, the time shift depends on the angle between the anisotropy axis and the incident angle of the S wave and therefore it depends also on the epicentral distance. This allows to determine the degree of (specifically horizontal) anisotropy and its orientation in space by restoring iteratively the linear motion of SV expected for isotropic media. This is illustrated, e.g., by Fig. 2.7 in Chapter 2 for teleseismic SKS waves. The time shift between the first S arrivals in R and T may range between several tenths of seconds and several seconds, depending on what type of S waves one analyzes at which frequencies and source distances and with focus on which parts of the Earth (e.g., crust, upper mantle or deeper). Also in this case it is helpful to analyse arrivals in the S-wave range and their polarization in a ZRT or LQT system in order to discriminate between onsets in different components due to wave conversion at a discontinuity or caused by true S-wave splitting due to travel through anisotropic Earth material.

### 3. Consistent hand picking procedure

The findings discussed in the previous sections will be combined to a standard procedure for consistent P- (and S-) phase picking in the following sections. This requires, first and foremost, that the picking is made on some standardized type of seismic record and that time window length and the amplitude scaling are chosen consistently. Therefore, we will discuss in detail these major sources for inconsistent picking.

#### 3.1 Consistent waveform filtering

As demonstrated in section 2.5, filtering may lead to inconsistent picking. Therefore, we suggest the following guidelines:

- The original seismic record itself is the output of a filter, which distorts the true ground motion input depending on its relative bandwidth. Only a record of infinite bandwidth could reproduce the ground motion undistorted (see Chapter 4).
- Since for local/regional applications emphasis is on high-frequency signals one should preferably use either directly original short-period (SP) records or filter BB records accordingly so that the lower corner frequency  $f_{cl}$  of the passband is about 1 Hz.
- Short-period (SP) channels (EH, SH):  
Typical SP sensors are electromagnetic seismographs of the geophone type with a seismometer eigenfrequency  $f_s \approx 1\text{ Hz}$  (or higher). They have a response proportional to ground motion velocity between  $f_s = f_{cl}$  (lower corner frequency) up to the corner frequency  $f_{cu}$  of the anti-aliasing filter. If  $f_{cu} = 10\text{ Hz}$ , respectively  $50\text{ Hz}$  (i.e., if the data are sampled with  $\Delta f \approx 40\text{ Hz}$ , respectively  $200\text{ Hz}$ ) then this corresponds to a velocity-proportional relative bandwidth (RBW) between some 3.5 to 5.5 octaves or 1 to 1.7 decades. Note that this is much less than for modern broadband seismographs,

which also have a velocity-proportional response but within the range  $f_{cl} \ll 1$  Hz (up to  $\approx 0.3$  mHz) and  $f_{cl} \approx 10$  to  $20$  Hz (with  $\Delta f \geq 40$  Hz), corresponding to an RBW between some 2 to 4 decades. Note, however, that the RBW of common bandpass-filtered SP seismographs used for teleseismic observations is only about 1 and 2 octaves (e.g., Fig. 4 in Granville et al., 2005) and should not be mistaken with the SP sensors considered here for local and regional travel-time studies.

- To avoid aliasing-effects, be sure that the upper corner frequency  $f_{cu}$  of the filter is well below Nyquist frequency, typically at  $\leq 1/4^{\text{th}}$  of the sampling rate  $\Delta f$  and with a steep roll-off for  $f > f_{cu}$ .
- Unfiltered SP sensors have already a limited bandwidth with a step  $3^{\text{rd}}$  order roll-off of the response for  $f < f_{cu}$  (see Fig. 5.3 in Chapter 5) which results in significant phase shift and distortion of the earliest parts of the waveform (for frequencies up to about 3-4 times  $f_{cl}$ ).
- Therefore, one should avoid further frequency filtering when ever possible. Apply a  $2^{\text{nd}}$  order 1 Hz high-pass filter only if necessary, e.g. in presence of low-frequency noise. This filter will not affect the velocity-proportional passband, yet be aware that it will steepen the roll-off for  $f < f_{cl}$  to  $5^{\text{th}}$  order, thus causing larger phase shifts and waveform distortions up to  $f \gg f_{cl}$ . Therefore, compare original waveform with filtered one before final picking. Also check, if the filter changes first motion polarity.
- Apply possible low-pass or band-pass filters only if necessary, e.g., in presence of high-frequency noise. Compare original SP waveform with filtered one before final picking.
- Broadband (BB) channels (HH, BH):  
Modern feedback-controlled BB seismographs are widely in use now. They have also velocity-proportional response for  $f > f_{cl}$  (see Fig. 11.27 in Chapter 11). If one applies a high-pass filter with corner frequency 1 Hz to such a wider broadband records with  $f_{cu} \ll 1$  Hz it will remove the long-period noise often present in BB recordings, and in addition, will simulate well the response of an original 1 Hz SP record of the geophone type. Thus, filtering BB records with a constant  $f_{cu} = 1$  Hz will minimize the inconsistency due to different types of BB sensors used in practice.
- S-wave identification and picking:  
Application of a Wood-Anderson simulation filter ( $f_{cl} = 1$  Hz in combination with an integration from velocity to displacement) can be useful to enhance the S-wave onset. In addition, component rotation to the RT-system can help to identify the S-wave on the T-component.

### 3.2 Consistent window size and amplitude scaling

A uniform procedure is required to minimize subjectivity in phase picking due to irregular choice of time window size and amplitude scaling as shown in Figure 6. On the other hand, it turns out to be very difficult to define universal rules, especially, if the method should be feasible for picking of large data sets. A consistent choice of time window and amplitude

scaling should satisfy two conditions: A fixed time window length and amplitude normalisation with respect to a reference scale.

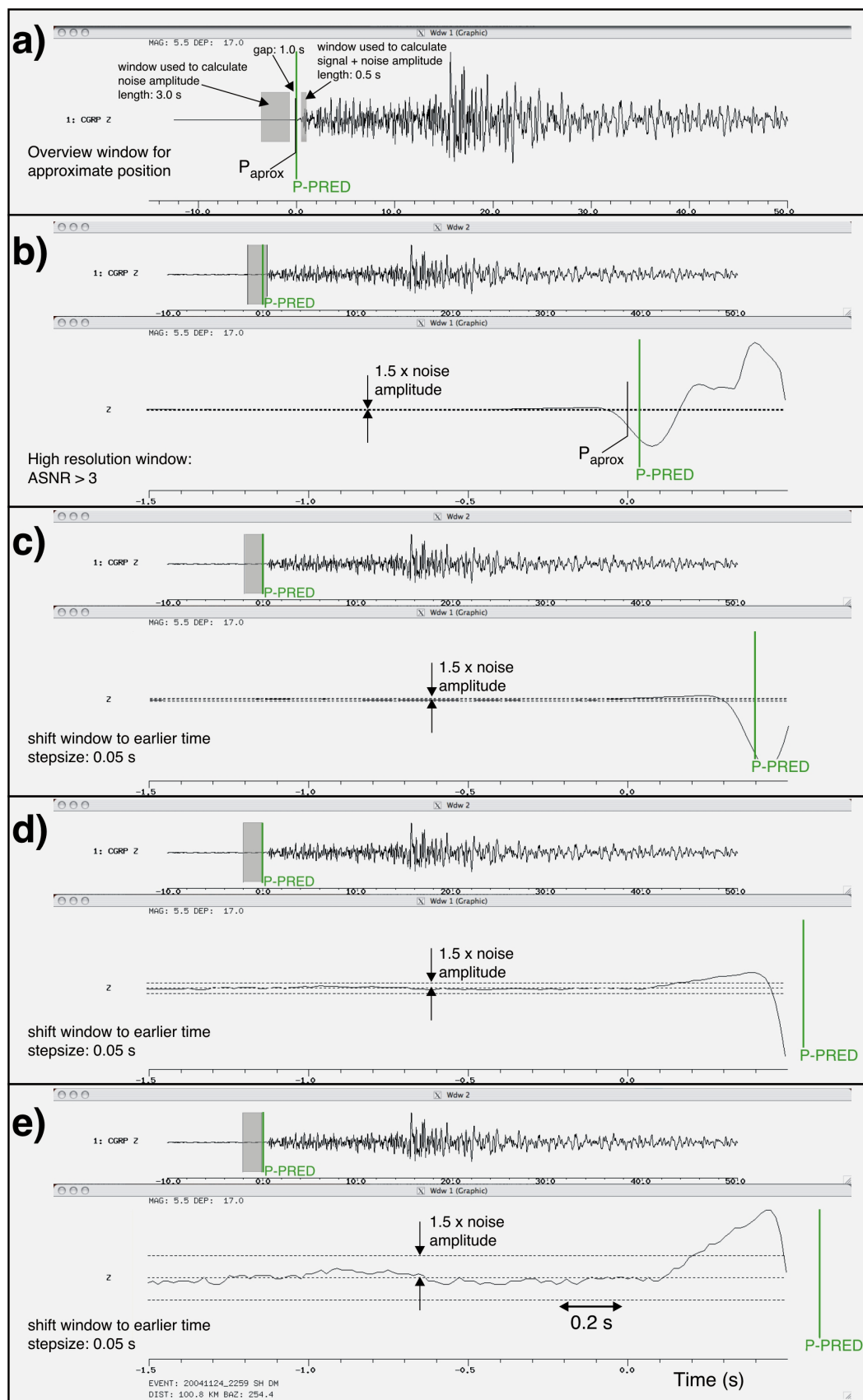
As already mentioned, universal rules for these conditions are difficult to define. Concerning the time window length, the choice depends on the dominant frequency of the signal, but also on the  $ASNR$ . In the presence of strong noise or in the case of a weak and emergent signal, it is often better to use a broader time window to determine the change in frequency and follow the transition from the signal to the noise part. In addition, many picking tools normalise the amplitude to the maximum amplitude within the selected time window and therefore amplitude scaling depends also on the size and the position of the time window. A possible approach to minimize the inconsistency related to time window choice and amplitude scaling may consist of a mixture of iterative selection of window sizes and sliding them stepwise from the signal part towards the noise dominated part. For a further description, we apply this approach to the example shown in Figure 6. A priori, we define four possible time windows used for picking:

|                                                                      |      |                   |
|----------------------------------------------------------------------|------|-------------------|
| Length of <b>H</b> igh- <b>R</b> esolution <b>W</b> indow (HRW):     | 2 s  | $ASNR \geq 3$     |
| Length of <b>M</b> id- <b>R</b> esolution <b>W</b> indow (MRW):      | 5 s  | $2 \leq ASNR < 3$ |
| Length of <b>L</b> ow- <b>R</b> esolution <b>W</b> indow (LRW):      | 10 s | $ASNR < 2$        |
| Length of <b>V</b> ery-low- <b>R</b> esolution <b>W</b> indow (VRW): | 15 s |                   |

In the first step we choose a broad overview window, which includes the whole coda of P or even parts of the S-coda (Fig. 11a). ‘P-PRED’ denotes the theoretical arrival time predicted in a velocity model for a preliminary hypocenter. The predicted arrival time can be useful, but only as a rough guide and is not necessary. We pick the approximate position  $P_{approx}$  of the earliest arrival visible in this window. Based on  $P_{approx}$ , the maximum noise and signal+noise amplitude is estimated. To avoid parts of the signal in the noise window, we define a safety gap of  $\pm 0.5$  s around  $P_{approx}$ . We use a length of 3 s for the noise window and of 0.5 s for the signal+noise window. We define the noise threshold as  $\pm 1.5$  times the absolute noise amplitude within the noise window. In addition, we calculate  $ASNR$  from the absolute noise and signal+noise amplitudes. Depending on the resulting  $ASNR$ , we choose the appropriate window size as described above.

In Figure 11b  $ASNR$  is several times larger than 3, therefore we use the HRW-size for picking. Between Figure 11b and 11e, the time window is shifted stepwise towards earlier times. The maximum amplitude within each time window is used for amplitude scaling. As long as the upper and lower noise threshold (dashed horizontal lines) are close to each other, the time window can be shifted further to earlier times (i.e. enlarging the amplitude scale). Finally, in Figure 11e, the amplitude scaling allows an almost perfect discrimination between noise and signal. In this example, the suggested procedure prevents missing the small precursors. If we could not clearly distinguish between noise and signal with the actual window size, or if we are not sure whether it is definitely the first arrival (e.g., in the presence of strong noise), we repeat the sliding procedure with a larger time window (MRW or LRW). Subsequently, we apply the consistent error assessment as described in section 2.1 to determine  $t_L$ ,  $t_E$ , and  $t_A$ . The upper and lower noise thresholds (dashed horizontal lines) are used for the consistent determination of  $t_L$ .





**Figure 11** Sliding time window procedure used for consistent phase picking. See text for further details.

### 3.3 Consistent phase identification

Provided that a hypocenter can be calculated, the actual phase identification can be crosschecked against a simplified crustal model (similar to Figure 4b). Another helpful tool to minimize inconsistency due to phase misinterpretation can be the use of synthetic or predicted arrival times, especially for re-picking of seismograms. However, as the comparison of Figures 3 and 4 shows, the actual phases may look very different and also their onset times may significantly deviated from the synthetic ones. Therefore, one should never pick the onsets at the theoretically predicted arrival times and identify the phase on waveform synthetics, unless all the subsequently applied criteria for phase identification and picking, as described above, are sufficiently well fulfilled. Because our goal is not to confirm already existing, usually grossly simplified 1-D models but to substitute them by high-resolution 2- and/or 3-D models by picking well-established deviations of correctly identified real phase onsets from theoretically predicted ones.

If, however, a reasonably accurate local or regional velocity model is available and a preliminary hypocenter has been calculated on the basis of such a model, we can calculate synthetic first arrival times for each event-station pair, prior to the picking process. This provides three major advantages. First, they may guide the picker to the right phase (but as said above, with the reliability depending on model quality) and thus reduce phase misinterpretations. Second, the synthetics can also be used to arrange the seismograms according to their expected arrival times. Thus, one can start the picking at the closest station and does not waste too much time with likely low quality arrivals. In addition, the synthetics may give at least some rough guidance for recognizing and tracking possible, theoretically expected changes in waveforms from station to (neighbouring) station. Third, it represents a rough test for correct synchronization of the timing system. In case of obvious timing errors of a station, we would observe a large discrepancy between synthetic and actual waveform arrival.

### Acknowledgments

Seismograms were processed and displayed using the ‘SeismicHandler’ package (Stammler 1993) and additional figures were generated with the Generic Mapping Tool by Wessel and Smith (1995)

### References

- Bormann, P., Klinge, K., and Wendt, S. (2002). Data analysis and seismogram interpretation. In: Bormann (ed), *IASPEI New Manual of Seismological Observatory Practice (NMSOP)*, GeoForschungsZentrum, Potsdam, ISBN 3-9808780-0-7, Vol. 1, Chap. 11, 100 pp. or <http://nmsop.gfz-potsdam.de>.
- Di Stefano, R., Aldersons, F., Kissling, E., Baccheschi, P., Chiarabba, C., Giardini, D. (2006). Automatic seismic phase picking and consistent observation error assessment: Application to the Italian seismicity. *Geophys. J. Int.*, 165, 121–134.

- Diehl, T., Kissling, E., Husen, S., Aldersons, F. (2009). Consistent phase picking for regional tomography models: application to the greater Alpine region, *Geophys. J. Int.* **176**, 542–554.
- Douglas, A., Bowers, D., and Young, J. B. (1997). On the onset of P seismograms. *Geophys. J. Int.*, **129**, 681–690.
- Fuchs, K., Müller, G. (1971). Computation of synthetic seismograms with the reflectivity method and comparison with observations, *Geophys. J. Roy. Astron. Soc.*, **23**, 417–433.
- Husen, S., Diehl, T., Kissling, E. (2009). The effects of data quality in local earthquake tomography: Application to the Alpine region. *Geophysics*, **74**, WCB71-WCB79.
- Kulhánek, O. (1990). Anatomy of seismograms. *Developments in Solid Earth Geophysics* **18**, Elsevier, 78 pp.
- Kulhánek, O. (2002). The structure and interpretation of seismograms. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (eds.), *International Handbook of Earthquake and Engineering Seismology*. Academic Press, Part A, 333–348.
- Scherbaum, F. (2001). Of poles and zeros: Fundamentals of Digital Seismology. *Modern Approaches in Geophysics*, Kluwer Academic Publisher, 2<sup>nd</sup> edition,.
- Scherbaum, F. (2002). Analysis of digital earthquake signals. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (eds.), *International Handbook of Earthquake and Engineering Seismology*. Academic Press, Part A, 349–355.
- Seidl, D., Stammer, W. (1984). Restoration of broad-band seismograms (Part I). *J. Geophys.*, **54**, 114–122.
- Simon, R. B. (1981). Earthquake interpretations: A manual of reading seismograms. William Kaufmann Inc., 150 pp.,
- Stammer, K. (1993). Seismic Handler – programmable multichannel data handler for interactive and automatic processing of seismological analysis. *Comp. Goesci.*, **19**, 135–140.
- Wessel, P., and Smith, W. H. F. (1995). New version of the Generic Mapping Tool released. *EOS, Trans. Am. Geophys. Un.*, **76**, 329.

| Topic   | Seismogram Picking Error from Analyst Review (SPEAR)                                                                                                                                                                                                                                                                                                                                         |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Author  | <b>Cleat Zeiler</b> , University of Texas at El Paso, El Paso, Texas, USA;<br>now: Air Force Technical Application Center, Patrick AFB, Florida, USA,<br>E-mail: <a href="mailto:cpzeiler@miners.utep.edu">cpzeiler@miners.utep.edu</a><br><b>Aaron Velasco</b> , University of Texas at El Paso, El Paso, Texas, USA,<br>E-mail: <a href="mailto:aavelasco@utep.edu">aavelasco@utep.edu</a> |
| Version | July 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_IS_11.5">10.2312/GFZ.NMSOP-2_IS_11.5</a>                                                                                                                                                                                                                                                                                        |

## 1 Introduction

The Seismogram Picking Error from Analyst Review (SPEAR) project was initiated in 2006 to determine the reliability and accuracy of picks made by analysts. The process of picking is defined as the timing, measuring and naming of seismic phases used to locate and classify seismic events. Prior to this study only one analyst-based study had been conducted (Freedman, 1966a). While the general practice and theory of picking has not changed significantly from 1966, the technology and tools used to analyze waveforms has improved, along with the resolution of digitized waveforms (Lomnitz, 2006). Additionally, the 1966 study was limited to a small subset of analysts that used paper records. To continue and modernize the study of analyst based picks, we implemented a two-part study by first looking at available catalogue data (Zeiler and Velasco, 2009) and then by creating a common data set for analysts to pick.

The initial looks at archived data provided a limited data set to further understand pick error. Zeiler and Velasco (2009) established a pick model for a single analyst and found that picks from different institutions could be biased. However, they could not establish a complete error model. To produce a more reliable error model, a more robust technique to test and study the effects of different analysts reviewing the same seismograms must be implemented. The current SPEAR effort is designed to provide a common data set for analysts from varying experience levels to pick. An updated model is generated with each new participant. The current approach is designed to focus on signal quality and analyst experience. While many studies have tried to estimate pick error (Freedman, 1966a, 1966b, 1967a, 1967b and 1968; Buland, 1976; Pavlis, 1986 and 1992; Billings et al.; 1994, Sipkin et al., 2000), our approach can uniquely remove pick errors from the travel-time modeling errors.

## 2 History of the study

Zeiler and Velasco (2009) studied catalogue data to test the reliability of a single analyst's picks and compare the repeated picks from different institutions reported in the catalogue of the International Seismology Centre (ISC). The single analyst picks were obtained from a local network operator, who reviewed or made each pick on local to near regional events. With over twenty years' experience picking the waveforms, the operator showed consistency within 0.10 seconds when working on waveforms with high signal-to-noise ratio (SNR > 10). Since, the signal quality is not documented in the ISC catalogue, Zeiler and Velasco (2009) were only able to show that there can be a bias based on institution or who is making the

picks. The initial study confirmed that the study of pick error could not fully be conducted on pre-existing data base information.

The current iteration of the SPEAR study relies on a common data set of varying types of waveforms for analysts to pick. The waveforms include the same event recorded at various distances, different types of sources, and synthetic pulses with varied SNR. A broader suite of seismograms would be ideal, but limiting the time constraint to pick the seismograms has been a primary goal of the experiment. Currently the data set can be picked in under an hour by most analysts. Since the project is an on going effort, **we encourage interested institutions or analysts to contact the authors** for any related question **and download the common data set and user guidelines** via the link *Download programs and files* from the NMSOP-2 front page.

The SPEAR study has collected picks during two main efforts (Table 1). In each effort the analysts have only picked phases they could identify. Then, after a minimum of six months, the analysts have re-picked the data to test their consistency of picking over time. The commitment to pick the data set twice is not required for the study, but is encouraged to help understand picking precision and its improvement with time and experience.

**Table 1** The growth of the SPEAR project from the inception to the last review of all picks that have been submitted by 2011.

| Year | No. of Analysts | No. of institutions | No. of waveforms | No. of picks |
|------|-----------------|---------------------|------------------|--------------|
| 2006 | 12              | 4                   | 25               | 647          |
| 2011 | 27              | 7                   | 39               | 1633         |

### 3 Participation in the study and picking procedure

#### 3.1 Assumptions

Within the NMSOP (Chapter 11) one finds general guidelines of how to analyze and pick seismograms. We therefore assume that an analyst is familiar with these techniques and the IASPEI approved standard phase nomenclature (IS 2.1). Moreover, IS 11.4 offers a “Tutorial for consistent phase picking at local to regional distances”. However, we do not want to discourage other possible techniques and **welcome participation from all analysts and automated systems**. In part, the study is to collect the methodology employed in analyzing seismograms and we encourage participants to provide their methodology once they have picked the common data set. This would help in generalizing and spreading the awareness of **best practice** and to identify and avoid in future frequently made mistakes. Since, most routine analysis is performed on events, we understand that not all picks can be associated to a particular phase; the naming of phases will need to be addressed with future work. We are interested in where an analyst records the time of an arrival. This picking process is initially performed on the raw seismograms and only then, if an analyst would like to introduce filtering and waveform enhancement to refine their picks, we will allow these procedures to identify additional biases. Windowing or zooming of the seismogram is encouraged to improve the recognition and timing of onset arrivals.

### 3.2 General procedure

After downloading or receiving the data set, follow the naming convention outlined in Table 2. If you are using a picking program that does not follow this convention, you will need to include a conversion table of the naming convention used when returning your picks. If additional phases are observed, place additional markers as needed and record the observed phase to be reported with the submission of the picks.

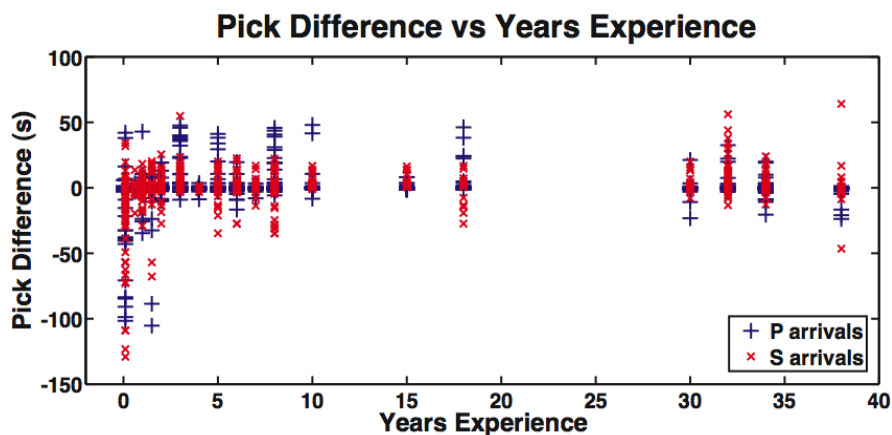
**Table 2** Naming convention for picks using SAC header variables.

| SAC header | Arrival type |
|------------|--------------|
| Amarker    | P or Pg      |
| T0         | S or Sg      |
| T1         | Pn           |
| T2         | Sn           |
| T4         | Lg           |
| T5         | Rg           |

In addition to the recorded phase picks from the data set, we record basic information on each analyst. We use the analyst information to derive correlation of picks based on years of experience, average number of seismograms picked in a year, number of seismograms picked in the past six months, institution/organization, and types of seismograms picked (local, regional, teleseismic, explosion). This profile, along with participant names and results, will not be released for comparison of analysts or institutions.

## 4 Results from SPEAR

The most striking result from the experiment is the trend observed in pick difference versus years of experience (Figure 1). The cut-off of having more than five years experience seems to group the analysts with the lowest pick difference and measurement repeatability. This phenomenon of 5 years experience has been observed in other research as the minimum time to achieve expertise in a given skill (Gladwell, 2008). However, we also note that the years experience is not always indicative of how the overall population will pick and further work is needed to clearly identify analysts that have achieved a competent level of picking.

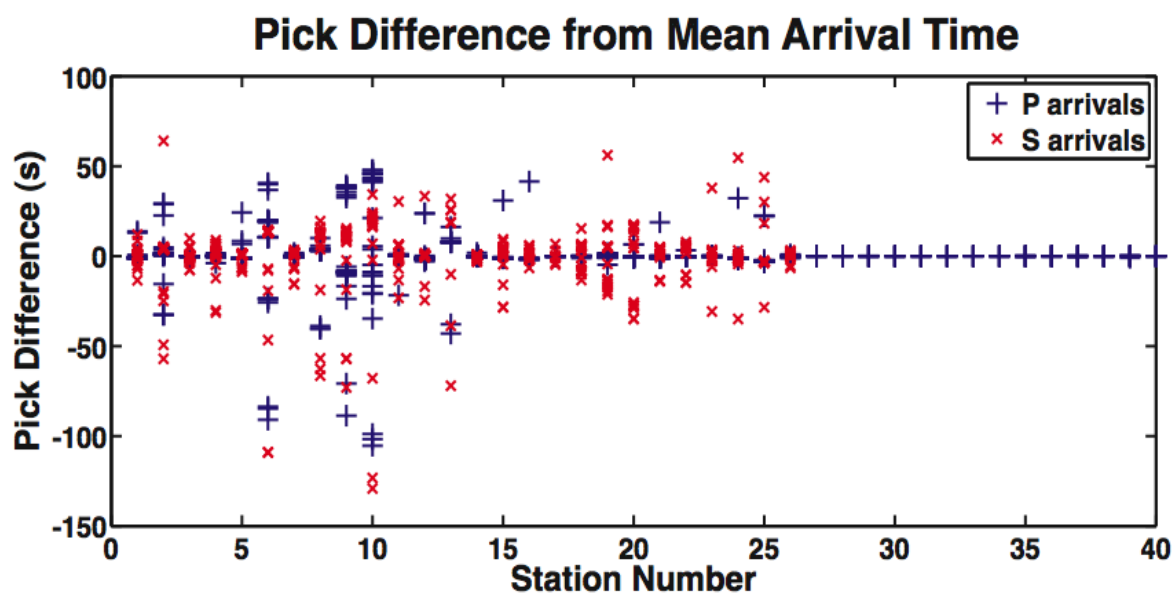


**Figure 1** The years experience of each analyst versus the pick difference calculated from the mean arrival time for S and P picks.

#### 4.1 Pick difference

The goal is to determine the arrival time of all phases recorded and we assume each analyst is measuring the same observation. This assumption has resulted in pick differences over a minute (Figure 2). While for standard locating procedures this magnitude of error would be unacceptable, for our purposes we can identify features in the noise field that cause false detections. In many cases, the observations that are over a minute from the mean arrival time have detections that are within seconds of them; this suggests that multiple analysts observed the same feature in the waveform. When participating in the study we encourage analysts to only record observations they can clearly identify. In every case there is at least one event recorded, but if an analyst is unable to detect a significant arrival, we encourage them to not make any measurements. However, if multiple events are observed or if additional arrivals are observed, we encourage the analyst to time and record these detections. In the case an arrival is misidentified we will correct this type of error and add the observation to the correct population.

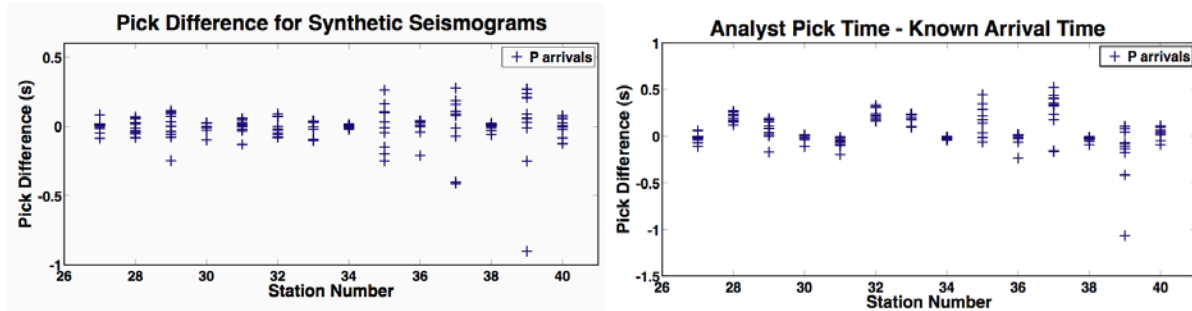
The pick difference is calculated from the average arrival time for each common observation. When new data is added the average arrival time is recalculated and a new pick difference is established (Figure 2). Since the true arrival time is known for the synthetic seismograms we also compute the pick error for stations 27-40; pick error for these stations can be calculated as an absolute term (Figure 3). Since, the other stations do not have a known arrival, only a relative pick error can be determined. To establish the reliability of the relative pick error, the synthetic measurements will be compared to the mean pick time (average arrival time measured by the analysts) and the known arrival time.



**Figure 2** The pick differences of P and S phases computed from the average arrival time. The synthetic traces did not have an S phase to pick.

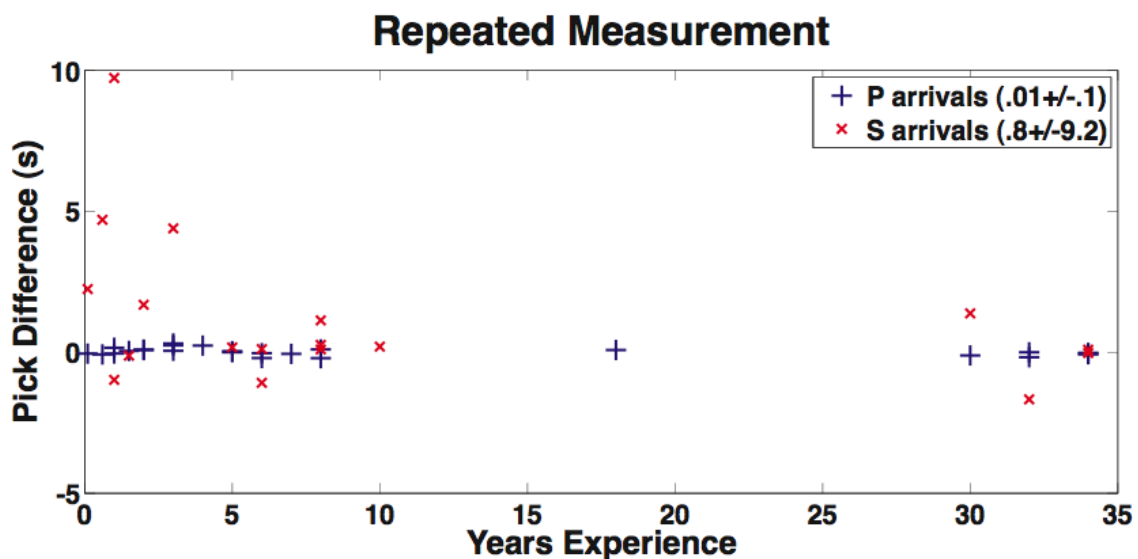
When you compare the analyst made picks on the synthetic seismograms, you see that overall the analysts made similar picks. However, when you compare the analyst picks to the known

arrival time, stations 28, 29, 32, 33, 35 and 38 show a bias of up to 0.5 seconds (Figure 3). All of the late picks are a result of emergent arrivals and analysts choosing to measure the arrival in the coda energy of the phase (the other stations that are biased late are due to the SNR). This presents a difficult variable to the model, since the traditional SNR cannot capture the true amplitude difference between the measured signal and noise.



**Figure 3** The pick difference for the synthetic pulses, (left) based of the pick difference from the mean arrival time and (right) based off the known arrival time.

Additionally we have placed a repeated seismogram in the data set to determine an analyst's repeatability (Figure 4). Overall the analysts are able to repeat their measurements with a high level of precision, with the type of phase measured and the years experience being the main factors influencing repeatability.



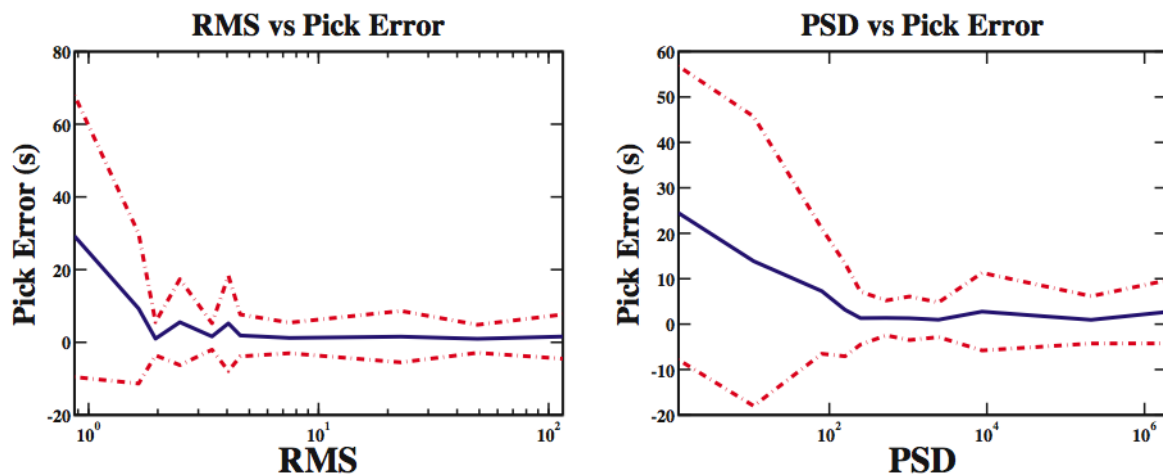
**Figure 4** The pick difference between repeated measurements on the same seismogram during the initial round of picking.

The overall goal of the experiment is to derive a pick model based on signal quality. However, no signal quality measurement has been found that adequately models the pick error associated with the process of timing seismic phases. We applied the traditional Short Term Average versus Long Term Average (STA/LTA) using an RMS method of computing the SNR (Figure 5). The STA/LTA model was broke into a two part model of picks with an SNR of 2 or less could have up to a 20 second pick error and all other picks have a pick error



of 0.1 seconds. This model is a bit disappointing because all picks under SNR 2 would be thrown out of location solutions. We tested other models to establish signal quality, however, to keep the focus and brevity of this information sheet, this topic needs to be studied further and reported in additional papers.

The best model we found so far to model the pick error is using the peak average ratio of the Power Spectral Density (PSD) of the signal and noise. This approach is similar to the Wide-Spectral band Ratio (WSR) described in Zeiler and Velasco (2009). The PSD ratio provided a broader window to establish a linear relationship between the lower quality signals and then a general pick error for high quality signals (Figure 5). The PSD ratio shows that a signal 10 times larger in amplitude than the back ground noise can be picked with the highest precision.



**Figure 5** The traditional signal quality of RMS (left) and peak average PSD ratio between signal and noise (right) plotted against the pick error to establish a pick model. The red dashed lines mark the standard deviation of the pick error for the signal quality measurement.

## 4 Summary

To be successful the SPEAR project needs the continued support of the seismological community by picking the common data set (obtained by contacting the authors or downloading from the NMSOP site). The initial results of SPEAR are promising, but we still need more participants to establish the statistical significance of the measurements. Therefore, we invite all interested seismological institutions, data analysis centers and observatories to participate in the SPEAR project. For **user guidelines** see the **Annex**.

Additional work is also needed to establish the best criteria for quantifying signal quality. The initial measurements are crucial to the event location processes along with the process of event refinement and classification. By fully understanding the errors associated with picking, seismological observatories can better tune their automated systems and improve velocity models. Along with these crucial elements a better understanding of achievable location precision can be assessed from the quality of picks. However, the biggest contribution will be to facilitate the sharing of picks for global studies and work load reduction.

## Annex

**SPEAR – User Guidelines**

If you are a single analyst, you can download the data sets via the link [Download programs and files](#) from the NMSOP-2 front page. For institutions with more than 2 analysts that will be participating in the study please contact the author for additional data sets. The data needs to be randomly ordered to prevent any learning bias. Once the data sets have been downloaded follow the guidelines in section 3 of IS 11.5. Also, please do not use any of the results provided in IS 11.5 to influence the method of picking.

Depending on the software that is used to analyze the data, common picking procedures need to be adopted. If you are using SAC the provided shell script will allow you to scroll through and pick the events while recording your picks in an output file. However, if you are using a different program the standard procedure is to review the three component data for each station. Only review one stations data at a time and proceed to the next stations data once all picks have been recorded. Do not return to any of the picks to correct or modify them once you have proceeded to the next station. The only enhancement to the waveforms is zooming or windowing the data. Please mark any phases you can identify (following Table 2 in IS 11.5), whether you can name them or not. Generally there is only one event per station, however, if additional phases are observed please measure the arrival time and report them. Stations 27-40 have only a single pulse to identify. Save the measured arrival times and e-mail them back to the first author: [cpzeiler@miners.utep.edu](mailto:cpzeiler@miners.utep.edu).

Once you have recorded the initial picks, feel free to re-pick the waveforms with any filtering or waveform analysis techniques that are common to your observatory. Send these picks in a separate file with a description of the technique used to record the observations. With the pick files also please submit the following information:

- years of experience
- average number of seismograms picked in a year
- number of seismograms picked in the past six months; institution/organization
- types of seismograms picked (local, regional, teleseismic, explosion)
- identify if you would be willing to re-pick the waveforms in 6 months to a year.

Feel free to contact the author for any additional help or clarification of the procedures.

**Acknowledgment**

The authors would like to thank the analysts that have voluntarily contributed time and energy to pick the common data set. The analysts are highly tasked and their donated time is greatly appreciated. The authors also thank the editor P. Bormann for valuable comments that helped to focus and improve the manuscript as well as his offer to make the SPEAR data set and user guidelines accessible via the NMSOP website.

## References

- Billings, S. D., M. S. Sambridge, and B. L. N. Kennett (1994). Errors in hypocenter location: picking, model, and magnitude dependence, *Bull. Seism. Soc. Am.* **84**, 1978-1990.
- Buland, R. (1976). The mechanics of locating earthquakes, *Bull. Seism. Soc. Am.* **66**, 173-187.
- Freedman, H. W. (1966a). The “little variable factor” a statistical discussion of the reading of seismograms, *Bull. Seism. Soc. Am.* **56**, 593-604.
- Freedman, H. W. (1966b). A statistical discussion of *Pn* residuals from explosions, *Bull. Seism. Soc. Am.* **56**, 677-695.
- Freedman, H. W. (1967a). Estimating the accuracy of source “Parameters”, *Bull. Seism. Soc. Am.* **57**, 373-379.
- Freedman, H. W. (1967b). A statistical discussion of *P* residuals from explosions: Part II, *Bull. Seism. Soc. Am.* **57**, 545-561.
- Freedman, H. W. (1968). Seismological measurements and measurement error, *Bull. Seism. Soc. Am.* **58**, 1261-1271.
- Gladwell, M. (2008). *Outliers: the story of success*, Little, Brown and Company, New York, 309.
- Leonard, M. (2000). Comparison of Manual and Automatic Onset Time Picking, *Bull. Seism. Soc. Am.* **90**, 1384-1390.
- Lomnitz, C. (2006). Three theorems of earthquake location, *Bull. Seism. Soc. Am.* **96**, 306-312.
- Pavlis, G. L. (1986). Appraising earthquake hypocenter location errors: a complete, practical approach for single-event locations, *Bull. Seism. Soc. Am.* **76**, 1699-1717.
- Pavlis, G. L. (1992). Appraising relative earthquake location errors, *Bull. Seism. Soc. Am.* **82**, 836-859.
- Sipkin, S. A., W. J. Person, and B. W. Presgrave (2000). Earthquake bulletins and catalogs at the USGS National Earthquake Information Center, *Incorporate Research Institutions for Seismology Newsletter*, **2000**, No. 1, 2-4.
- Steck, L. K., A. A. Velasco, A. H. Cogbill, and H. J. Patton (2001). Improving regional seismic event location in China, *Pure Appl. Geophys.* **158**, 211-240.
- Tull, J. E. (1987). SAC reference, tutorial guide for new users, Lawrence Livermore National Laboratory.
- Zeiler, C. P. And Velasco, A. A. (2009). Seismogram Picking Error from Analyst Review (SPEAR): Single-analyst and institution analysis, *Bull. Seism. Soc. Am.* **99**, 2759-2770.

| Topic   | Examples of interactive data analysis of seismic records<br>using the SEISAN software                                                                                                                                                                                                                                                           |
|---------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | <b>Lars Ottemöller</b> , Department of Earth Science, University of Bergen, Bergen, Norway, <a href="mailto:lars.ottemoller@geo.uib.no">lars.ottemoller@geo.uib.no</a><br><b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, Dept. 2, D-14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version | April 2013; DOI: 10.2312/GFZ.NMSOP-2_IS_11.6                                                                                                                                                                                                                                                                                                    |

|                                                                                                                               | page         |
|-------------------------------------------------------------------------------------------------------------------------------|--------------|
| <b>1 Introduction</b>                                                                                                         | <b>1</b>     |
| <b>2 Analysis of a local earthquake in southern Norway recorded between 50 km and 650 km epicentral distance</b>              | <b>2</b>     |
| <b>3 Analysis of a regional earthquake recorded at distance between 440 and 2950 km (<math>\Delta &lt; 30^\circ</math>)</b>   | <b>16</b>    |
| <b>4 Analysis of a shallow earthquake recorded at teleseismic distances (<math>30^\circ &lt; \Delta &lt; 90^\circ</math>)</b> | <b>27</b>    |
| <b>5 Analysis of an intermediate depth earthquake at teleseismic distances</b>                                                | <b>37</b>    |
| <b>6 Single station 3-component source location</b>                                                                           | <b>44</b>    |
| <b>7 Concluding remarks</b>                                                                                                   | <b>48</b>    |
| <b>Acknowledgment</b>                                                                                                         | <b>48</b>    |
| <b>References</b>                                                                                                             | <b>49-50</b> |

## 1 Introduction

The purpose of this information sheet is to illustrate the steps involved when analyzing seismic records by means of the SEISAN software. Examples are given for local, regional and teleseismic earthquakes. This is done by giving representative screen plots together with explanatory texts that follow the figures.

For details on the SEISAN software the reader is referred to section 11.4.2 (Data Analysis and Seismogram Interpretation - Software for routine analysis – SEISAN) and the references therein. Details of the software commands are not given here, and the reader is referred to the SEISAN manual for details, which can be downloaded either from NMSOP-2 via the link ‘Download programs and files’ or from <ftp://ftp.geo.uib.no/pub/seismo/SOFTWARE/SEISAN/>.

The events are included with the sample data that is available with SEISAN (from version 9.2).

The main background material provided by NMSOP-2 (<http://nmsop.gfz-potsdam.de>), explaining the methodology behind this information sheet, is:

- CHAPTER 2: Seismic Wave Propagation and Earth Models
- CHAPTER 3: Seismic Sources and Source Parameters
- CHAPTER 5: Seismic Sensors and their Calibration
- CHAPTER 6: Seismic Recording Systems
- CHAPTER 11: Data Analysis and Seismogram Interpretation
- IS 2.1: Standard nomenclature of seismic phases
- IS 3.3: The new IASPEI standards for determining magnitudes from digital data and their relation to classical magnitudes
- IS 11.1: Seismic source location
- IS 11.4: Tutorial for consistent phase picking at local to regional distances
- DS 2.1-2.4: Record examples for local, regional and teleseismic earthquakes and explosions

Note that there are in the following some, only apparent, differences between SEISAN internal nomenclature and procedures and those recommended by the IASPEI standards for measuring magnitudes which do not contradict. While SEISAN mb and IASPEI mb are identical, SEISAN mB corresponds to IASPEI mB\_BB. Similarly holds that SEISAN Ms = Ms\_20 and MS = Ms\_BB. Moreover, IASPEI standards recommend to measure mB\_BB and Ms\_BB on unfiltered broadband velocity traces, however accompanied by the statement that their passband should cover at least the period range within which these two magnitudes should be measured. SEISAN, however, always deconvolves any given actual record so as to simulate for mB\_BB a velocity-proportional broadband response in the period range between 0.2 and 30 s and for Ms\_BB a response in the period range between 3 and 60 s. This allows in SEISAN to read the broadband magnitudes also on short-period records, provided that the SNR is sufficient to allow a broadband velocity restitution.

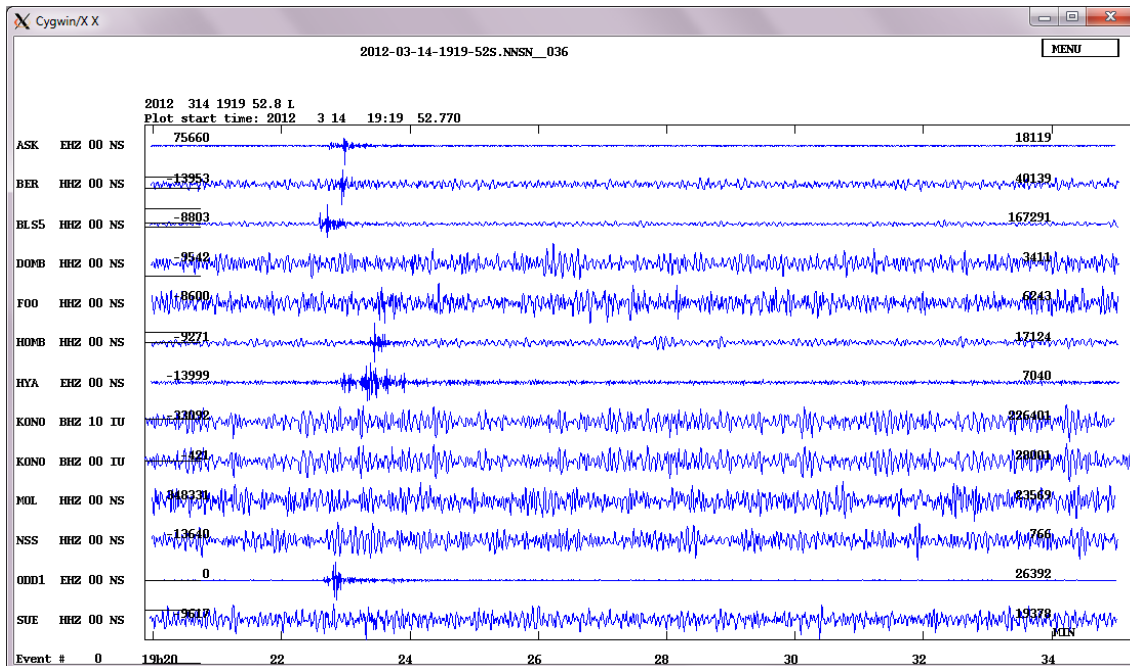
## 2 Analysis of a local earthquake in southern Norway recorded between 50 km and 650 km epicentral distance

The earthquake chosen here as example occurred in southern Norway and was recorded on the Norwegian National Seismic Network operated by the University of Bergen. The event was recorded on a mix of short-period and broadband seismic stations. It was automatically detected and then interactively processed.

The earthquake parameters are:

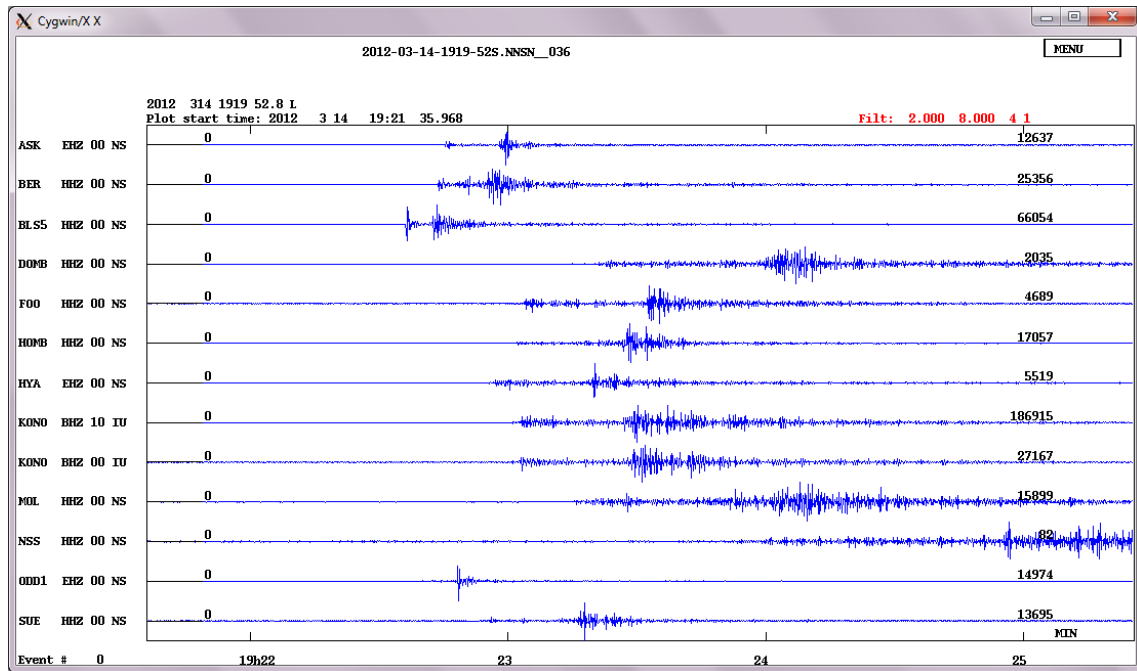
|              |                     |
|--------------|---------------------|
| Date:        | 14 Mar 2012         |
| Origin time: | 19:22:27.9<br>(UTC) |
| Latitude:    | 59.526°N            |
| Longitude:   | 5.615°E             |
| Depth:       | 13.9 km             |
| ML:          | 3.4                 |

We now show the step-by-step procedure to analyze the earthquake.

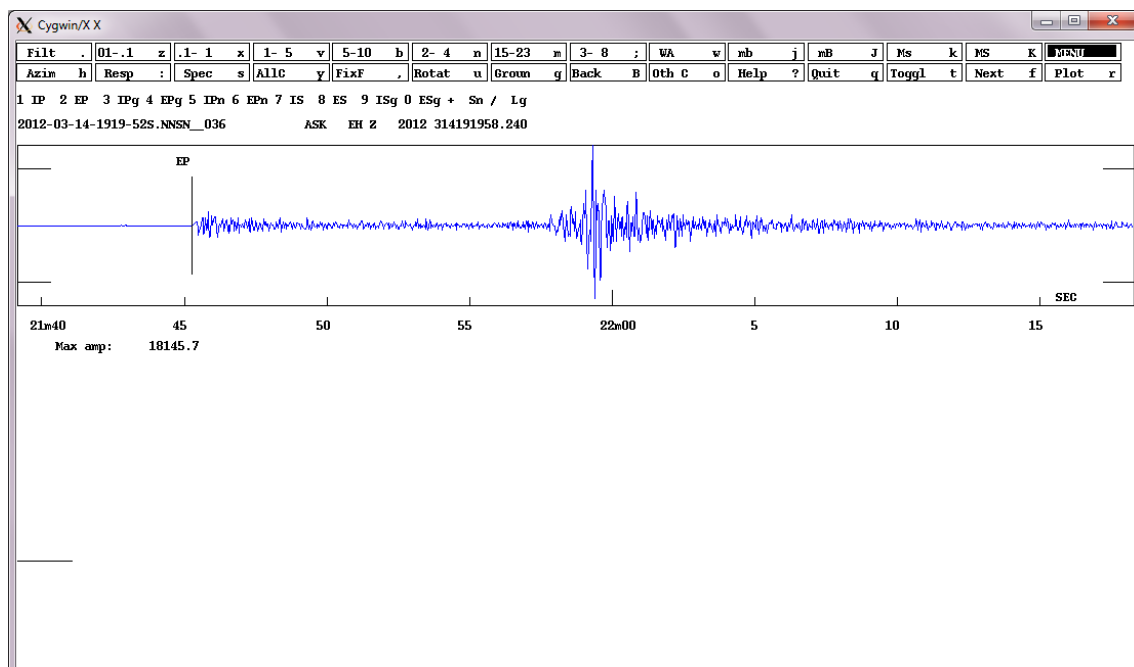


*Vertical-component traces, unfiltered:* Our starting point is a waveform file that is produced by an automatic detection system. No additional information is given, although, depending on the automated processing system, additional information including type of event, location and magnitude may already be available. It is our first task to identify the type of event that is recorded. If the detection is not false, the initial choices are between local, regional and teleseismic event. It may also be possible to identify if the event is likely an explosion based on the characteristics of the waveforms and the pattern of phase arrival times. The above multi-trace plot shows records of a small local earthquake, characterized by the short signal duration of less than about 1 min and a relatively high frequency signal content.

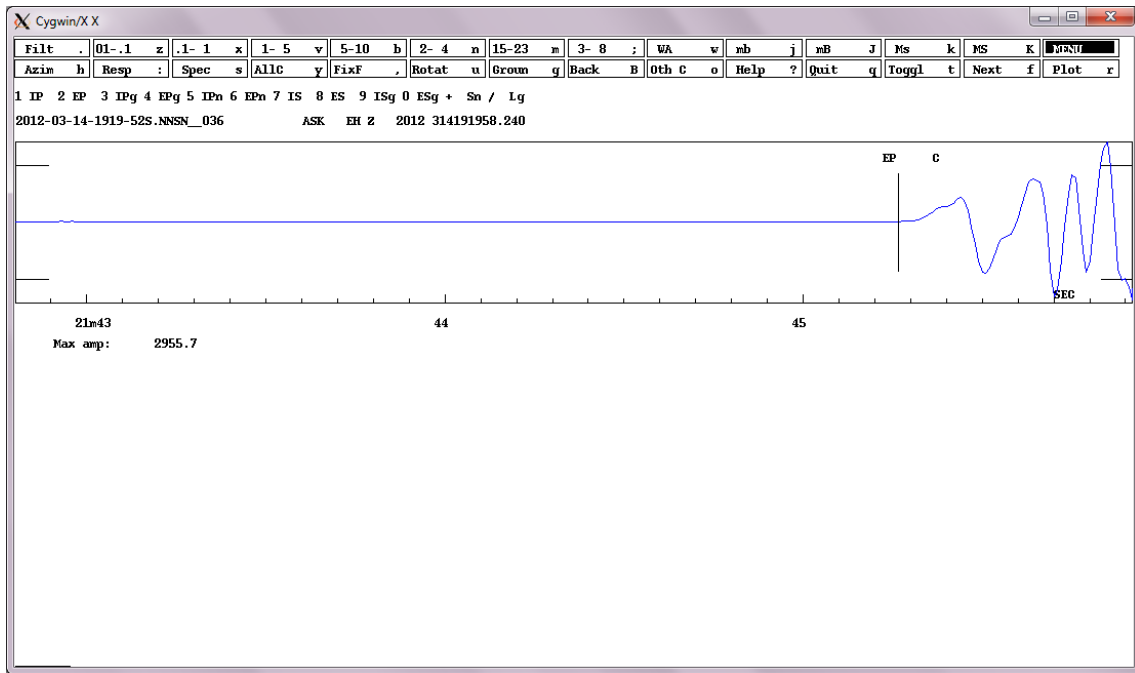
The event is recorded on 13 seismic stations. The numbers on the SEISAN multi-trace plot gives the DC of individual traces on the left and the maximum zero to peak amplitude on the right. The units are counts unless instrument correction is used to compute displacement, velocity or acceleration with the units nm (nanometers =  $10^{-9}$  m), nm/s and nm/s<sup>2</sup>, respectively. Station and component codes are given left of each trace together with the location code and network code. For the first trace, the station is ASK, the component is EHZ (short-period seismometer, with high sampling rate of more than 80 samples per second), the location code is 00 (used to indicate differences between equipment that gives the same component code at the same site, for example an STS2 at 40sps is labeled BHZ.00 and an STS1 at 20sps is labeled BHZ.10) and the network is NS (indicating who operates the station, in this case NS = Norwegian National Seismic Network). The event is well visible on the short-period seismometer records, which have the component code EHZ. However, it is not seen on the broadband traces, which are labeled HHZ, as this is a small earthquake where the signal content is mostly above 0.5 Hz.



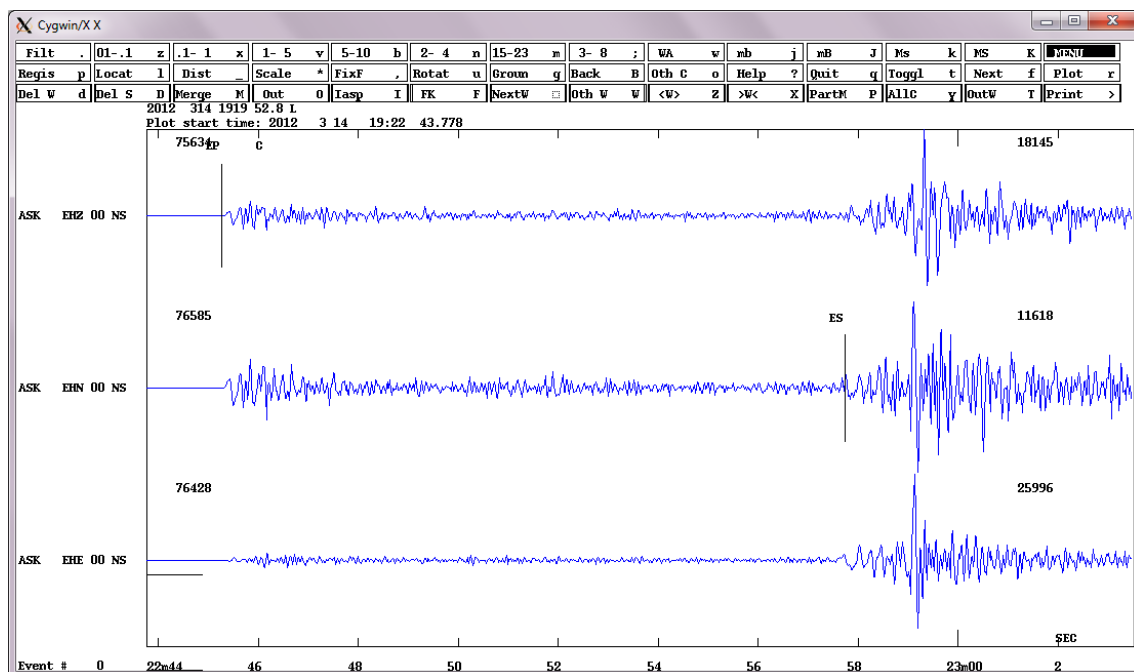
*Vertical-component traces, filtered 2-8 Hz:* Applying this high-frequency band-pass filter, which suppresses the more long-period microseismic noise in the broadband records, the earthquake is now clearly seen on all the traces. As compared to the previous plot, the time window of signal arrivals has been cut out and the time scale been stretched. With distinct primary P- and secondary S-wave arrivals in this frequency band it is very clear that this is a local earthquake. We can see from the arrival times that the earthquake was closest to station BLS5. In SEISAN we then put this event into the database and start the processing.



*Single trace plot of station ASK at epicentral distance of 109 km, unfiltered:* In SEISAN, we mostly do the phase picks on single trace plots, like shown here, or on three-component trace plots. For this recording on a short-period seismometer it is not necessary to filter the data in order to pick phase arrivals. The P arrival is marked.

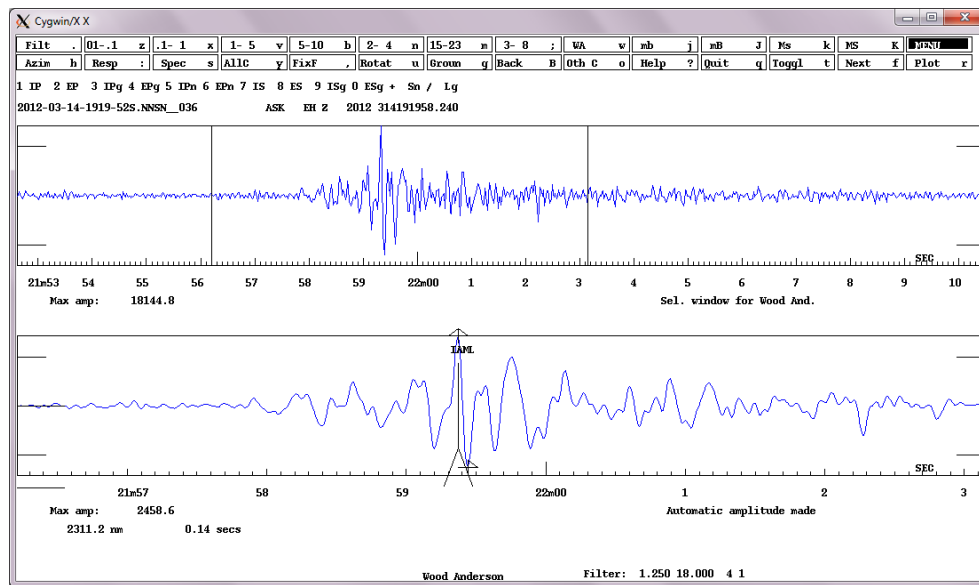


*Single-trace plot of station ASK at epicentral distance of 109 km, unfiltered:* This is the same station as on the previous plot, but this time we have zoomed in on the P arrival. It is now possible to read the P arrival time more accurately, and we can also identify the polarity of the first arrival, which is compression (upward motion on the vertical component) in this case. The phase arrivals will be used for the earthquake location (see IS 11.1) and the polarity can be used to determine the fault plane solution (see also EX 3.2).



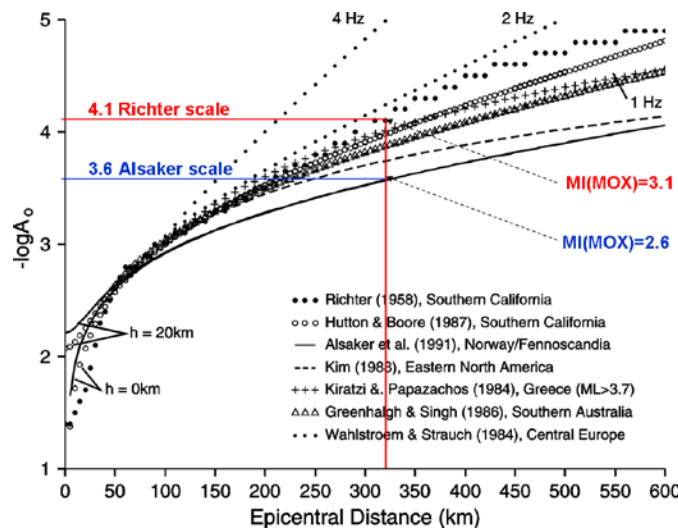
*Three-component trace plot of station ASK at epicentral distance of 109 km, unfiltered:* We normally read the S arrival on one of the two horizontal channels and not on the vertical as SP conversions may arrive prior to S and result in an incorrect reading of the onset time.



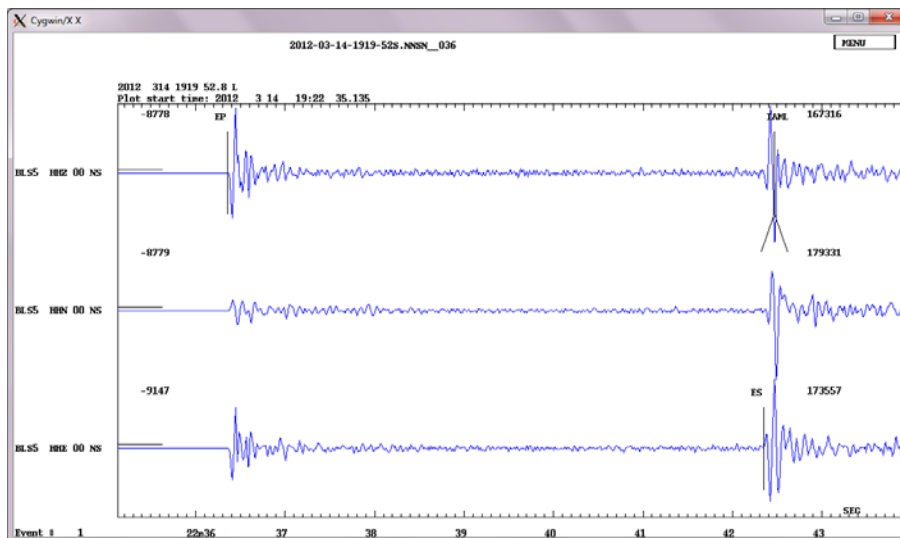


*Single-trace plots of station ASK at epicentral distance of 109 km, Wood-Anderson filter:* To read the ML amplitude, the instrument response is removed and a filter equivalent to the revised Wood-Anderson seismograph response according to Uhrhammer and Collins (1990) (see also IS 3.3) is applied. The bottom trace is a zoomed-in version of the area between the two vertical lines on the top trace. In SEISAN the amplitude is given in units of nanometer. It is measured peak-to-adjacent trough of the maximum deflection of the seismogram trace, where peak and trough are separated by one crossing of the zero-line (sometimes described as “peak-to-peak amplitude”) but stored in the database is peak-to-peak/2 according to the IASPEI (2013) standard. This maximum amplitude is normally read from the S or Lg waves. It is stored as the IAML amplitude phase. ‘I’ stands for IASPEI standard, assuring that the standard Wood-Anderson filter response and the recommended amplitude measurement procedure has been applied (see IS 3.3), and ‘A’ stands for displacement amplitude in nm. IASPEI standard for ML is to measure A on horizontal components. In Norway, however, it is routine practice to use the vertical component for the amplitude reading of ML and the scale given by Alsaker et al. (1991; see also DS 3.1, Tab. 2). This scale was derived for Norway and scaled so that its results match at 60 km distance with those of the horizontal component California standard formula for equal IAML input (see next figure and recommendations in IS 3.3). The default ML formula implemented in SEISAN is that of Hutton and Boore (1987; see next figure and formula in DS 3.1, Tab. 2), on which also the IASPEI standard formula for ML is based (see IS 3.3). If, however, this default procedure is used in other regions with other attenuation conditions then the calculated ML values may differ from standard ML. DS 3.1. Table 2, offers several ML calibration functions for other regions together with comments on their scaling, components to be used and the distance range of their applicability. You may implement in SEISAN either your own properly to the ML standard scaled local calibration function, or, if not yet available, one of those which you believe to match best with the seismotectonic conditions in the area. Note, however, that local ML data based on not yet tested and proven standard scaling should not be published with the nomenclature for standard ML. That is, why we use in NMSOP *M<sub>l</sub>* as the general nomenclature for local magnitude, and *M<sub>L</sub>* only for *M<sub>l</sub>* that has been properly scaled to the new standard. Otherwise, users of local magnitude data do not know whether or not they have been correctly scaled according to the now globally set standard. E.g., according to Braunmiller et al. (2005) (see also section 3.2.9.6 in the forthcoming new Chapter 3) the *M<sub>l</sub>* scales used by various agencies in Southern Germany, France, Italy and Switzerland yield results for equal input data that differ on average between 0.2 to 0.6 m.u. when scaled to *M<sub>w</sub>*.

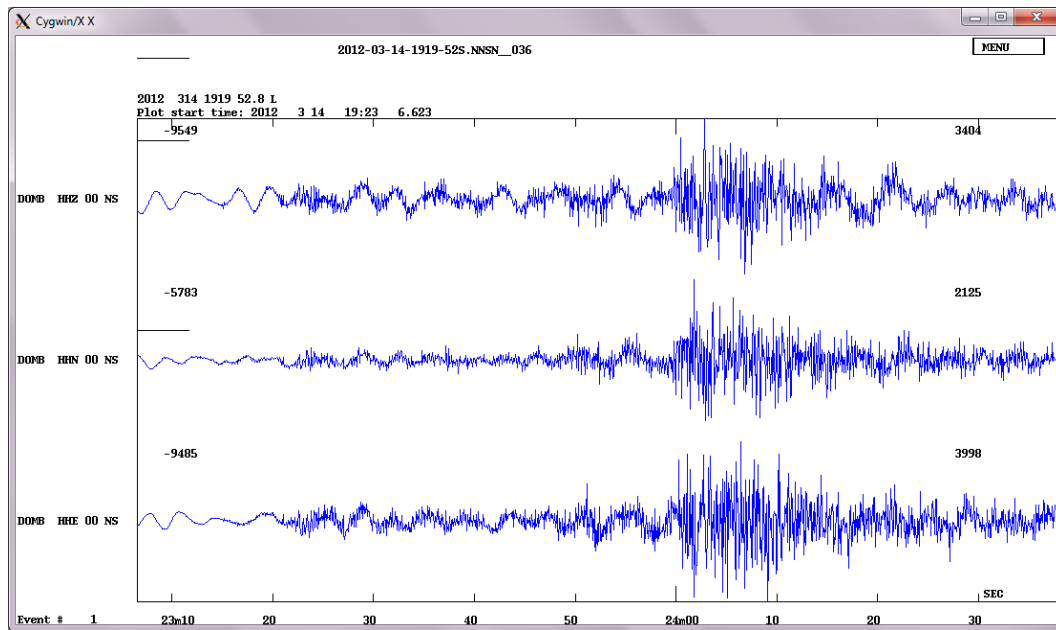
Some examples for other regional ML calibration functions have been plotted in the diagram below. Note that the Alsaker et al. (1991) scale (lowermost full line) for the continental shield areas of Scandinavia and Eurasia reveals significantly lower attenuation when compared with Southern California calibration curves of both Richter (1935) and Hutton and Boore (1987) (see full dot and open dot curves in the Figure below). Accordingly, since properly scaled to the latter at about 60 km epicentral distance, the Alsaker et al. calibration curve yields at several 100 km distance for equal amplitude input data already ML values that are several tenths of magnitude units smaller than those derived by applying the original Richter scale. This difference increases to 1.7 m.u. at a distance of 1500 km, up to which the Alsaker et al. (1991) scale is defined (although strictly Hutton and Boore (1987) is limited to 600 km).



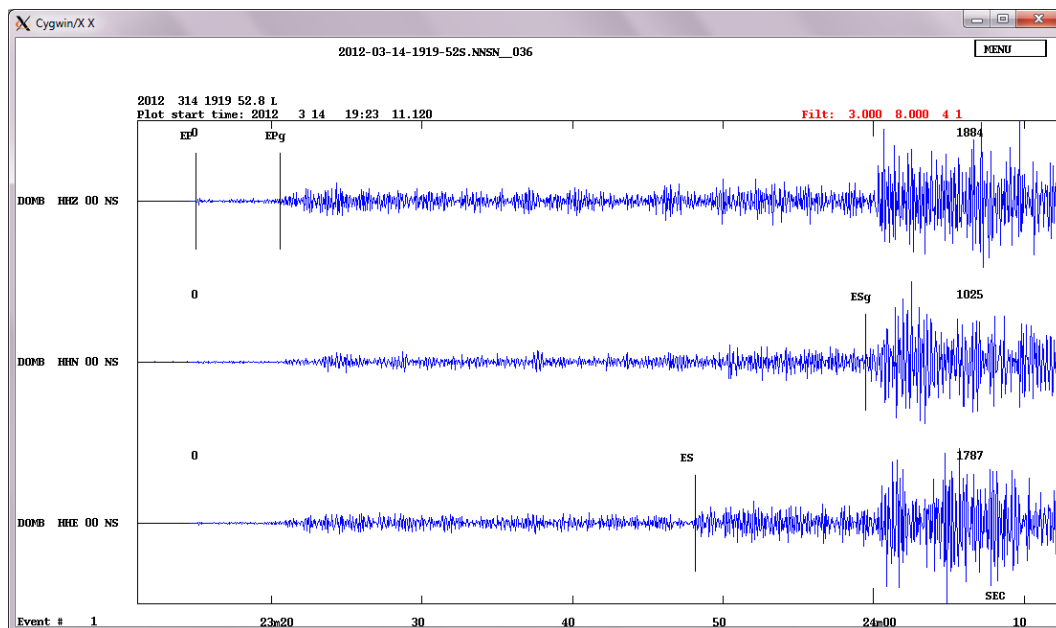
We now repeat the same procedure of measuring IAML for the other stations, however show only examples from some selected stations to explain specific aspects of the data processing.



*Three-component traces of station BLS5 at epicentral distance of 49 km, unfiltered:* Plotted are the three-component broadband traces from the nearest station. P and S arrivals, and the ML amplitude were read. In this case the SNR was high enough so that it was not necessary to filter the broadband recordings to identify the phase arrivals.

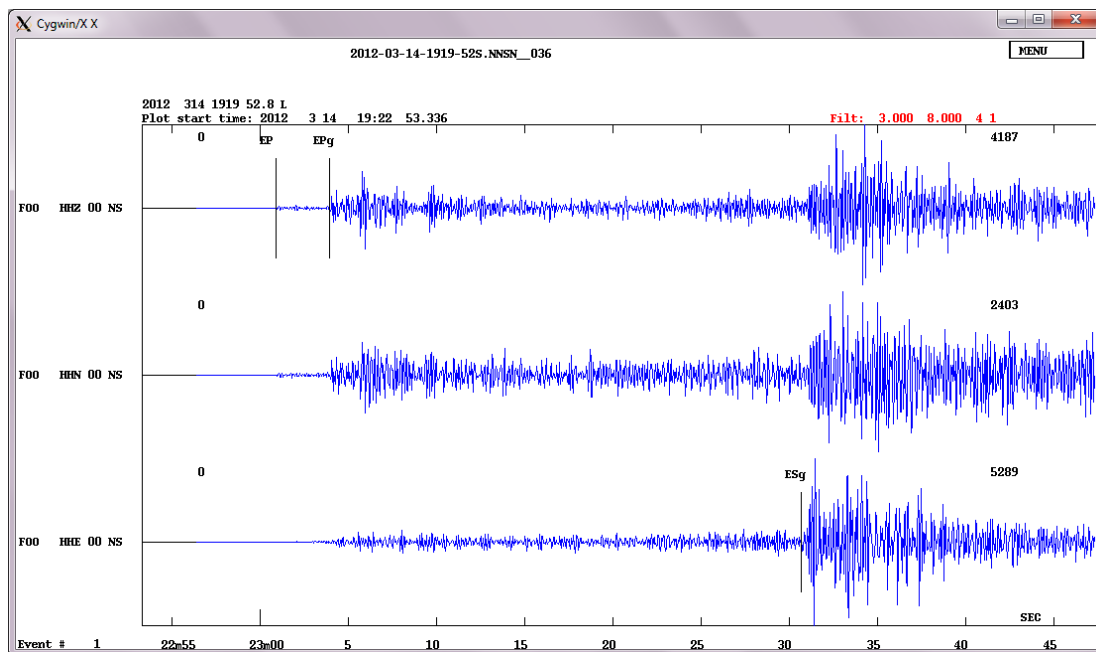


Three-component traces of station DOMB at epicentral distance of 342 km, unfiltered: Going to larger distances, the signal amplitudes at higher frequencies decrease relative to the more low-frequency microseismic noise, which peaks around 5 seconds. Here, we would normally filter the traces to read the phase arrivals.

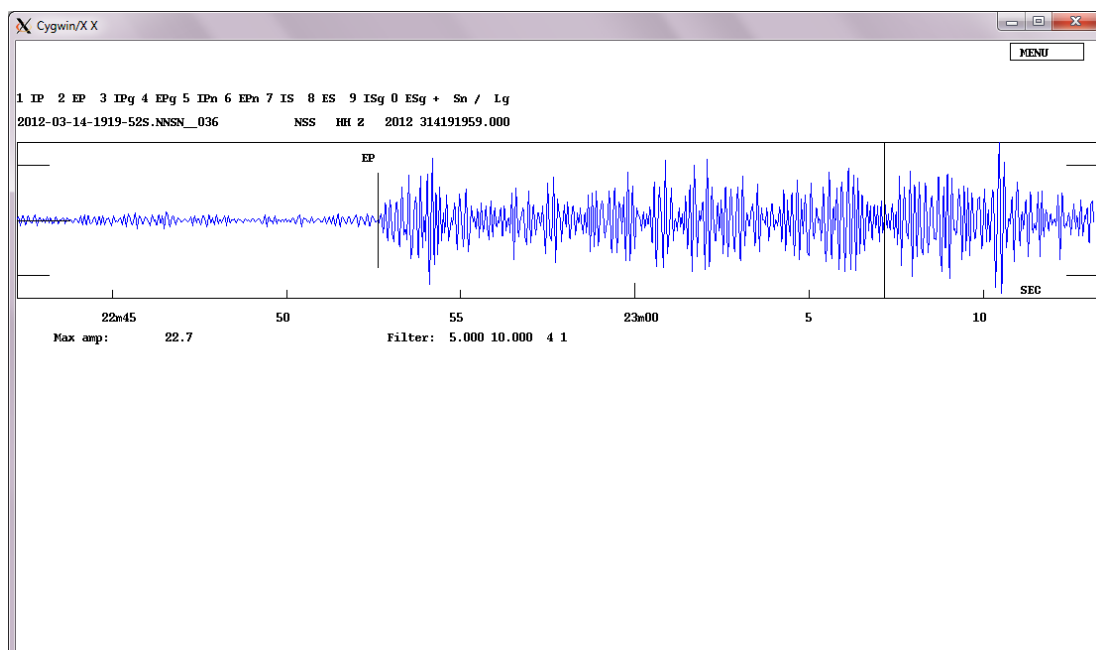


Three-component traces of station DOMB at epicentral distance of 342 km, filtered 3-8 Hz: For this station, it is possible to read besides the larger amplitude P and S arrivals, termed Pg and Sg, additionally the commonly weaker but earlier arrivals Pn and Sn. In SEISAN we generally mark the first P- and S-wave arrivals as P and S, respectively. The location program then determines which phase arrives first. Sometimes, Sn (here marked as ES) is weak and not recognizable in the signal-generated noise of scattered P waves. Then one has to mark Sg as such, since the location program otherwise would use it as the first arriving S (Sn in this case).

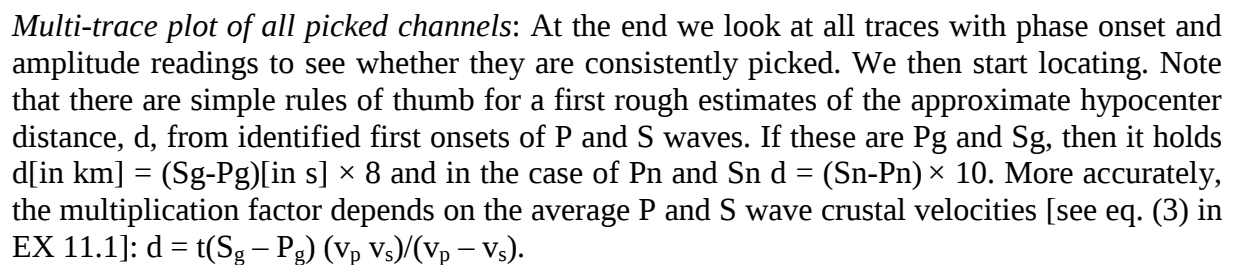
Reading different phases with different take-off angle from the source, such as Pg/Pn and Sg/Sn, significantly improves the depth determination. However, one should be quite certain that these identifications are correct. This requires fairly clear arrivals, an experienced analyst and that the phases match the solution when locating. For standard nomenclature, nature and travel paths of crustal phases see IS 2.1 and for many more record examples DS 11.1.



Three-component traces of station FOO at epicentral distance of 233 km, filtered 3-8 Hz: This is an example where Sn cannot be read, and S has to be read as Sg.



Single-trace plot of station NSS at epicentral distance of 649 km, filtered 5-10 Hz: Going to larger distances, the P arrival becomes more emergent and would not be read without applying a filter.



|                                                        |      |        |          |         |         |       |      |        |       |       |       |      |      |      |   |   |
|--------------------------------------------------------|------|--------|----------|---------|---------|-------|------|--------|-------|-------|-------|------|------|------|---|---|
| #                                                      | 1    | 14     | Mar      | 2012    | 19:22   | 28    | L    | 59.526 | 5.615 | 13.9  | 0.4   | 3.4  | LBER | 12   | ? | 1 |
| date                                                   | hrmn | sec    | lat      | long    | depth   | no    | m    | rms    | damp  | erln  | erlt  | erdp |      |      |   |   |
| 12 314                                                 | 1922 | 28.19  | 5931.58N | 5 36.9E | 13.9    | 29    | 3    | 0.36   | 0.000 | 4.7   | 2.0   | 3.6  |      |      |   |   |
| stn                                                    | dist | azm    | ain      | w phas  | calcphs | hrmn  | tsec | t-obs  | t-cal | res   | wt    | di   |      |      |   |   |
| BLS5                                                   | 49   | 103.3  | 96.7     | 0 S     | SG      | 1922  | 42.3 | 14.14  | 14.18 | -0.04 | 1.00  | 7    |      |      |   |   |
| BLS5                                                   | 49   | 103.3  | 96.7     | 0 P     | D PG    | 1922  | 36.4 | 8.15   | 8.15  | 0.00  | 1.00  | 4    |      |      |   |   |
| BLS5                                                   | 49   | 103.3  |          | 0 IAML  |         | 1922  | 42.5 | 14.3   |       |       |       |      |      |      |   |   |
| BLS5                                                   | 49   | 103.3  |          | 0 AMPG  |         | 1922  | 36.4 | 8.2    |       |       |       |      |      |      |   |   |
| BLS5                                                   | 49   | 103.3  |          | 0 AMSG  |         | 1922  | 42.4 | 14.2   |       |       |       |      |      |      |   |   |
| ODD1                                                   | 71   | 52.6   | 93.0     | 0 S     | SG      | 1922  | 48.0 | 19.81  | 20.02 | -0.21 | 1.00  | 10   |      |      |   |   |
| ODD1                                                   | 71   | 52.6   | 93.0     | 0 P     | D PG    | 1922  | 39.8 | 11.54  | 11.51 | 0.04  | 1.00  | 1    |      |      |   |   |
| ODD1                                                   | 71   | 52.6   |          | 0 IAML  |         | 1922  | 48.3 | 20.1   |       |       |       |      |      |      |   |   |
| BER                                                    | 97   | 350.7  | 91.8     | 0 S     | SG      | 1922  | 55.0 | 26.75  | 26.70 | 0.05  | 1.00  | 4    |      |      |   |   |
| BER                                                    | 97   | 350.7  | 91.8     | 0 P     | C PG    | 1922  | 43.4 | 15.22  | 15.35 | -0.12 | 1.00  | 2    |      |      |   |   |
| BER                                                    | 97   | 350.7  |          | 0 IAML  |         | 1922  | 56.9 | 28.7   |       |       |       |      |      |      |   |   |
| ASK                                                    | 109  | 347.7  | 91.5     | 0 S     | SG      | 1922  | 57.7 | 29.52  | 29.95 | -0.43 | 0.98* | 4    |      |      |   |   |
| ASK                                                    | 109  | 347.7  | 91.5     | 0 P     | C PG    | 1922  | 45.3 | 17.06  | 17.21 | -0.15 | 0.98* | 3    |      |      |   |   |
| ASK                                                    | 109  | 347.7  |          | 0 IAML  |         | 1922  | 59.4 | 31.2   |       |       |       |      |      |      |   |   |
| SUE                                                    | 177  | 344.9  | 90.8     | 0 Sg    | SG      | 1923  | 16.2 | 48.02  | 47.82 | 0.20  | 0.87* | 3    |      |      |   |   |
| SUE                                                    | 177  | 344.9  | 55.1     | 0 P     | C PN4   | 1922  | 53.8 | 25.56  | 25.93 | -0.37 | 0.87* | 4    |      |      |   |   |
| SUE                                                    | 177  | 344.9  | 90.8     | 0 Pg    | D PG    | 1922  | 55.8 | 27.61  | 27.49 | 0.13  | 0.87* | 2    |      |      |   |   |
| SUE                                                    | 177  | 344.9  |          | 0 IAML  |         | 1923  | 17.7 | 49.5   |       |       |       |      |      |      |   |   |
| HYA                                                    | 185  | 9.4    | 55.1     | 0 S     | SN4     | 1923  | 16.1 | 47.86  | 46.91 | 0.95  | 0.86* | 10   |      |      |   |   |
| HYA                                                    | 185  | 9.4    | 90.7     | 0 Sg    | SG      | 1923  | 19.0 | 50.78  | 50.00 | 0.78  | 0.86* | 3    |      |      |   |   |
| HYA                                                    | 185  | 9.4    | 55.1     | 0 P     | C PN4   | 1922  | 55.3 | 27.11  | 26.96 | 0.15  | 0.86* | 2    |      |      |   |   |
| HYA                                                    | 185  | 9.4    |          | 0 IAML  |         | 1923  | 20.1 | 51.9   |       |       |       |      |      |      |   |   |
| HYA                                                    | 185  | 9.4    | 90.7     | 0 Pg    | PG      | 1922  | 57.1 | 28.85  | 28.74 | 0.12  | 0.86* | 0    |      |      |   |   |
| HOMB                                                   | 217  | 128.8  | 90.6     | 0 Sg    | SG      | 1923  | 26.8 | 58.57  | 58.49 | 0.08  | 0.80* | 6    |      |      |   |   |
| HOMB                                                   | 217  | 128.8  | 90.6     | 0 Pg    | C PG    | 1923  | 2.1  | 33.86  | 33.62 | 0.25  | 0.80* | 4    |      |      |   |   |
| HOMB                                                   | 217  | 128.8  |          | 0 IAML  |         | 1923  | 27.2 | 59.0   |       |       |       |      |      |      |   |   |
| HOMB                                                   | 217  | 128.8  | 55.1     | 0 P     | C PN4   | 1922  | 59.4 | 31.24  | 30.96 | 0.28  | 0.80* | 6    |      |      |   |   |
| KONO                                                   | 225  | 84.8   | 55.1     | 0 P     | PN4     | 1923  | 0.3  | 32.12  | 31.95 | 0.17  | 0.79* | 3    |      |      |   |   |
| KONO                                                   | 225  | 84.8   | 90.6     | 0 Pg    | D PG    | 1923  | 2.9  | 34.70  | 34.81 | -0.11 | 0.79* | 1    |      |      |   |   |
| KONO                                                   | 225  | 84.8   | 90.6     | 0 Sg    | SG      | 1923  | 28.5 | 60.31  | 60.58 | -0.26 | 0.79* | 5    |      |      |   |   |
| KONO                                                   | 225  | 84.8   |          | 0 IAML  |         | 1923  | 30.2 | 62.0   |       |       |       |      |      |      |   |   |
| FOO                                                    | 233  | 352.5  | 90.6     | 0 Sg    | SG      | 1923  | 30.7 | 62.45  | 62.57 | -0.12 | 0.78* | 2    |      |      |   |   |
| FOO                                                    | 233  | 352.5  | 55.1     | 0 P     | PN4     | 1923  | 0.9  | 32.71  | 32.88 | -0.17 | 0.78* | 2    |      |      |   |   |
| FOO                                                    | 233  | 352.5  | 90.6     | 0 Pg    | PG      | 1923  | 3.9  | 35.69  | 35.96 | -0.27 | 0.78* | 1    |      |      |   |   |
| FOO                                                    | 233  | 352.5  |          | 0 IAML  |         | 1923  | 32.5 | 64.3   |       |       |       |      |      |      |   |   |
| DOMB                                                   | 342  | 32.3   | 55.1     | 0 S     | SN4     | 1923  | 48.1 | 79.93  | 80.82 | -0.89 | 0.60* | 6    |      |      |   |   |
| DOMB                                                   | 342  | 32.3   | 90.4     | 0 Sg    | SG      | 1923  | 59.5 | 91.25  | 91.28 | -0.03 | 0.60* | 3    |      |      |   |   |
| DOMB                                                   | 342  | 32.3   | 55.1     | 0 P     | C PN4   | 1923  | 14.9 | 46.70  | 46.45 | 0.26  | 0.60* | 1    |      |      |   |   |
| DOMB                                                   | 342  | 32.3   | 90.4     | 0 Pg    | PG      | 1923  | 20.6 | 52.35  | 52.46 | -0.11 | 0.60* | 0    |      |      |   |   |
| MOL                                                    | 355  | 16.3   | 55.1     | 0 P     | PN4     | 1923  | 15.4 | 47.23  | 48.03 | -0.80 | 0.57* | 1    |      |      |   |   |
| MOL                                                    | 355  | 16.3   |          | 0 IAML  |         | 1924  | 8.9  | 100.7  |       |       |       |      |      |      |   |   |
| NSS                                                    | 649  | 28.0   | 53.1     | 0 P     | PN5     | 1923  | 52.6 | 84.43  | 83.82 | 0.61  | 0.08* | 0    |      |      |   |   |
| ODD1                                                   | EZ   | hdist: | 72.6     | amp:    | 4529.1  | T:    | 0.4  | ml =   | 3.7   |       |       |      |      |      |   |   |
| BER                                                    | HZ   | hdist: | 97.8     | amp:    | 1863.0  | T:    | 0.4  | ml =   | 3.5   |       |       |      |      |      |   |   |
| ASK                                                    | EZ   | hdist: | 109.9    | amp:    | 2311.9  | T:    | 0.1  | ml =   | 3.6   |       |       |      |      |      |   |   |
| SUE                                                    | HZ   | hdist: | 177.5    | amp:    | 1124.8  | T:    | 0.1  | ml =   | 3.6   |       |       |      |      |      |   |   |
| HYA                                                    | EZ   | hdist: | 185.5    | amp:    | 629.6   | T:    | 0.4  | ml =   | 3.4   |       |       |      |      |      |   |   |
| HOMB                                                   | HZ   | hdist: | 218.4    | amp:    | 1355.1  | T:    | 0.4  | ml =   | 3.8   |       |       |      |      |      |   |   |
| KONO                                                   | BZ   | hdist: | 225.4    | amp:    | 144.8   | T:    | 0.2  | ml =   | 2.8   |       |       |      |      |      |   |   |
| MOL                                                    | HZ   | hdist: | 355.3    | amp:    | 112.3   | T:    | 0.1  | ml =   | 3.0   |       |       |      |      |      |   |   |
| FOO                                                    | HZ   | hdist: | 233.4    | amp:    | 388.3   | T:    | 0.4  | ml =   | 3.3   |       |       |      |      |      |   |   |
| BLS5                                                   | HZ   | hdist: | 50.9     | amp:    | 6142.4  | T:    | 0.1  | ml =   | 3.7   |       |       |      |      |      |   |   |
| BLS5                                                   | HZ   | gdist: | 50.9     | mom:    | 13.5    | mw =  | 2.9  |        |       |       |       |      |      |      |   |   |
| FOO                                                    | HZ   | gdist: | 231.4    | mom:    | 13.6    | mw =  | 3.0  |        |       |       |       |      |      |      |   |   |
| Number of spectra available and number used in average |      |        |          |         |         |       |      |        |       | 2     | 2     |      |      |      |   |   |
| 2012                                                   | 314  | 1922   | 28.2     | L       | 59.526  | 5.615 | 13.9 | BER    | 12    | 0.4   | 3.4   | LBER | 3.0  | WBER |   |   |

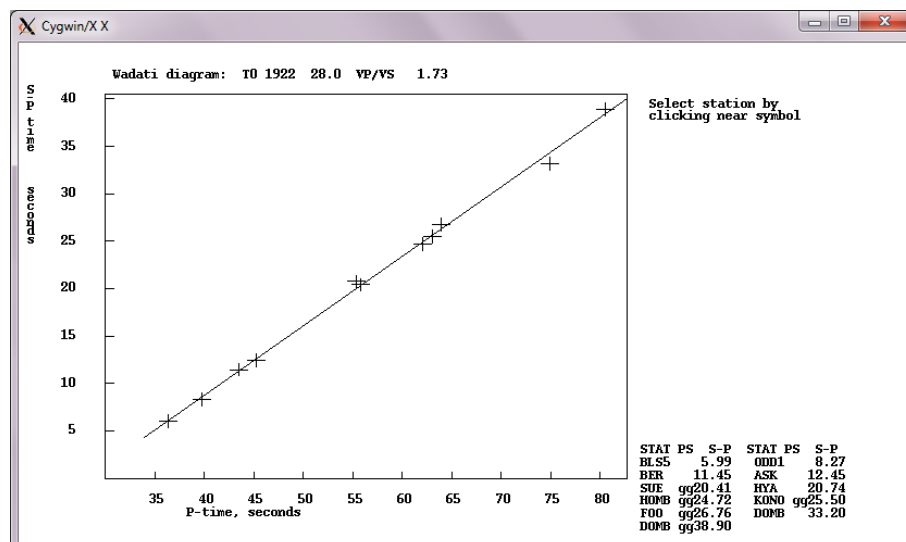
*Output from location program:* We use a standard linearized inversion program (HYPOCENTER, Lienert et al. (1986, 1995)) to locate the event that makes use of the common seismic phases at local distances. The location program tries to minimize the differences between the observed (t-obs) and calculated arrival times (t-cal) that are computed for a 1D layered model. The residuals are given in the column labeled 'res'. For our example event these travel-time residuals are relatively small. Note that for station DOMB the residuals are given for both Pn/Pg and Sn/Sg and that our interpretation fits the solution.

Direct (Pg and Sg) and critically refracted wave types (Pn and Sn) were very clearly visible from this earthquake on the records of several stations. This helped to constrain the earthquake depth, which was estimated to be 13.9 km (see last but one line). At the end the location procedure the program gives the magnitude values for the individual components and stations, and calculates the average, which is 3.4 in this case. The text 'LBER' refers to the magnitude type (L for ML) and the reporting agency (BER = Bergen). Some spread in the magnitude residuals (difference between single-component magnitude and average) is expected and mostly due to local site effects. In our case the largest residual are -0.6 and +0.4 magnitude units.

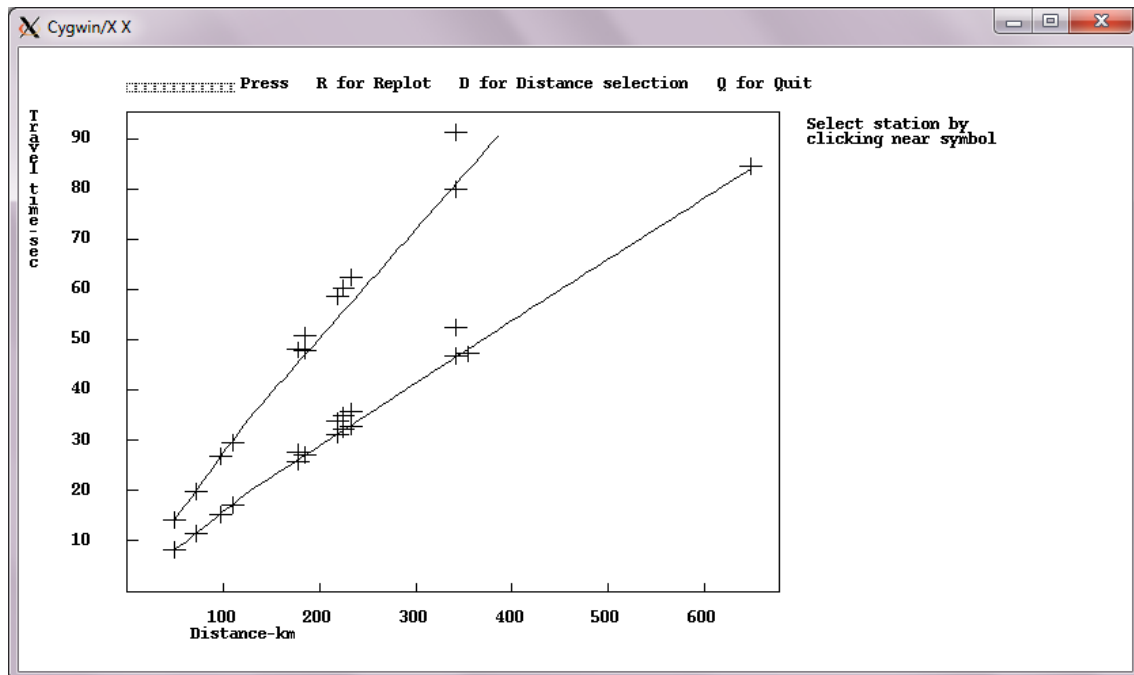
A number of polarities were read, 7 compressional (C) and 4 dilatational (D), which are given in the column after the phase name read and the phase used by the location program. For example, in the second phase line of the location output for stations BLS5, the first phase arrival was read as P, the location program decides that this is the direct Pg arrival based on the hypocentral distance. The polarity is read as D, standing for down or dilatation. The location program computes the azimuth (azm) and angle of incidence (ain), which in this example are 103.3° and 96.7°, respectively. The polarity data can be used to determine the fault plane solution if there are sufficient observations. In this example we can not find a robust solution and do not show the procedure here.

The velocity model used is given by the following table:

| <i>Depth of top of layer in km</i> | <i>P-wave velocity in km/s</i> |
|------------------------------------|--------------------------------|
| 0                                  | 6.2                            |
| 12                                 | 6.6                            |
| 23                                 | 7.1                            |
| 31                                 | 8.05                           |
| 50                                 | 8.25                           |
| 80                                 | 8.5                            |



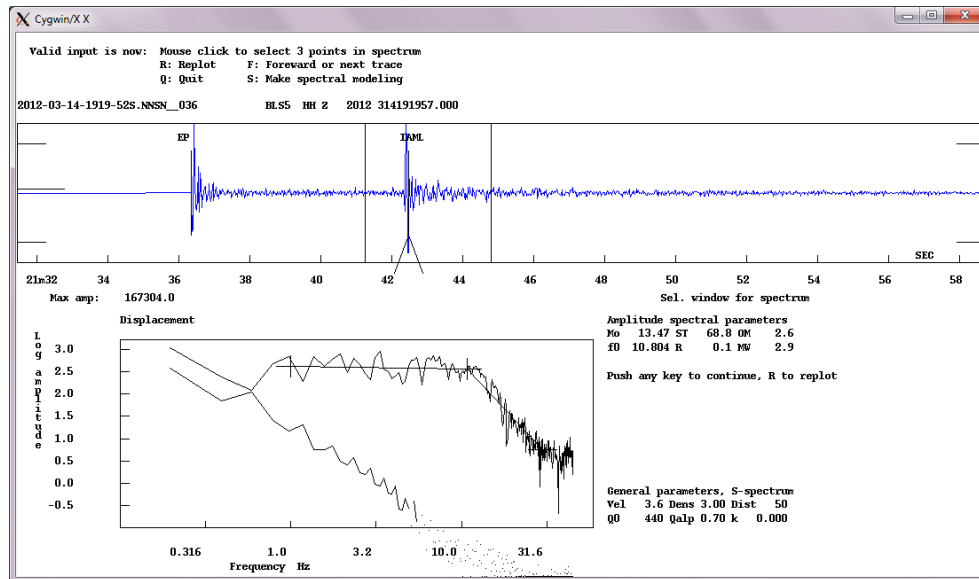
**Wadati diagram:** To check our readings we can plot the Wadati diagram (plotting S-P times against absolute P arrival times (labels on x-axis given to show scaling), and the data points should fall onto a line like in this example. Data points (+) that are significantly off the line should be checked and re-read.



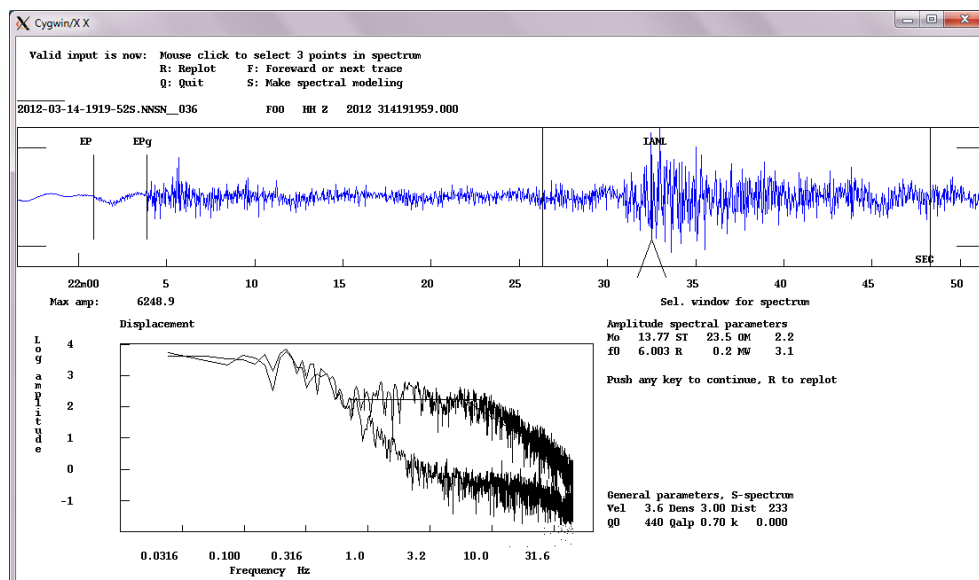
*Travel time plot:* We can also produce a simple travel-time plot that shows the observations (+) together with the calculated travel-time curves for P and S first arrivals according to the local velocity model implemented in the SEISAN location program (solid lines). For distances greater than about 150 km, our Pg (around the lower curve) and Sg arrivals (around the upper curve) come in mostly after the predicted first arrivals.

Note that SEISAN uses layered 1D velocity models and the user needs to enter the model for his or her respective area. SEISAN allows the use of a number of 1D models for events depending on their location to account for changes in structure within the area of interest or network coverage.

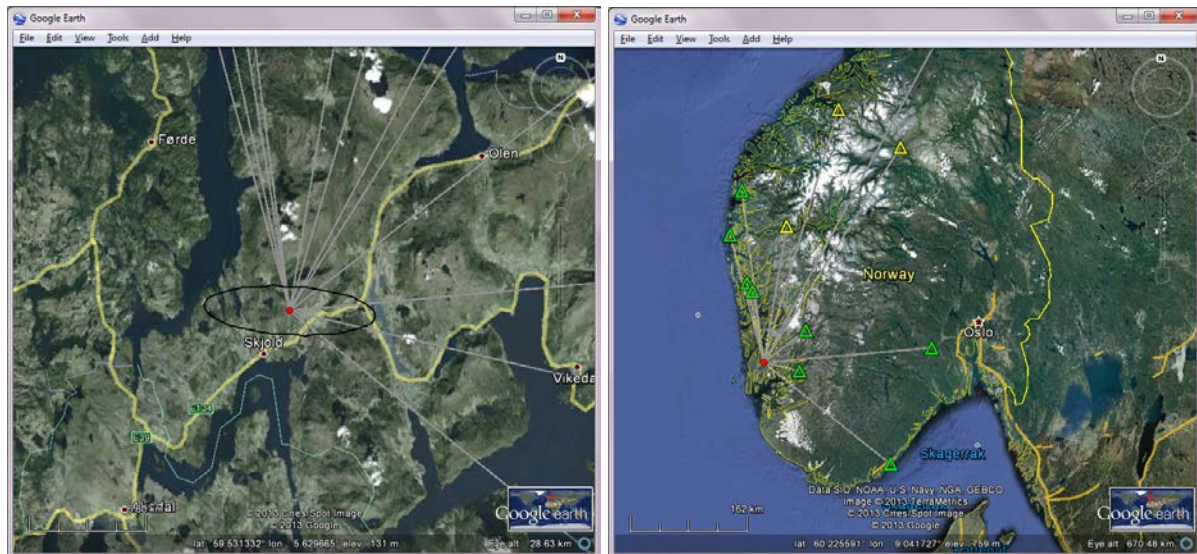




Source spectrum plot for station BLS5 at epicentral distance of 49 km: We compute the source displacement spectrum (upper curve in the bottom diagram above the noise spectrum) for the vertical component of this station. The spectrum is corrected for attenuation, which for Norway is given by  $Q(f) = 440 \times f^{0.7}$  (see data at bottom right). From the spectrum we can approximately read the long-period spectral level, which is translated via the assumed velocity and density model into seismic moment (here  $\log_{10}M_0 = 13.47$  Nm) and moment magnitude ( $M_W = 2.9$ ). We also measure the corner frequency ( $f_c = 10.8$  Hz), which is translated into stress drop  $\Delta\sigma = 68.8$  bars = 6.88 MPa using  $\Delta\sigma = (7/16) M_0/R^3$  (Eshelby, 1957).



Source spectrum plot for station FOO at epicentral distance of 233 km: The spectral analysis is normally done for a number of stations with good SNR. When comparing the results for this station with those of the previous station (BLS5) one finds the difference between the moment magnitudes to be small, but the stress drop difference to be large because of the reading uncertainty in  $f_c$  and  $\Delta\sigma \sim f_c^3$ . (For related formulas and discussions see EX 3.4). At the end, the averages are computed from all observations and standard deviations are calculated.



*Location maps in GoogleEarth:* SEISAN creates files that make it possible to look at the location with GoogleEarth immediately. The figure on the left shows the epicenter area with the error ellipse plotted as a solid black line. The extent of the 90 percent confidence error ellipse is about 2.5 by 9.0 km. It is also possible to compare the epicenter location to known fault structures. In the case of explosions we try to see if the event is located close to a known mine or quarry.

On the right, we show a more regional GoogleEarth map, which shows the epicenter as a red dot. The stations are color coded depending on their residual from the hypocenter location where green is low residual, yellow is moderate residual and red is high residual (the color levels are configurable). In this example, there are no stations towards the west and southwest of the epicenter and stations are sparse toward the southeast, resulting in an azimuthal gap of 216 degree. This, together with the dominating N-S extension of the Norwegian network causes the much larger location uncertainty in E-W direction (see error ellipse in the left-hand panel). We can also plot the epicenter together with previously located earthquakes.

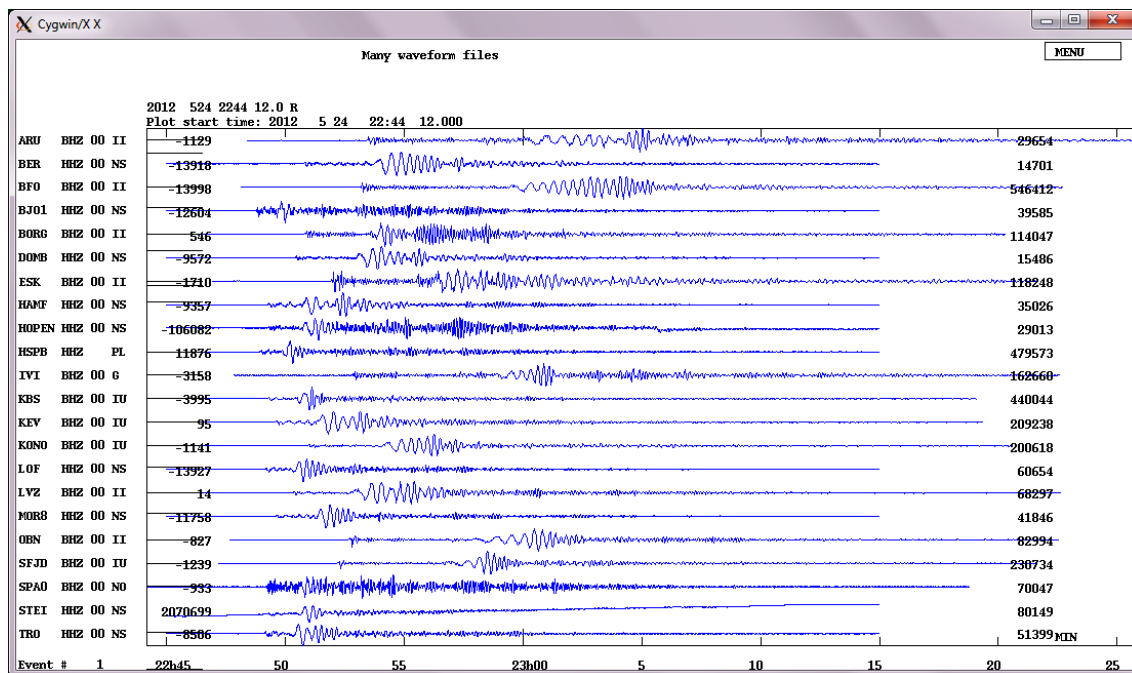
### 3 Analysis of a regional earthquake recorded at distance between 440 and 2950 km ( $\Delta < 30^\circ$ )

As an example we have chosen an earthquake that occurred on the Mohns Ridge, offshore northern Norway. The earthquake with a magnitude greater than 6 was large enough to be recorded globally. Here, we process data that were recorded on the Norwegian National Seismic Network operated by the University of Bergen and data provided by IRIS from the Global Seismograph Network (GSN). All data used here was recorded on broadband stations.

The earthquake parameters provided by the USGS are:

|              |                     |
|--------------|---------------------|
| Date:        | 24 May 2012         |
| Origin time: | 22:47:46.6<br>(UTC) |
| Latitude:    | 72.99°N             |
| Longitude:   | 5.65°E              |
| Depth:       | 8.8 km              |
| Mw:          | 6.2                 |

We now show the step-by-step procedure to analyze the earthquake.

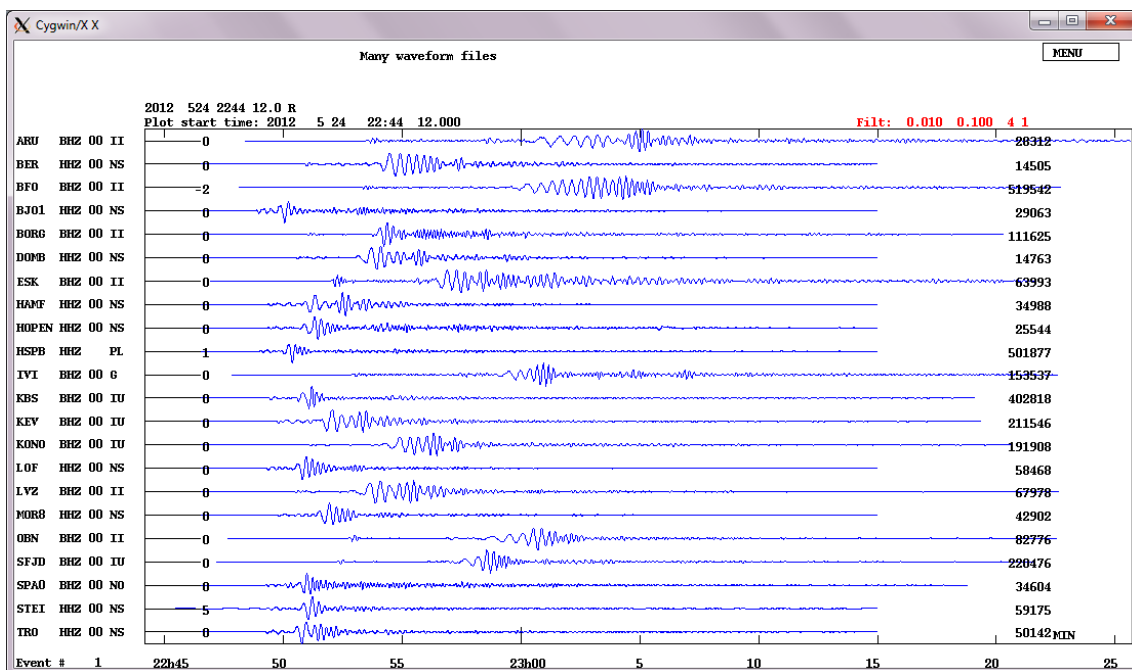


*Vertical component multi-trace plot, unfiltered:* On the seismograms we clearly see long-period surface waves, with large amplitudes and pronounced dispersion. This indicates that the earthquake is relatively large and shallow. For the stations, where we see the first arriving P as well we can estimate the approximate distance from the time difference between the P and the Rayleigh surface wave maximum (which is most pronounced in vertical component records). For station BFO this time difference is about 10 min. According to Table 1 below this means that station BFO is about 20 to 25° degrees away ( $1^\circ = 111.2$  km) from the epicenter. Being aware of the station locations, considering the pattern of arrival times, and,

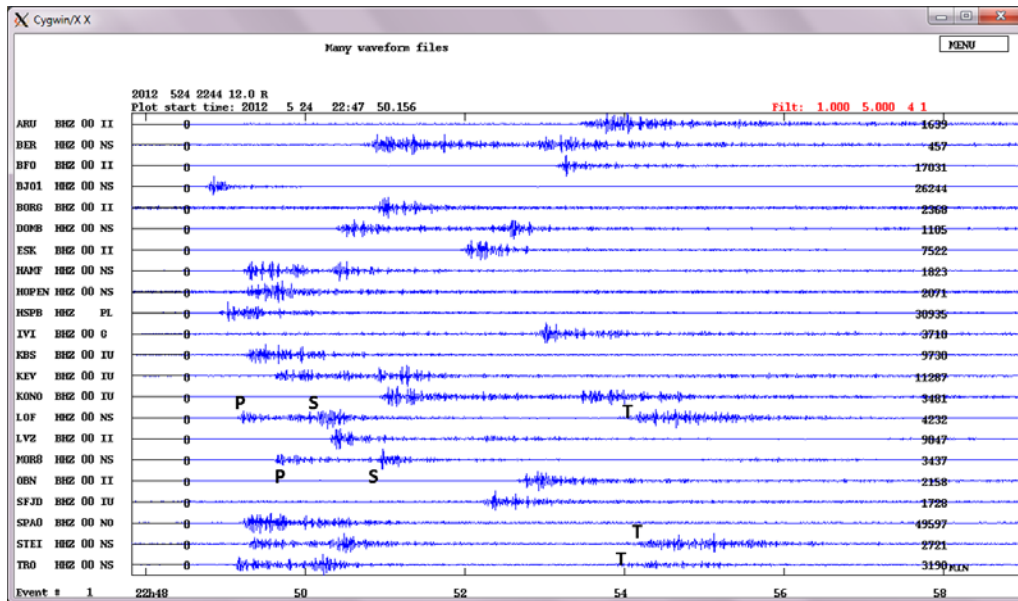
if three-component records are available, also of the first-motion polarity patterns and the better constraining S-P travel-time differences, we can also get a first guess of the epicenter location (see EX 11.2).

**Table** Approximate time interval ( $t_{Rmax} - t_p$ ) between the arrival of the maximum Rayleigh wave amplitude and the first onset of P waves as a function of epicentral distance  $\Delta$  according to Archangelskaya (1959) and Gorbunova and Kondorskaya (1977) for dominantly continental travel paths.

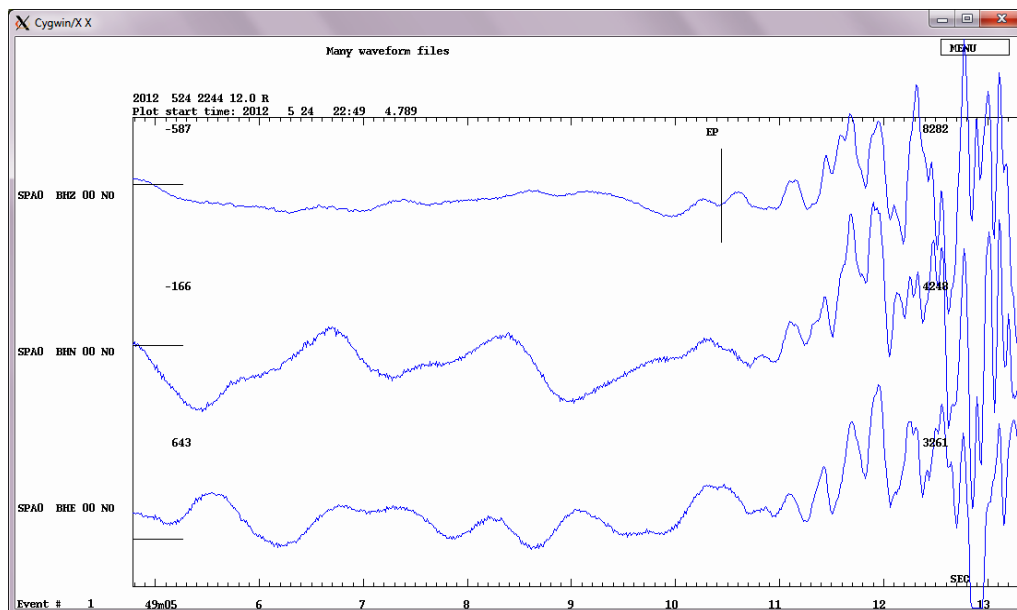
| $\Delta^\circ$ | $t_{Rmax} - t_p$<br>(min) | $\Delta^\circ$ | $t_{Rmax} - t_p$<br>(min) | $\Delta^\circ$ | $t_{Rmax} - t_p$<br>(min) |
|----------------|---------------------------|----------------|---------------------------|----------------|---------------------------|
| 10             | 4-5                       | 55             | 26                        | 100            | 45-46                     |
| 15             | 6-8                       | 60             | 28-29                     | 105            | 47-48                     |
| 20             | 9-10                      | 65             | 31                        | 110            | 48-50                     |
| 25             | 10-12                     | 70             | 33                        | 115            | 53                        |
| 30             | 13-14                     | 75             | 35                        | 120            | 55                        |
| 35             | 15-16                     | 80             | 37                        | 125            | 57                        |
| 40             | 18-19                     | 85             | 39-40                     | 130            | 60                        |
| 45             | 21                        | 90             | 42                        | 140            | 64                        |
| 50             | 24                        | 95             | 43                        | 150            | 70                        |



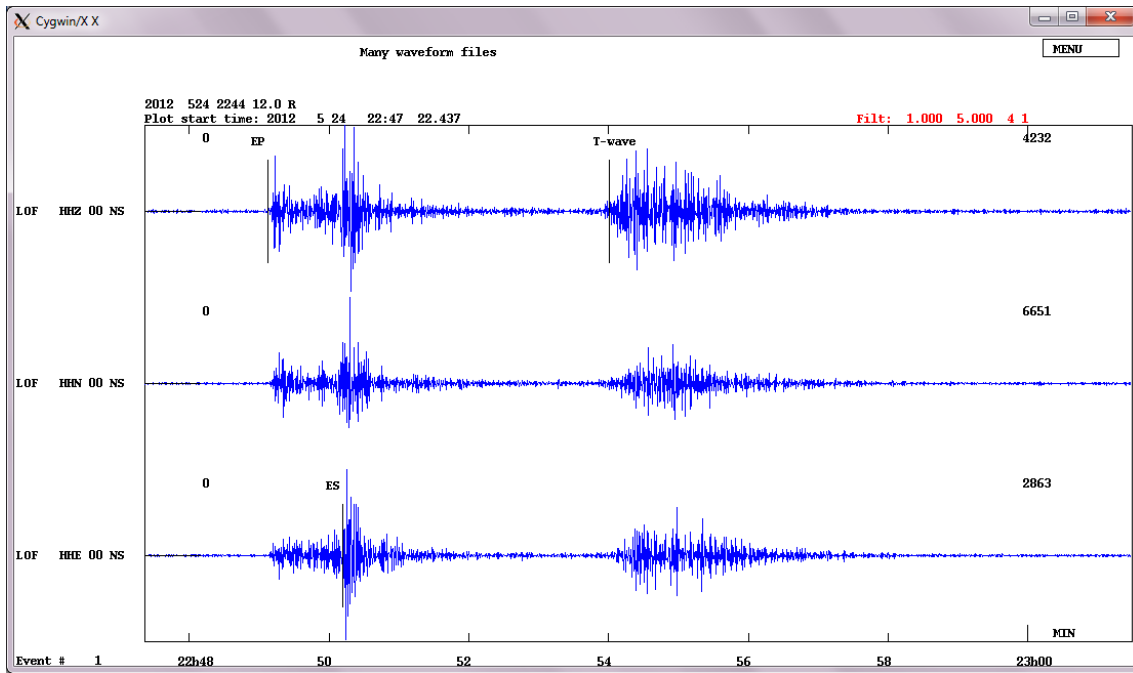
*Vertical component multi-trace plot, filtered 0.01-0.1 Hz:* Filtering the data at low frequencies, we see both body and surface waves. However, the seismograms are dominated by the surface waves as they have larger amplitudes in this frequency range. The shape of the surface wave trains depends on the Earth structure along the paths (see, e.g., Chapter 2, Figs. 2.10, 2.12 and 2.13) and, because of the dispersion, i.e., the dependence of the surface-wave velocity on the period, also on the epicentral distance (see Figure above). The seismograms also depend on the mechanism of faulting.



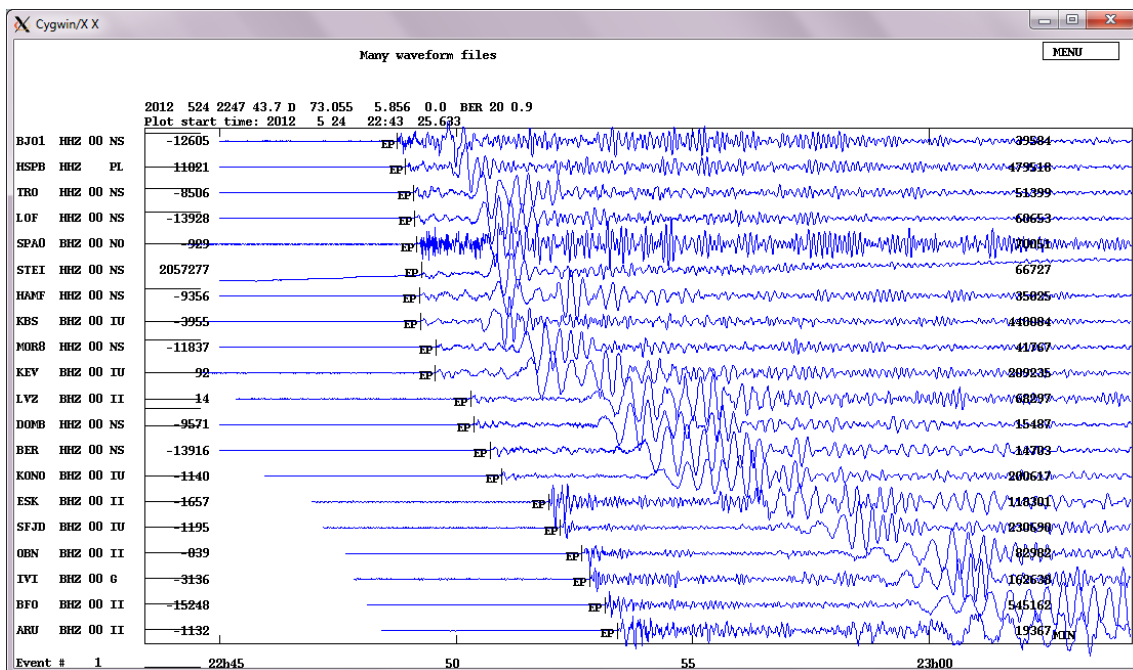
*Multi-trace plot, filtered 1-5 Hz:* Looking at higher frequencies, the surface waves disappear from the seismogram and we look at P and (at these high frequencies and in vertical component records usually less clear) S body waves. If the first arriving P and S phases are  $P_n$  and  $S_n$ , respectively, then there is a simple “rule of thumb” to estimate the hypocenter distance, namely time difference  $S_n - P_n$  [in s]  $\times 10$ . For station LOF with a time difference of about 1 min this means a distance of  $\approx 600$  km. This agrees with the rough estimate from the time difference between the Rayleigh-wave maximum and the P first arrival. On some stations (STEI, TRO and LOF) we can also see a signal in this frequency range arriving much later (about 5 minutes after P for station TRO). This is the T(tertiary)-wave, generated by an earthquake in or near the oceans, propagating in the oceans as an acoustic wave and converted back to seismic wave at the ocean-land boundary near the recording site. (for more explanations see Chapter 2, section 2.6.5). It should not be mistaken for another event.



*Three-component traces of station SPA0 at epicentral distance of 648 km, unfiltered:* Zooming around the P arrival the phase can be picked with or without filter. The error on the phase reading here may be as large as 0.5 sec.

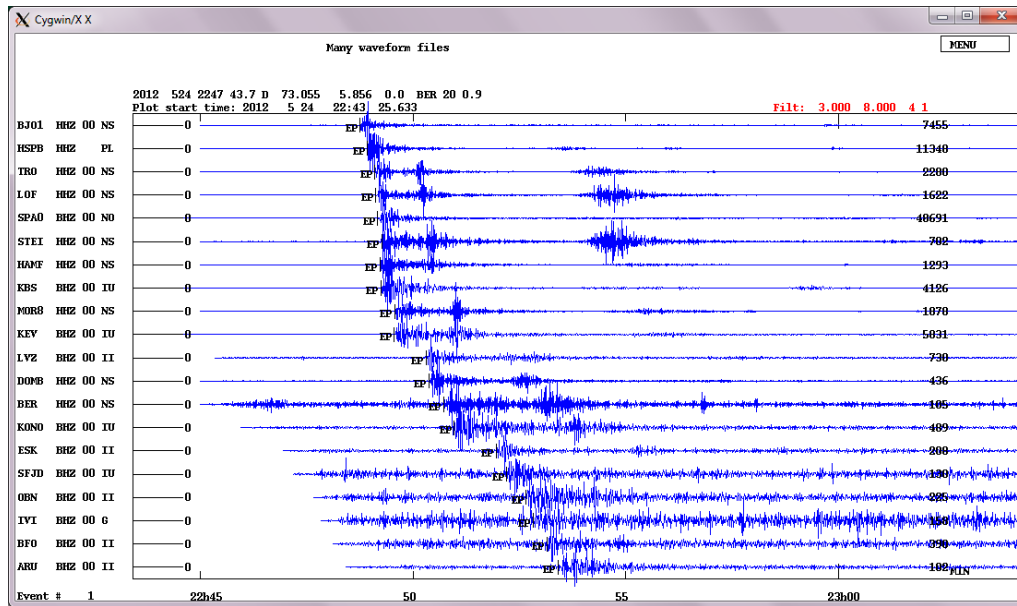


*Three-component traces of station LOF at epicentral distance of 603 km, filtered 1-5 Hz:* Looking at this coastal station, we clearly see the arrival of the T wave. The amplitude of the T wave is in this case comparable to the S-wave amplitude. According to the time difference (S-P)[in s]×10 the station is about 600 km away from the source.

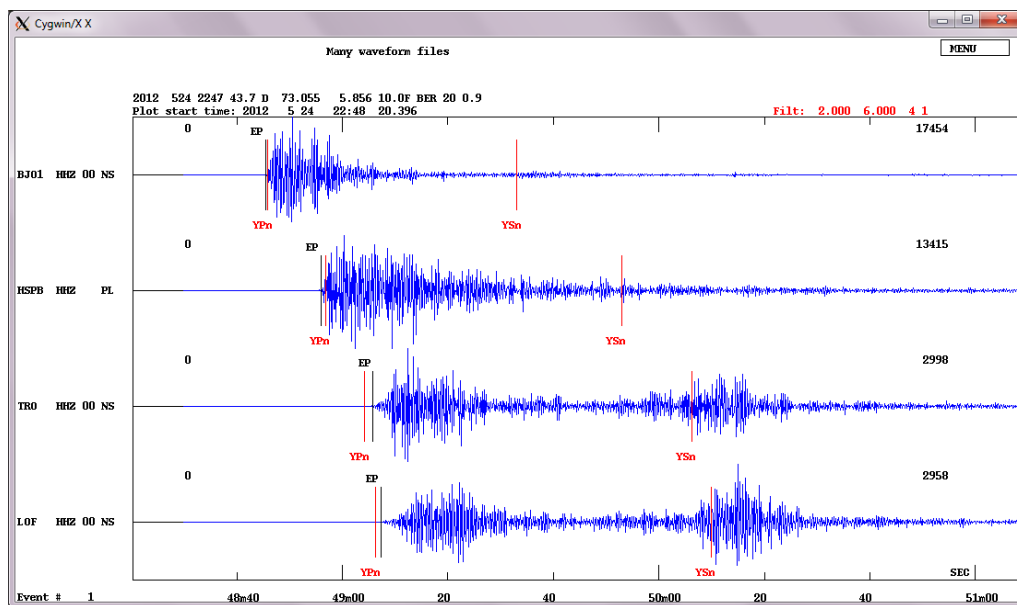


*All vertical traces, unfiltered:* The traces are sorted by distances and the P arrival picks are shown. This type of plot is useful to check the consistency of the picks. Based on the P arrival times, the epicenter is estimated at 73.055°N and 5.856°E. The hypocenter depth is not constrained by the data and given as 0 km by the location program. On stations like SPA0 we can see that we have both high and low frequency. Towards larger distances the high frequencies get attenuated more strongly than lower frequency signals.

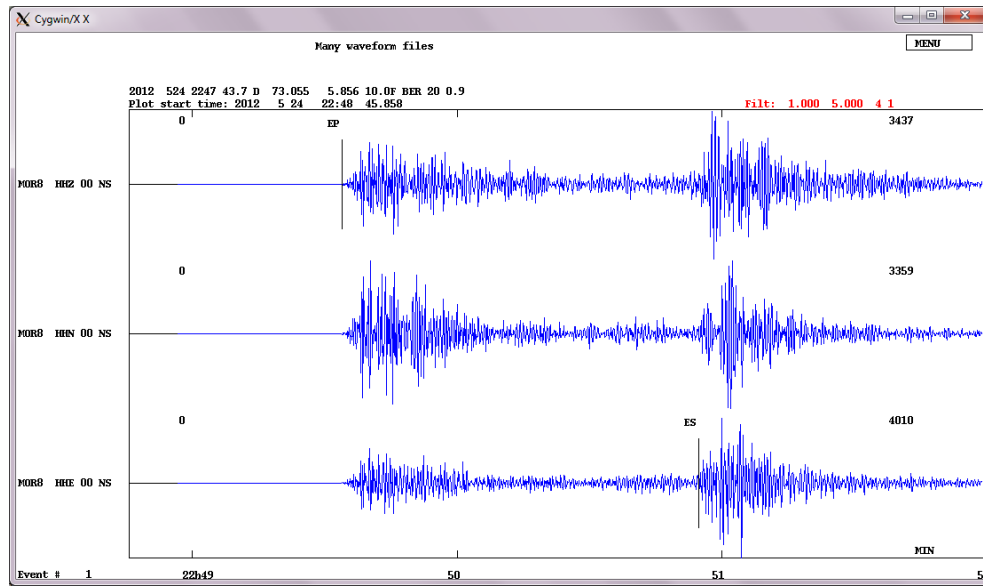




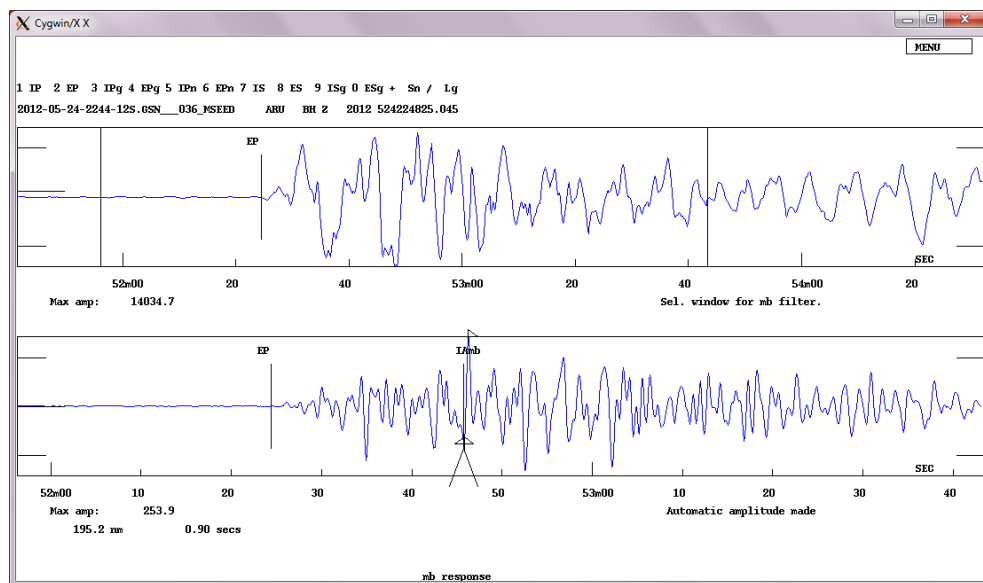
*All vertical traces, filtered 3-8 Hz:* We can see that the signal amplitude of the higher frequencies decreases as we go to larger distances due to attenuation. The distance range here is approximately 440 km (station BJO1) to 2960 km (station ARU).



*Selected vertical traces, filtered 2-6 Hz:* Computed Pn and Sn phase arrival times have been marked by red vertical lines and the text 'yPn' and 'ySn'. No S-waves are recognizable in the records of the two stations BJO1 and HSPB closer to the epicenter. They are likely situated in an azimuth range of minimum S-wave and strong P-wave radiation (see EX 3.2). While S appears at larger distances, the phase is emergent and from this plot its travel-time does not match well with the global velocity model used (IASP91). The earthquake is located on the Mid-Atlantic Ridge. The seismograms are complicated both due to source and path effects, including the transition from oceanic to continental crust. The display of theoretical arrivals should only serve as a guide on whether the considered phase is plausible to arrive in this time window. However, it would be **wrong** to pick a phase close to the predicted time if there is no visible change in the signal.

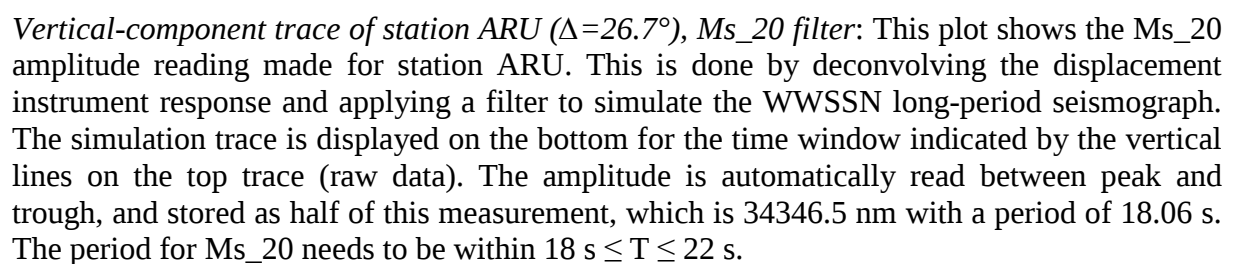
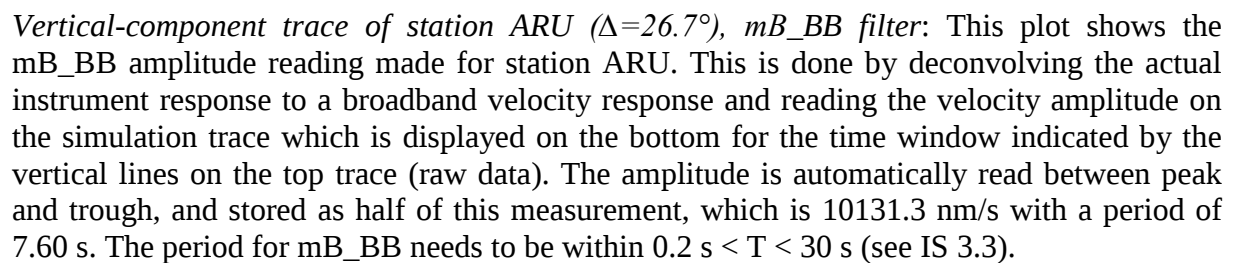


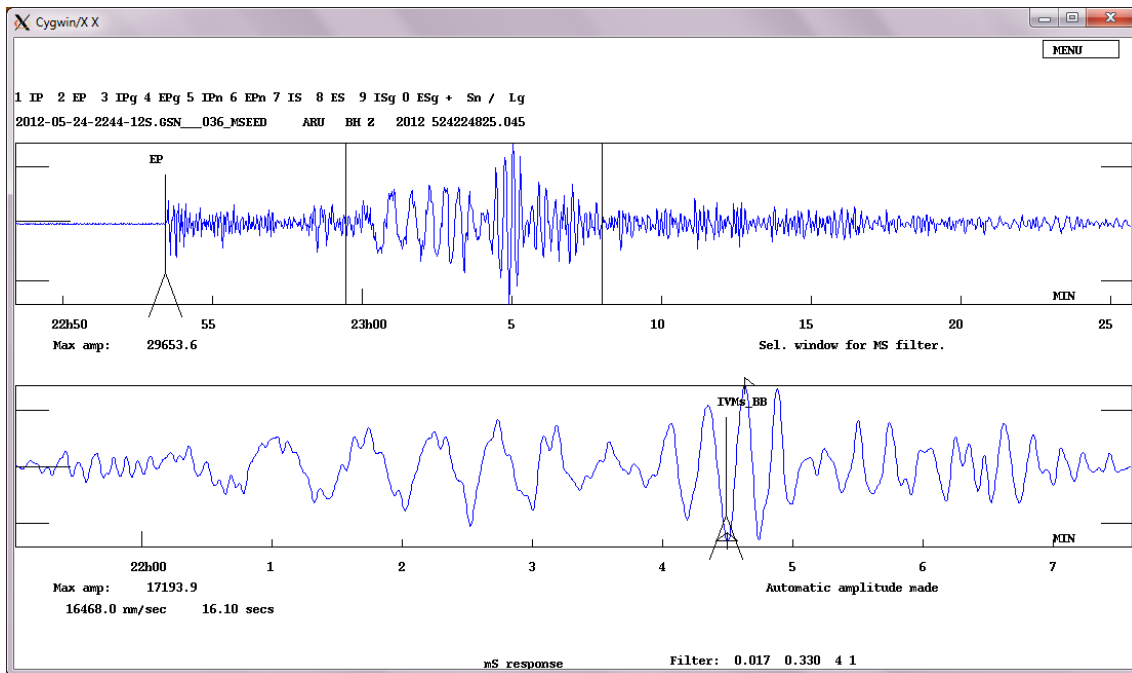
*Three-component trace plot of station MOR8 at epicentral distance of 813 km, filtered 1-5 Hz: This is an example where the S arrival can be read quite well on one of the horizontal channels.*



*Vertical-component trace of station ARU ( $\Delta=26.7^\circ$ ), mb filter: While it is possible to read ML amplitudes from this earthquake for stations within about 1500 km, this is not done because the ML scale for Norway is not appropriate for sources on the Mid-Atlantic Ridge. We, therefore, read amplitudes for the teleseismic magnitudes which can be used down to distances of  $20^\circ$  and even down to  $2^\circ$  for standard  $M_s$ \_BB. Moment magnitude for this event can be measured using either regional or teleseismic moment tensor inversion (USGS  $M_w = 6.2$ ). The plot above shows the mb amplitude reading for station ARU. This is done by deconvolving the broadband instrument response and applying a filter to simulate the WWSSN short-period seismograph (see IS 3.3). The simulation trace is displayed on the bottom trace for the time window indicated by the vertical lines on the top trace (raw data). The amplitude  $I_{amb}$  is automatically read between the largest peak and adjacent trough. It is stored as half of this measurement, which is 195.2 nm, together with the related period  $T$  of 0.90 s which is twice the time difference between peak and trough.*





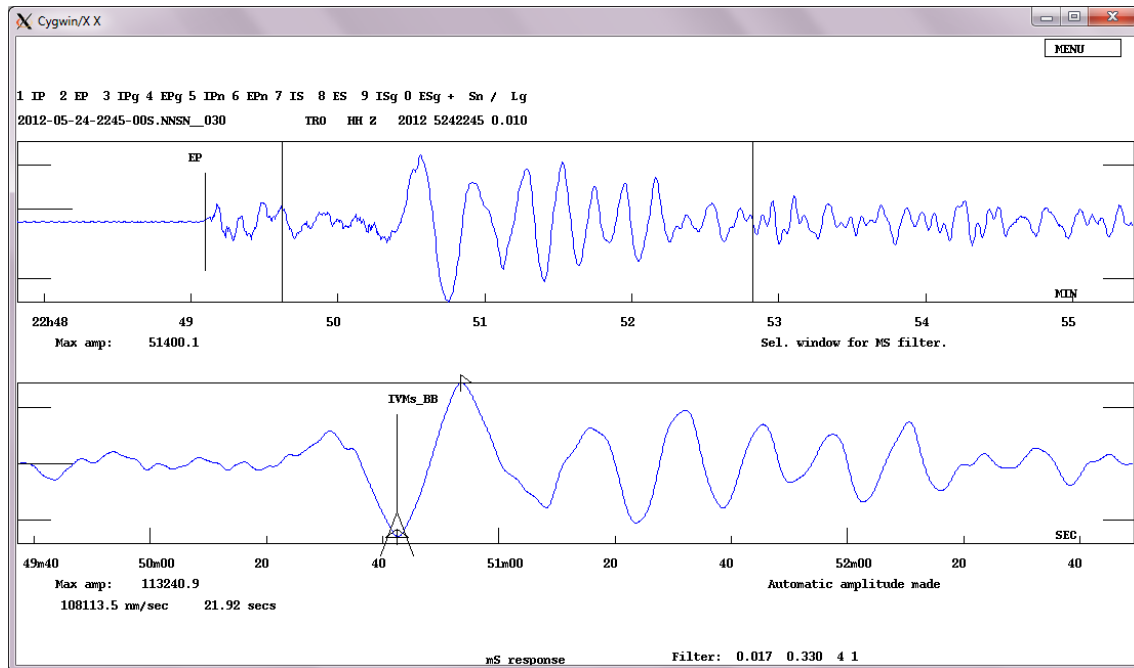


*Vertical-component trace of station ARU ( $\Delta = 26.7^\circ$ ),  $Ms\_BB$  filter:* This plot shows the  $Ms\_BB$  amplitude reading made for station ARU. If the primary record is not yet a velocity broadband trace then the latter has to be synthesized by deconvolving the actual instrument response to a broadband velocity response and reading the velocity amplitude on the simulation trace. The latter is displayed on the bottom trace for the time window indicated by the vertical lines on the top trace (raw data). The amplitude is automatically read between the largest peak and adjacent trough, and stored as half of this measurement, which is 16468.0 nm/s, together with a period of 16.10 s, i.e., twice the time difference between the peak and adjacent trough. The period needs to be within  $3 \text{ s} \leq T \leq 60 \text{ s}$ , and thus maybe significantly shorter than for  $Ms\_20$  measurement, allowing  $Ms$  measurements at distances well below  $20^\circ$  (down to  $2^\circ$ ).

Based on these four amplitude and period measurements from station ARU, we can compare the magnitudes to  $M_w$ :

| Magnitude scale | Value |
|-----------------|-------|
| $M_w$ (USGS)    | 6.2   |
| mb              | 5.9   |
| $mB\_BB$        | 6.7   |
| $Ms\_20$        | 6.0   |
| $Ms\_BB$        | 6.1   |

We see that the values for this size of earthquake around magnitude 6 compare well, with the exception of  $mB\_BB$  which is on global average about 0.3 -0.4 m.u. greater than  $M_w$  at  $M_w = 6.2$  and about 0.6 m.u. larger than mb.



Vertical-component trace of station TRO at  $\Delta = 5.3^\circ$ ,  $M_s_{BB}$  filter:  $M_s_{BB}$  can be used down to distances of  $2^\circ$ . This plot shows the  $M_s_{BB}$  amplitude reading made for station TRO. The amplitude is automatically read between peak and trough, and stored as half of this measurement, which is 108113.5 nm/s with a period of 21.92 s. The computed magnitude is  $M_s_{BB} = 5.7$ , compared to  $M_s_{BB} = 6.1$  for station ARU.

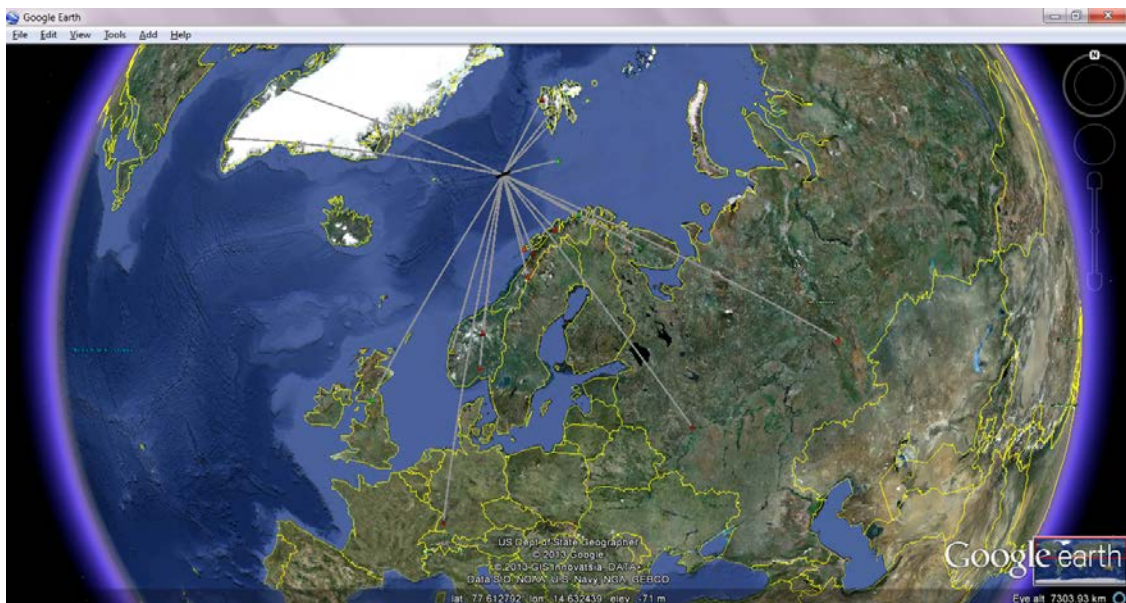
|        |      |                      |          |         |        |       |          |         |        |        |         |         |         |
|--------|------|----------------------|----------|---------|--------|-------|----------|---------|--------|--------|---------|---------|---------|
| #      | 1    | 24 May 2012 22:47 45 | D        | 72.943  | 6.096  | 8.8F  | 1.8      | 6.3sBER | 20     | ? 1    |         |         |         |
| date   | hrmn | sec                  | lat      | long    | depth  | no    | m        | rms     | damp   | erln   | erlt    | erdp    |         |
| 12 524 | 2247 | 45.09                | 7256.61N | 6       | 5.7E   | 8.8*  | 24 2     | 1.77    | 0.000  | 47.0   | 9.2     | 0.0     |         |
| stn    | dist | azm                  | ain      | w       | phas   | calcp | h        | t       | obs    | t-cal  | res     | wt      | di      |
| BJO1   | 439  | 60.6                 | 45.8     | 0       | P      | Pn    | 2248     | 45.4    | 60.36  | 60.82  | -0.46   | 1.00    | 6       |
| HSPB   | 528  | 26.6                 | 45.8     | 0       | P      | Pn    | 2248     | 56.0    | 70.88  | 71.85  | -0.97   | 1.00    | 7       |
| TRO    | 587  | 122.6                | 45.8     | 0       | P      | Pn    | 2249     | 5.8     | 80.70  | 79.19  | 1.51    | 1.00    | 1       |
| TRO    | 587  | 122.6                | 0        | IVMs_BB |        |       | 2250     | 42.4    | 177.3  |        |         |         |         |
| LOF    | 603  | 149.2                | 45.8     | 0       | P      | Pn    | 2249     | 7.4     | 82.27  | 81.18  | 1.09    | 1.00    | 1       |
| LOF    | 603  | 149.2                | 48.3     | 0       | S      | Sn    | 2250     | 6.4     | 141.35 | 144.88 | -3.53   | 1.00    | 5       |
| SPA0   | 648  | 21.2                 | 45.8     | 0       | P      | Pn    | 2249     | 10.4    | 85.35  | 86.84  | -1.49   | 1.00    | 7       |
| STEI   | 654  | 144.2                | 45.8     | 0       | P      | Pn    | 2249     | 15.7    | 90.63  | 87.49  | 3.14    | 1.00    | 1       |
| STEI   | 654  | 144.2                | 48.2     | 0       | S      | Sn    | 2250     | 18.9    | 153.78 | 156.21 | -2.43   | 1.00    | 6       |
| HAMF   | 662  | 104.2                | 45.8     | 0       | P      | Pn    | 2249     | 13.8    | 88.70  | 88.50  | 0.20    | 1.00    | 3       |
| KBS    | 685  | 10.5                 | 45.8     | 0       | P      | Pn    | 2249     | 14.6    | 89.50  | 91.27  | -1.77   | 1.00    | 8       |
| MOR8   | 813  | 151.6                | 45.7     | 0       | P      | Pn    | 2249     | 34.0    | 108.87 | 107.24 | 1.63    | 1.00    | 1       |
| MOR8   | 813  | 151.6                | 48.1     | 0       | S      | Sn    | 2250     | 54.7    | 189.65 | 191.69 | -2.04   | 1.00    | 5       |
| KEV    | 821  | 105.3                | 45.7     | 0       | P      | Pn    | 2249     | 33.8    | 108.76 | 108.20 | 0.56    | 1.00    | 3       |
| LVZ    | 1191 | 103.8                | 45.6     | 0       | P      | Pn    | 2250     | 18.8    | 153.67 | 153.99 | -0.32   | 1.00    | 3       |
| DOMB   | 1219 | 172.5                | 45.6     | 0       | P      | Pn    | 2250     | 22.6    | 157.46 | 157.34 | 0.12    | 1.00    | 2       |
| DOMB   | 1219 | 172.5                | 47.8     | 0       | S      | Sn    | 2252     | 25.0    | 279.88 | 281.53 | -1.65   | 1.00    | 7       |
| BER    | 1401 | 181.7                | 45.5     | 0       | P      | Pn    | 2250     | 43.7    | 178.61 | 179.70 | -1.09   | 1.00    | 3       |
| KONO   | 1490 | 172.3                | 45.4     | 0       | P      | Pn    | 2250     | 58.2    | 193.13 | 190.69 | 2.44    | 1.00    | 2       |
| ESK    | 2010 | 197.3                | 39.8     | 0       | P      | Pn    | 2251     | 57.3    | 252.18 | 251.84 | 0.34    | 1.00    | 5       |
| ESK    | 2010 | 197.3                | 0        | IVmB_BB |        |       | 2252     | 6.0     | 260.9  |        |         |         |         |
| ESK    | 2010 | 197.3                | 0        | IAMB    |        |       | 2252     | 12.8    | 267.7  |        |         |         |         |
| ESK    | 2010 | 197.3                | 0        | IAMS_20 |        |       | 2257     | 17.9    | 572.8  |        |         |         |         |
| SFJD   | 2172 | 281.2                | 34.8     | 0       | P      | P     | 2252     | 11.8    | 266.68 | 267.71 | -1.03   | 1.00    | 12      |
| SFJD   | 2172 | 281.2                | 0        | IAMB    |        |       | 2252     | 23.3    | 278.2  |        |         |         |         |
| SFJD   | 2172 | 281.2                | 0        | IVMs_BB |        |       | 2258     | 30.4    | 645.3  |        |         |         |         |
| SFJD   | 2172 | 281.2                | 0        | IAMS_20 |        |       | 2258     | 34.0    | 648.9  |        |         |         |         |
| OBN    | 2420 | 128.4                | 34.0     | 0       | P      | P     | 2252     | 39.3    | 294.24 | 291.96 | 2.28    | 1.00    | 0       |
| OBN    | 2420 | 128.4                | 0        | IVmB_BB |        |       | 2252     | 50.0    | 304.9  |        |         |         |         |
| OBN    | 2420 | 128.4                | 0        | IAMB    |        |       | 2252     | 55.4    | 310.3  |        |         |         |         |
| OBN    | 2420 | 128.4                | 0        | IAMS_20 |        |       | 23 0     | 30.7    | 765.6  |        |         |         |         |
| OBN    | 2420 | 128.4                | 0        | IVMs_BB |        |       | 23 0     | 39.0    | 773.9  |        |         |         |         |
| IVI    | 2570 | 268.3                | 33.4     | 0       | P      | P     | 2252     | 50.1    | 304.96 | 306.21 | -1.25   | 1.00    | 11      |
| BFO    | 2743 | 176.4                | 28.4     | 0       | P      | P     | 2253     | 9.1     | 324.01 | 321.15 | 2.86    | 1.00    | 1       |
| ARU    | 2954 | 100.0                | 28.1     | 0       | P      | P     | 2253     | 25.2    | 340.13 | 338.25 | 1.87    | 1.00    | 1       |
| ARU    | 2954 | 100.0                | 0        | IAMB    |        |       | 2253     | 45.7    | 360.6  |        |         |         |         |
| ARU    | 2954 | 100.0                | 0        | IVmB_BB |        |       | 2253     | 45.8    | 360.7  |        |         |         |         |
| ARU    | 2954 | 100.0                | 0        | IAMS_20 |        |       | 23 4     | 49.9    | 1024.8 |        |         |         |         |
| ARU    | 2954 | 100.0                | 0        | IVMs_BB |        |       | 23 4     | 54.1    | 1029.0 |        |         |         |         |
| TRO    | HZ   | dist:                | 587.0    | amp:    |        |       | 108000.0 | T:      | 21.9   | MS =   | 5.7     |         |         |
| ESK    | BZ   | dist:                | 2010.0   | amp:    |        |       | 154000.0 | T:      | 7.1    | mB =   | 7.3     |         |         |
| ESK    | BZ   | dist:                | 2010.0   | amp:    |        |       | 11890.7  | T:      | 1.8    | mb =   | 6.7     |         |         |
| ESK    | BZ   | dist:                | 2010.0   | amp:    |        |       | 363000.0 | T:      | 21.6   | Ms =   | 6.6     |         |         |
| SFJD   | BZ   | dist:                | 2172.0   | amp:    |        |       | 4562.0   | T:      | 2.9    | mb =   | 6.2     |         |         |
| SFJD   | BZ   | dist:                | 2172.0   | amp:    |        |       | 158000.0 | T:      | 18.0   | Ms =   | 6.4     |         |         |
| OBN    | BZ   | dist:                | 2420.0   | amp:    |        |       | 31652.9  | T:      | 5.2    | mB =   | 6.9     |         |         |
| OBN    | BZ   | dist:                | 2420.0   | amp:    |        |       | 6234.6   | T:      | 2.4    | mb =   | 6.6     |         |         |
| OBN    | BZ   | dist:                | 2420.0   | amp:    |        |       | 168000.0 | T:      | 21.8   | Ms =   | 6.4     |         |         |
| OBN    | BZ   | dist:                | 2420.0   | amp:    |        |       | 49453.8  | T:      | 19.6   | MS =   | 6.4     |         |         |
| ARU    | BZ   | dist:                | 2954.0   | amp:    |        |       | 269.7    | T:      | 1.0    | mb =   | 5.9     |         |         |
| ARU    | BZ   | dist:                | 2954.0   | amp:    |        |       | 10130.1  | T:      | 7.6    | mB =   | 6.7     |         |         |
| ARU    | BZ   | dist:                | 2954.0   | amp:    |        |       | 36322.6  | T:      | 18.1   | Ms =   | 6.0     |         |         |
| ARU    | BZ   | dist:                | 2954.0   | amp:    |        |       | 16468.0  | T:      | 16.1   | MS =   | 6.1     |         |         |
| 2012   | 524  | 2247                 | 45.1     | D       | 72.943 | 6.096 | 8.8F     | BER     | 20     | 1.8    | 6.3sBER | 6.1sBER | 6.4bBER |

*Output from location program:* We use the same standard linearized inversion program (HYPOCENTER) to locate this event as used for the local event. The location is based on the IASP91 global travel-time model. The residuals are much larger than for the local event as the difference between the true regional model and IASP91 is significant. The largest residual here is more than 3 seconds. In the hypocenter determination we fixed the depth to 8.8 km and, using both P and S readings, we obtain a latitude of 72.943°N and longitude of 6.096°E. This compares to 73.055°N and 5.856°E for the preliminary location based on P arrivals only. The solution based on P and S is 15 km east of the USGS solution.

Based on all our amplitude and period measurements the following network average magnitude values were calculated:

| Magnitude scale | Value |
|-----------------|-------|
| Mw (USGS)       | 6.2   |
| mb (USGS)       | 5.7   |
| mb              | 6.4   |
| mB_BB           | 7.0   |
| Ms_20           | 6.3   |
| Ms_BB           | 6.1   |

Note that Ms\_20 and Ms\_BB are within 0.1 m.u. of the USGS Mw. In contrast, both our mb and mB\_BB are significantly larger, with mb, based on only 4 measurements in the near teleseismic distance range between 2000 and 3000 km, even by 0.7 m.u. larger than USGS mb, based on 299 measurements at a much larger range of distances up to about 11,000 km. This may hint to larger calibration uncertainties of the Gutenberg-Richter (1956) Q function in the near teleseismic distance range, where wave propagation is largely influenced by the greater inhomogeneity of upper mantle wave propagation and attenuation. This may result in larger uncertainty of body-wave magnitude estimates, if they are available in only very limited number in the near teleseismic range. On the other hand, this earthquake has taken place on the North-Atlantic Ridge (see Google map below). Intra-oceanic ridge strike-slip earthquakes are often high stress drop events characterized by a much larger than average ratio of  $E_s/M_0$  and thus much larger  $M_e$  than Mw (e.g. Bormann and Di Giacomo, 2011; Choy, 2012, in IS 3.5). Then, mb and mB\_BB may be much larger than usual too in their relation to long-period Mw and Ms. Therefore one might consider this as an alternative explanation, which is, however, neither supported by the much more representative USGS mb value (being rather normal for Mw = 6.2) nor by the relatively long and not very impulsive P waveforms. The latter are, in the case of high stress drop events, expected to be relatively short, impulse-like and of more high-frequency content (e.g., Fig. 3.6 in Chapter 3). Instead, the P-wave group is rather long for an Mw6.2 event, almost one minute, and relatively long-period.



*Location map in GoogleEarth:* This map shows the location with its error ellipse on the Mid-Atlantic ridge and the stations that were used for the analysis and location.

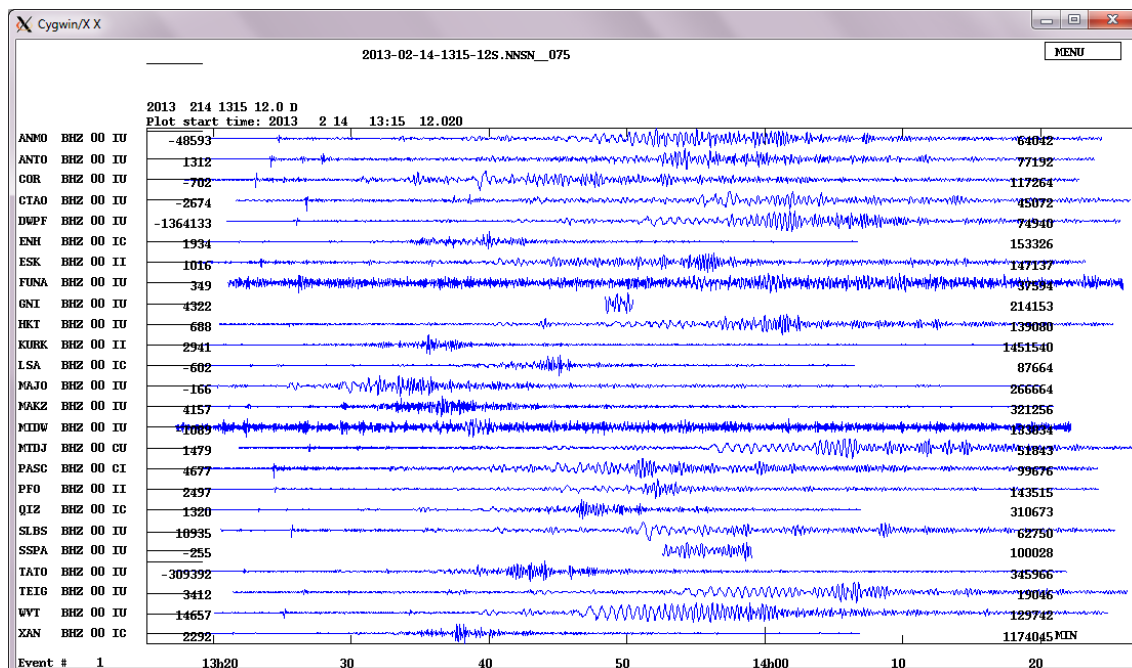
## 4 Analysis of a shallow earthquake recorded at teleseismic distances ( $30^\circ < \Delta < 90^\circ$ )

The earthquake chosen here as an example occurred in Russia at a shallow depth of 10 km (according to USGS). We extracted data from IRIS that was recorded on the Global Seismograph Network (GSN). All data used here were recorded on broadband stations.

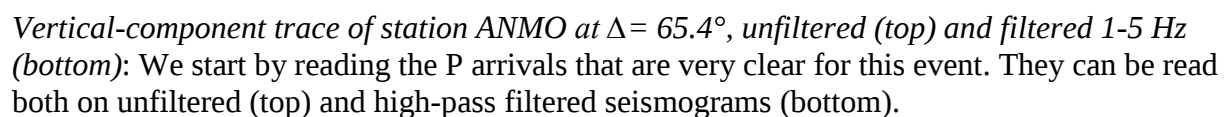
The earthquake parameters provided by the USGS are:

|              |                  |
|--------------|------------------|
| Date:        | 14/02/2013       |
| Origin time: | 13:13:53.1 (UTC) |
| Latitude:    | 67.61°N          |
| Longitude:   | 142.60°W         |
| Depth:       | 10 km            |
| Mw:          | 6.9              |

We now show the step-by-step procedure to analyze the earthquake.

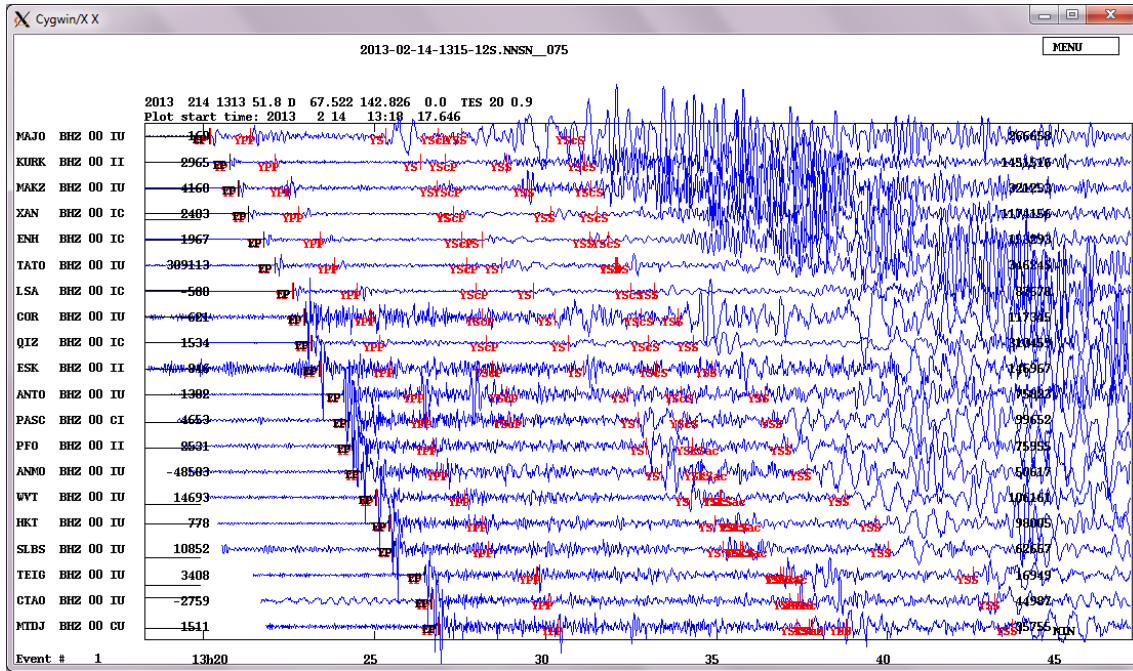


*Multi-trace plot of vertical components, unfiltered:* Record durations up to about 60 minutes at some stations and the observation of much larger surface-wave than body wave amplitudes hints to a large shallow earthquake at teleseismic distances.

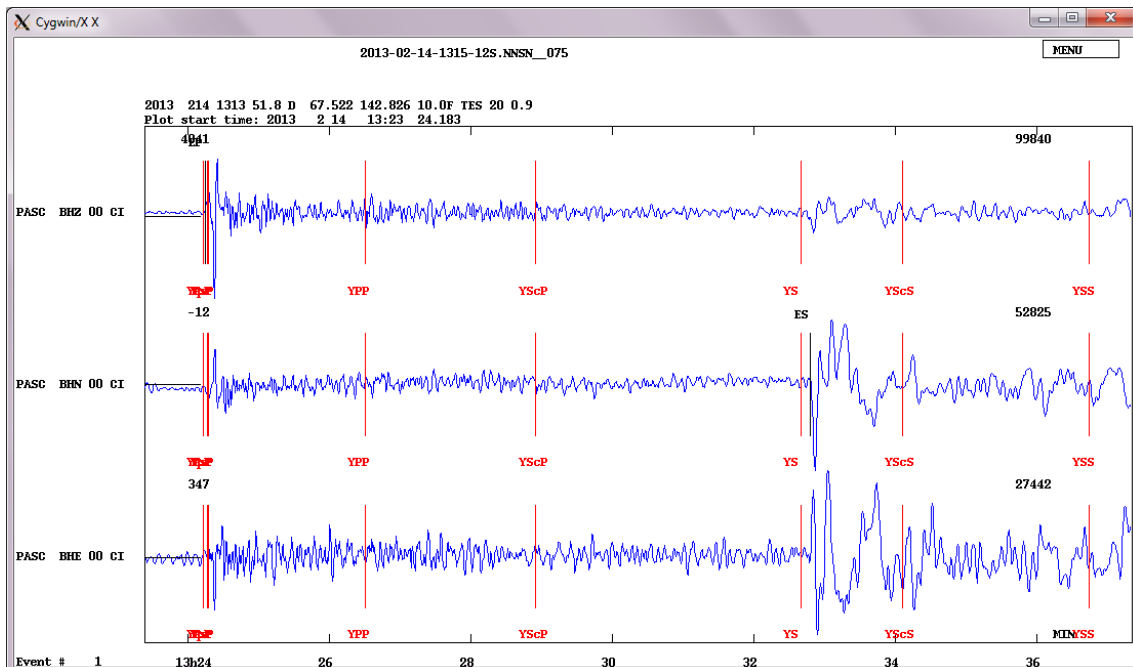


*Output from the location program:* Locating with the first P arrivals only, we obtain an epicenter location (67.522°N, 142.826°W) and depth = 0 km that is very close to the one given by the USGS. The depth is not constrained with the teleseismic P arrivals only.



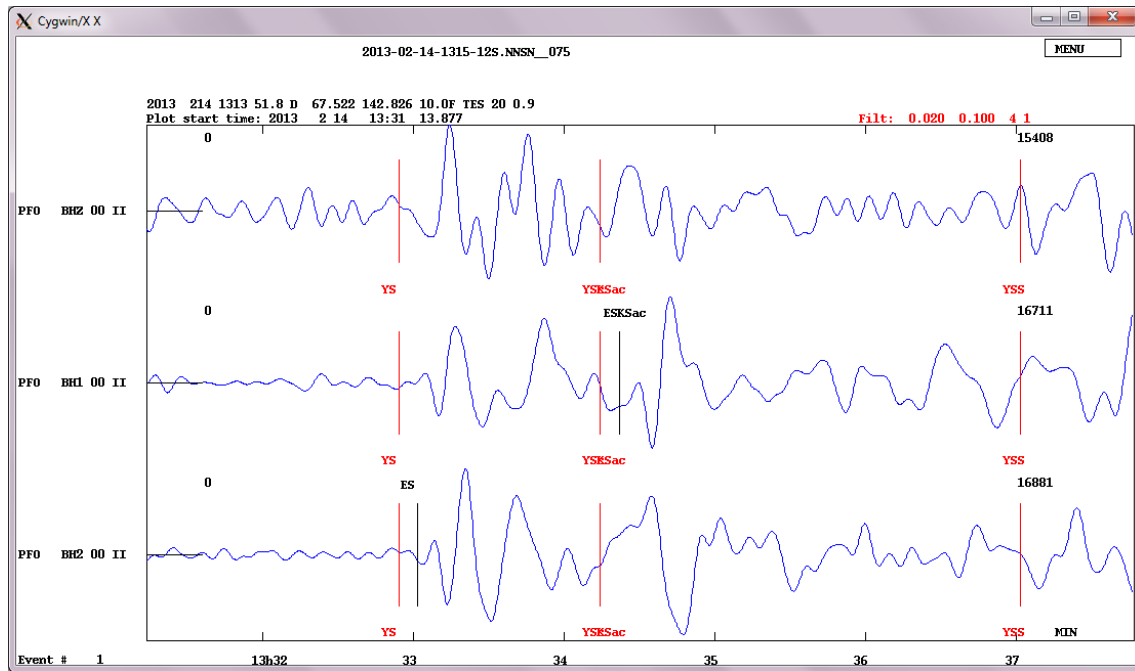


*Multi-trace plot of vertical components, unfiltered:* Plotted are the traces on which we have read the initial P. Based on the hypocenter obtained, we compute the synthetic arrival times of all phases. The traces are sorted by distance. The phases PP, S and SS are clear on all traces. We will now read the onset times of all these phases (plus P-wave first motion polarity), run the location program again with this additional information and see how the location changes.



*Three-component trace plot of station PASC at  $\Delta=62.2^\circ$ , unfiltered:* For this station P on the vertical component and S on the N component are very clear. Other phases are not very clear. Note that we do not pick the onsets at the calculated arrival times (red bars) but only at the distinct onset of the real phases with large SNR (black bars).

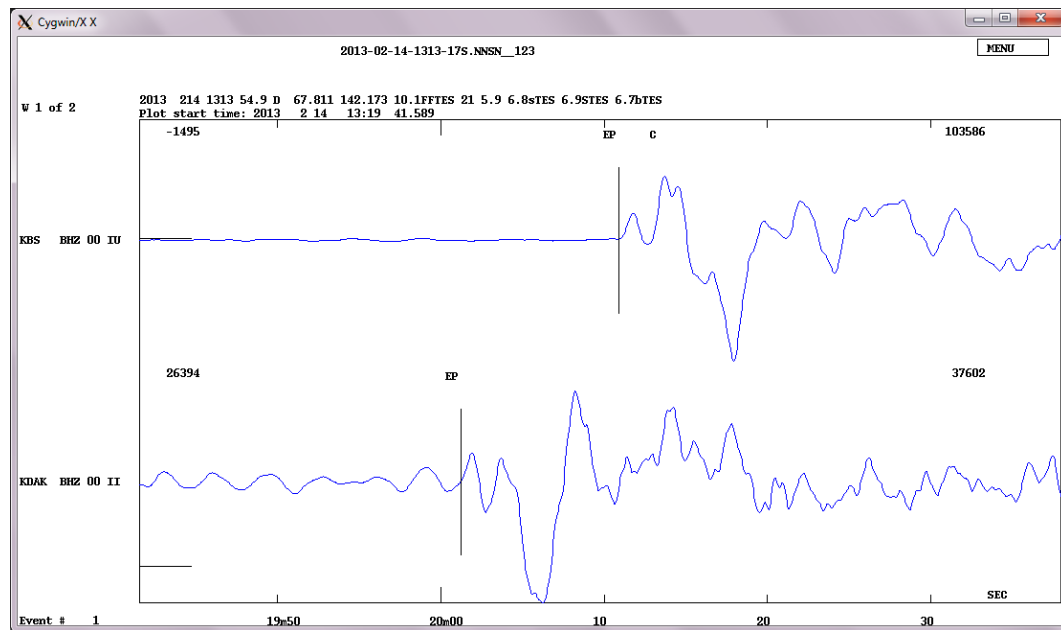




*Three-component trace plot of station PFO at  $\Delta = 63.3^\circ$ , filtered 0.02-0.1 Hz:* At this distance there is sufficient separation between S and SKS<sub>ac</sub> to read both phases. The component codes of the horizontal components here are BH1 and BH2 to indicate that the instrument is not exactly aligned with the geographic directions. In this case, the direction of orientation is given with the component's metadata and can be used to rotate into a NS-EW system.

|      |      |       |       |          |        |         |      |        |         |         |       |      |           |
|------|------|-------|-------|----------|--------|---------|------|--------|---------|---------|-------|------|-----------|
| #    | 1    | 14    | Feb   | 2013     | 13:13  | 51      | D    | 67.522 | 142.826 | 0.9     | TES   | 20   | ? 1       |
| date | hrmn | sec   | lat   | long     | depth  | no      | m    | rms    | damp    | erln    | erlt  | erdp |           |
| 13   | 214  | 1313  | 54.89 | 6748.67N | 142    | 10.4E   | 0.1  | 55     | 3       | 5.94    | 0.000 | 92.7 | 44.2 39.3 |
| stn  | dist | azm   | ain   | w        | phas   | calcp   | h    | tsec   | t-obs   | t-cal   | res   | wt   | di        |
| MAJO | 3485 | 186.1 | 42.8  | 0        | PP     | PnPn    | 1321 | 22.8   | 447.91  | 445.00  | 2.91  | 1.00 | 4         |
| MAJO | 3485 | 186.1 | 27.3  | 0        | P      | P       | 1320 | 10.0   | 375.15  | 382.25  | -7.11 | 1.00 | 2         |
| MAJO | 3485 | 186.1 | 28.1  | 0        | S      | S       | 1325 | 19.8   | 684.86  | 691.48  | -6.62 | 1.00 | 6         |
| KURK | 3858 | 273.1 | 26.8  | 0        | P      | P       | 1320 | 45.3   | 410.38  | 411.44  | -1.06 | 1.00 | 1         |
| KURK | 3858 | 273.1 | 40.6  | 0        | PP     | PnPn    | 1322 | 2.6    | 487.69  | 487.73  | -0.04 | 1.00 | 3         |
| KURK | 3858 | 273.1 | 27.7  | 0        | S      | S       | 1326 | 18.9   | 744.06  | 743.37  | 0.69  | 1.00 | 4         |
| MAKZ | 4072 | 266.2 | 26.4  | 0        | P      | P       | 1321 | 1.0    | 426.08  | 428.05  | -1.98 | 1.00 | 1         |
| MAKZ | 4072 | 266.2 | 39.7  | 0        | PP     | PnPn    | 1322 | 27.4   | 512.54  | 511.60  | 0.94  | 1.00 | 2         |
| MAKZ | 4072 | 266.2 | 46.9  | 0        | SS     | SnSn    | 1329 | 16.4   | 921.49  | 924.72  | -3.23 | 1.00 | 9         |
| XAN  | 4308 | 226.6 | 25.9  | 0        | P      | P       | 1321 | 18.1   | 443.21  | 446.13  | -2.92 | 1.00 | 0         |
| XAN  | 4308 | 226.6 | 34.8  | 0        | PP     | PP      | 1322 | 52.7   | 537.78  | 535.32  | 2.46  | 1.00 | 1         |
| XAN  | 4308 | 226.6 | 27.1  | 0        | S      | S       | 1327 | 16.3   | 801.38  | 805.45  | -4.08 | 1.00 | 2         |
| ENH  | 4687 | 224.0 | 25.2  | 0        | P      | P       | 1321 | 45.2   | 470.34  | 474.30  | -3.96 | 1.00 | 0         |
| ENH  | 4687 | 224.0 | 34.3  | 0        | PP     | PP      | 1323 | 30.7   | 575.79  | 572.36  | 3.43  | 1.00 | 1         |
| ENH  | 4687 | 224.0 | 26.5  | 0        | S      | S       | 1328 | 6.2    | 851.27  | 856.26  | -4.99 | 1.00 | 2         |
| ENH  | 4687 | 224.0 | 36.7  | 0        | SS     | SS      | 1331 | 20.0   | 1045.07 | 1046.08 | -1.01 | 1.00 | 3         |
| TATO | 4960 | 207.1 | 24.6  | 0        | P      | P       | 1322 | 5.6    | 490.73  | 494.12  | -3.39 | 1.00 | 1         |
| TATO | 4960 | 207.1 | 33.8  | 0        | PP     | PP      | 1323 | 57.4   | 602.48  | 598.76  | 3.72  | 1.00 | 2         |
| TATO | 4960 | 207.1 | 26.0  | 0        | S      | S       | 1328 | 43.4   | 888.52  | 892.20  | -3.68 | 1.00 | 3         |
| LSA  | 5373 | 244.7 | 23.7  | 0        | P      | P       | 1322 | 38.1   | 523.17  | 523.96  | -0.80 | 1.00 | 0         |
| LSA  | 5373 | 244.7 | 28.4  | 0        | PP     | PP      | 1324 | 35.9   | 641.06  | 636.87  | 4.19  | 1.00 | 1         |
| LSA  | 5373 | 244.7 | 25.3  | 0        | S      | S       | 1329 | 40.4   | 945.49  | 946.59  | -1.10 | 1.00 | 1         |
| LSA  | 5373 | 244.7 | 29.2  | 0        | SS     | SS      | 1333 | 24.0   | 1169.11 | 1158.32 | 10.78 | 1.00 | 2         |
| COR  | 5690 | 66.0  | 23.1  | 0        | P      | P       | 1322 | 57.4   | 542.53  | 545.00  | -2.47 | 1.00 | 1         |
| COR  | 5690 | 66.0  | 24.7  | 0        | S      | S       | 1330 | 21.9   | 986.97  | 985.30  | 1.67  | 1.00 | 2         |
| COR  | 5690 | 66.0  | 28.6  | 0        | SS     | SS      | 1334 | 25.7   | 1230.76 | 1203.01 | 27.74 | 1.00 | 3         |
| QIZ  | 5870 | 219.4 | 22.7  | 0        | P      | P       | 1323 | 9.3    | 554.36  | 557.21  | -2.85 | 1.00 | 0         |
| QIZ  | 5870 | 219.4 | 28.1  | 0        | PP     | PP      | 1325 | 15.6   | 680.66  | 677.02  | 3.64  | 1.00 | 1         |
| QIZ  | 5870 | 219.4 | 24.4  | 0        | S      | S       | 1330 | 41.4   | 1006.49 | 1007.80 | -1.32 | 1.00 | 2         |
| QIZ  | 5870 | 219.4 | 28.5  | 0        | SS     | SS      | 1334 | 45.3   | 1250.38 | 1228.62 | 21.76 | 1.00 | 2         |
| ESK  | 6053 | 336.5 | 22.3  | 0        | P      | P       | 1323 | 23.0   | 568.08  | 568.79  | -0.71 | 1.00 | 2         |
| ANTO | 6707 | 303.4 | 20.9  | 0        | P      | P       | 1324 | 5.6    | 610.70  | 610.87  | -0.17 | 1.00 | 2         |
| ANTO | 6707 | 303.4 | 27.5  | 0        | PP     | PP      | 1326 | 22.1   | 747.22  | 744.41  | 2.81  | 1.00 | 3         |
| ANTO | 6707 | 303.4 | 0     |          |        |         | 1327 | 47.7   | 832.8   |         |       |      |           |
| ANTO | 6707 | 303.4 | 22.8  | 0        | S      | S       | 1332 | 27.0   | 1112.14 | 1107.57 | 4.57  | 1.00 | 6         |
| PASC | 6918 | 67.4  | 20.5  | 0        | P      | P       | 1324 | 14.8   | 619.92  | 623.83  | -3.91 | 1.00 | 1         |
| PASC | 6918 | 67.4  | 22.4  | 0        | S      | S       | 1332 | 47.8   | 1132.86 | 1131.89 | 0.97  | 1.00 | 2         |
| PFO  | 7042 | 66.2  | 20.2  | 0        | P      | P       | 1324 | 22.4   | 627.56  | 631.41  | -3.85 | 1.00 | 1         |
| PFO  | 7042 | 66.2  | 22.2  | 0        | S      | S       | 1333 | 1.5    | 1146.59 | 1146.11 | 0.48  | 1.00 | 2         |
| PFO  | 7042 | 66.2  | 13.3  | 0        | SKSac  | SKSac   | 1334 | 22.1   | 1227.24 | 1226.38 | 0.86  | 1.00 | 1         |
| ANMO | 7276 | 57.2  | 19.7  | 0        | P      | P       | 1324 | 36.6   | 641.73  | 645.27  | -3.54 | 1.00 | 1         |
| ANMO | 7276 | 57.2  | 21.7  | 0        | S      | S       | 1333 | 24.0   | 1169.09 | 1172.25 | -3.16 | 1.00 | 2         |
| WVT  | 7747 | 41.4  | 18.8  | 0        | P      | P       | 1325 | 2.0    | 667.12  | 671.69  | -4.57 | 1.00 | 1         |
| WVT  | 7747 | 41.4  | 20.8  | 0        | S      | S       | 1334 | 11.7   | 1216.83 | 1222.45 | -5.62 | 1.00 | 2         |
| WVT  | 7747 | 41.4  | 27.7  | 0        | SS     | SS      | 1338 | 42.6   | 1487.75 | 1492.76 | -5.01 | 1.00 | 3         |
| HKT  | 8153 | 50.2  | 17.9  | 0        | P      | P       | 1325 | 25.0   | 690.14  | 693.79  | -3.65 | 1.00 | 1         |
| HKT  | 8153 | 50.2  | 20.0  | 0        | S      | S       | 1334 | 55.2   | 1260.27 | 1264.73 | -4.46 | 1.00 | 2         |
| SLBS | 8309 | 64.7  | 17.6  | 0        | P      | P       | 1325 | 33.8   | 698.88  | 702.32  | -3.44 | 1.00 | 1         |
| SLBS | 8309 | 64.7  | 19.7  | 0        | S      | S       | 1335 | 15.7   | 1280.78 | 1281.06 | -0.29 | 1.00 | 2         |
| SLBS | 8309 | 64.7  | 27.3  | 0        | SS     | SS      | 1340 | 12.2   | 1577.30 | 1570.17 | 7.13  | 1.00 | 3         |
| TEIG | 9412 | 46.6  | 15.2  | 0        | P      | P       | 1326 | 27.4   | 752.56  | 755.95  | -3.39 | 1.00 | 0         |
| TEIG | 9412 | 46.6  | 25.1  | 0        | PP     | PP      | 1329 | 48.6   | 953.66  | 952.02  | 1.64  | 1.00 | 1         |
| CTAO | 9733 | 176.2 | 14.4  | 0        | P      | P       | 1326 | 39.3   | 764.36  | 769.71  | -5.35 | 0.00 | 0         |
| CTAO | 9733 | 176.2 | 24.8  | 0        | PP     | PP      | 1330 | 6.6    | 971.71  | 975.50  | -3.79 | 1.00 | 2         |
| MTDJ | 9927 | 37.4  | 10.3  | 0        | SKSac  | SKSac   | 1337 | 24.1   | 1409.23 | 1409.91 | -0.68 | 1.00 | 1         |
| MTDJ | 9927 | 37.4  | 24.6  | 0        | PP     | PP      | 1330 | 25.7   | 990.79  | 989.60  | 1.18  | 1.00 | 1         |
| MTDJ | 9927 | 37.4  | 14.1  | 0        | P      | P       | 1326 | 52.7   | 777.80  | 778.52  | -0.72 | 1.00 | 0         |
| 2013 | 214  | 1313  | 54.9  | D        | 67.811 | 142.173 | 0.1  | TES    | 20      | 5.9     |       |      |           |

*Output from the location program:* Using additional phases the location changes slightly, but without readings of depth phases or P-wave onsets very near to this shallow source the depth cannot be better resolved compared to the USGS.



*Vertical component traces of stations KBS and KDAK, unfiltered:* We try to read first-motion P-wave polarities (upward motion = C = + = compressional, downward motion = D = - = dilatational) which can allow us to estimate a fault-plane solution for this earthquake. For station KBS (top) it is possible to read the polarity, which is upward (compression). For station KDAK it is not possible to read the polarity, although we were tempted to read this as a downward motion (dilatation). As seen further down, the selected data set does not have stations that are for sure in a dilatational quadrant, however, the position of all our compressional readings (see next figure) are in agreement with Global CMT fault-plane solution.

The following solution was provided by the Global CMT catalog ([www.globalcmt.org](http://www.globalcmt.org)) for this event. The solution shows an oblique thrust mechanism:

### 201302141313A EASTERN SIBERIA, RUSSIA

Date: 2013/ 2/14 Centroid Time: 13:13:58.6 GMT

[Lat= 67.71](#) [Lon= 142.62](#)

Depth= 12.0 Half duration= 5.2

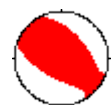
Centroid time minus hypocenter time: 5.5

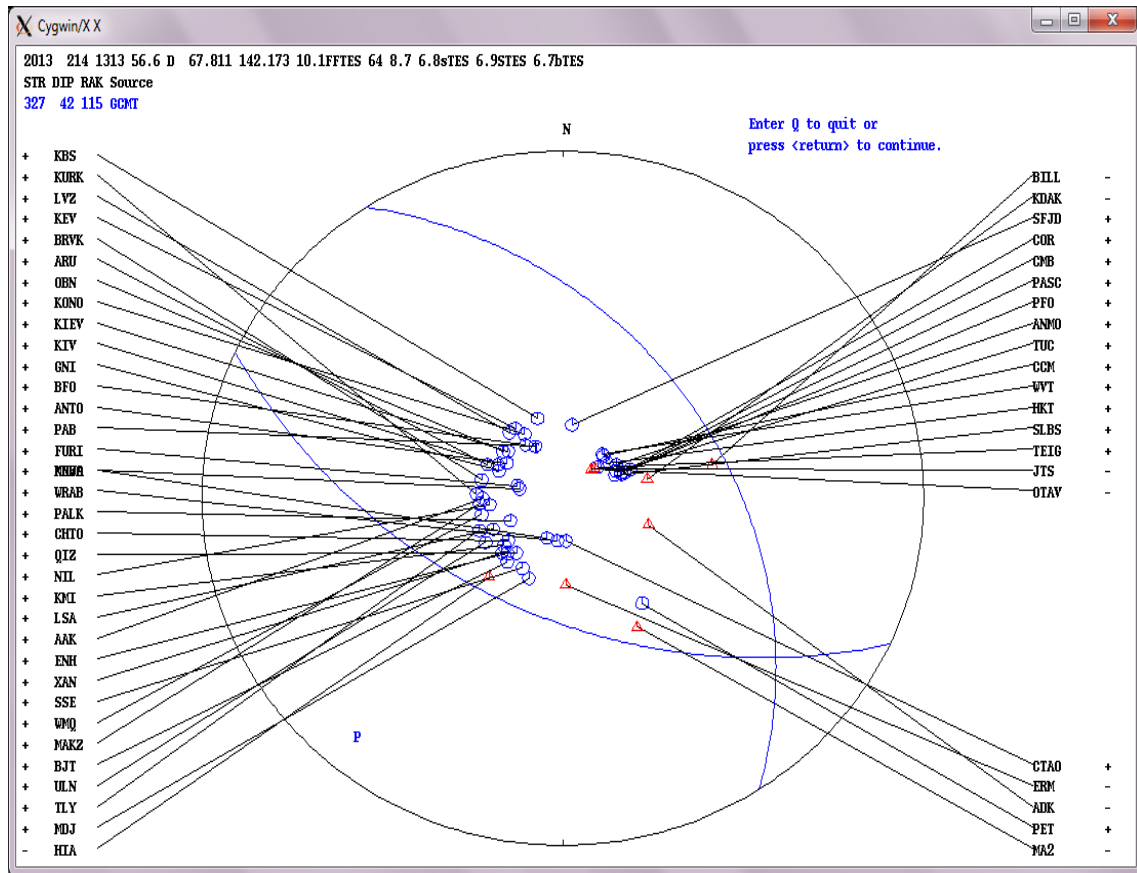
Moment Tensor: Expo=26 1.030 -0.627 -0.404 0.335 0.075 0.729

Mw = 6.7 mb = 0.0 Ms = 6.6 Scalar Moment = 1.2e+26

Fault plane: strike=327 dip=42 slip=115

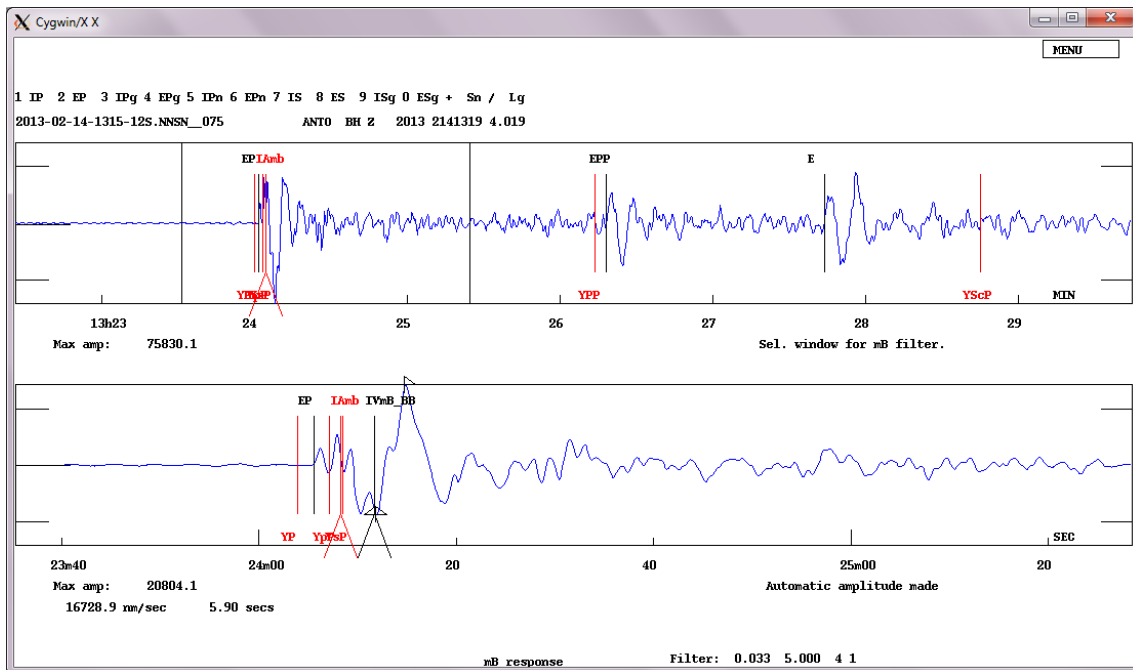
Fault plane: strike=115 dip=52 slip=69



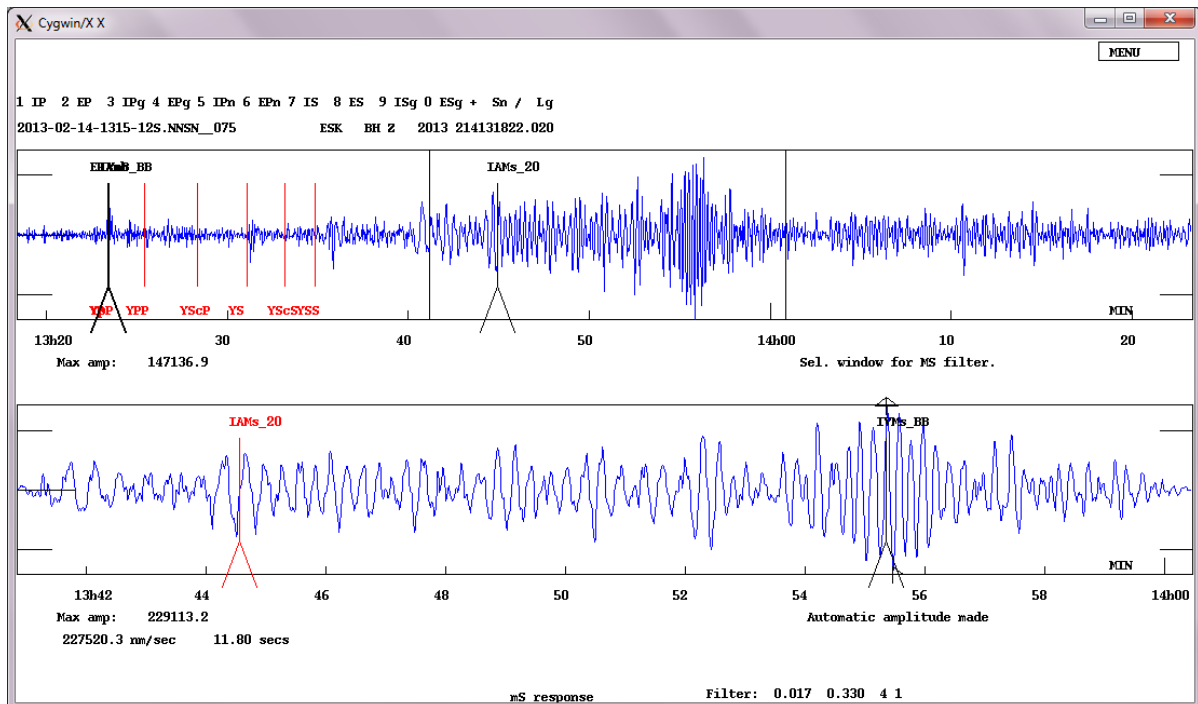


*Fault-plane solution plot:* This plot is created in SEISAN and shows the solution given by the Global CMT catalog together with polarities read from the example. We can see that all our observations lie near the center of the focal sphere and are likely compressions. However, the readings we made show compression and dilatation, and most likely the dilatational readings are wrong. Although looking at more than 100 stations from the global networks, we do not have good enough coverage of the focal sphere as all our rays to teleseismic stations only leave the source downward near the vertical. In contrast, the moment tensor solution (based on full waveforms from P, S and surface waves) normally requires only few stations to obtain a good solution. Using complementary to first-motion polarity readings in SEISAN also S/P amplitude ratios for better constraining fault-plane solutions is possible for local and teleseismic earthquakes. While the use of amplitude ratios did not help with this example, Figure 7.26 in Havskov and Ottemöller (2010) gives an example of teleseismic fault plane solution using amplitude ratios. The polarity readings made by us on individual station records (open circles = C, triangles = D) are given on the side with a link to the azimuth and angle of incidence of the corresponding phase projected onto the focal sphere.

Finally, we demonstrate the magnitude determinations for this earthquake:



Traces of station ANTO at  $\Delta = 60.3^\circ$ , mB\_BB filter: First we read amplitudes for mb and mB\_BB. Measured are the largest P-wave amplitude prior to PP, depth phases of P included, on the vertical component of simulated short-period WWSSN, respectively velocity broadband records. The position of the earlier Iamb measurement has been plotted here on the mB\_BB filtered trace only for comparison by a red arrow-bar. Details of teleseismic body-wave magnitudes are given in the regional event example in section 3 above and in IS 3.3.



Traces of station ESK at  $\Delta = 54.4^\circ$ , Ms\_BB filter: Next we determine the two IASPEI standard surface-wave magnitudes. The bottom trace shows the reading position of the Ms\_BB amplitude IVMs\_BB (black arrow-bar) on an instrument corrected broadband velocity trace.

The top trace is unfiltered and shows the time selection for the bottom trace. Note that the IAMS\_20 amplitude has already been read on another, WWSSN-LP filtered, record, before. However, the position of its measurement is shown here too for comparison (marked by a red arrow-bar). While the largest Ms\_20 displacement amplitude has a period of 21 s, the period of the 10.5 min later arriving largest Ms\_BB velocity amplitude is 11.80 s only. For more details on the determination of teleseismic surface-wave magnitude see the regional event example in section 3 as well as IS 3.3.

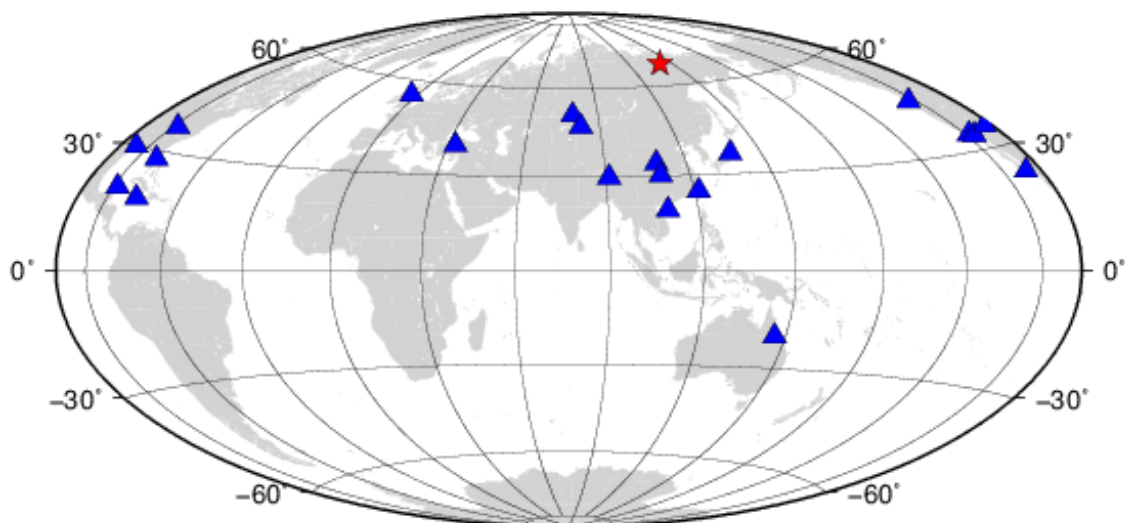
Computing the average magnitudes from all amplitude observations we get the values below:

| <b>Magnitude scale</b> | <b>Value</b> |
|------------------------|--------------|
| M <sub>w</sub> (USGS)  | 6.9          |
| m <sub>b</sub>         | 6.7          |
| m <sub>B_BB</sub>      | 7.3          |
| M <sub>s_20</sub>      | 6.9          |
| M <sub>s_BB</sub>      | 6.9          |

In this case the surface wave magnitudes Ms\_20 and Ms\_BB agree perfectly with Mw(USGS) and are equal amongst each other although Ms\_BB has been measured at a much shorter period and a later time than Ms\_20. mB\_BB is significantly larger than mb, and somewhat larger than the two Ms and the Mw magnitude, as expected for earthquakes of Mw = 6.9 according to Utsu (2002) and Fig. 3.70 in section 3.2.9.1 of the forthcoming revised Chapter 3).

The amplitudes and periods that were read for computing the magnitudes are summarized in the following table. Note that here the abridged internal magnitude nomenclature used in SEISAN stands for: mb = mb, mB = mB\_BB, Ms = Ms\_20 and MS = Ms\_BB.

|      |    |       |        |      |          |    |      |      |     |
|------|----|-------|--------|------|----------|----|------|------|-----|
| XAN  | BZ | dist: | 4308.0 | amp: | 1498.7   | T: | 1.8  | mb = | 6.3 |
| XAN  | BZ | dist: | 4308.0 | amp: | 22757.5  | T: | 8.8  | mB = | 7.0 |
| ENH  | BZ | dist: | 4687.0 | amp: | 959.5    | T: | 1.5  | mb = | 6.3 |
| ENH  | BZ | dist: | 4687.0 | amp: | 17567.1  | T: | 8.8  | mB = | 6.9 |
| ENH  | BZ | dist: | 4687.0 | amp: | 120000.0 | T: | 10.9 | MS = | 7.3 |
| ENH  | BZ | dist: | 4687.0 | amp: | 184000.0 | T: | 18.6 | MS = | 7.0 |
| TATO | BZ | dist: | 4960.0 | amp: | 436.4    | T: | 0.9  | mb = | 6.3 |
| TATO | BZ | dist: | 4960.0 | amp: | 20681.6  | T: | 9.0  | mB = | 7.2 |
| TATO | BZ | dist: | 4960.0 | amp: | 111000.0 | T: | 18.9 | MS = | 6.8 |
| TATO | BZ | dist: | 4960.0 | amp: | 81389.9  | T: | 11.8 | MS = | 7.2 |
| COR  | BZ | dist: | 5690.0 | amp: | 18587.7  | T: | 3.8  | mB = | 7.2 |
| COR  | BZ | dist: | 5690.0 | amp: | 778.2    | T: | 1.1  | mb = | 6.5 |
| COR  | BZ | dist: | 5690.0 | amp: | 11899.5  | T: | 15.7 | MS = | 6.4 |
| COR  | BZ | dist: | 5690.0 | amp: | 31700.7  | T: | 18.9 | MS = | 6.4 |
| ESK  | BZ | dist: | 6053.0 | amp: | 6275.4   | T: | 1.1  | mb = | 7.6 |
| ESK  | BZ | dist: | 6053.0 | amp: | 171000.0 | T: | 8.5  | mB = | 8.2 |
| ESK  | BZ | dist: | 6053.0 | amp: | 336000.0 | T: | 20.8 | MS = | 7.4 |
| ESK  | BZ | dist: | 6053.0 | amp: | 228000.0 | T: | 11.8 | MS = | 7.7 |
| ANTO | BZ | dist: | 6707.0 | amp: | 721.5    | T: | 1.2  | mb = | 6.7 |
| ANTO | BZ | dist: | 6707.0 | amp: | 16728.9  | T: | 5.9  | mB = | 7.3 |
| ANTO | BZ | dist: | 6707.0 | amp: | 57305.5  | T: | 19.2 | MS = | 6.7 |
| ANTO | BZ | dist: | 6707.0 | amp: | 20010.4  | T: | 20.0 | MS = | 6.8 |
| ANMO | BZ | dist: | 7276.0 | amp: | 9820.2   | T: | 3.8  | mB = | 7.2 |
| ANMO | BZ | dist: | 7276.0 | amp: | 892.4    | T: | 2.0  | mb = | 6.6 |
| ANMO | BZ | dist: | 7276.0 | amp: | 18195.2  | T: | 21.0 | MS = | 6.8 |
| ANMO | BZ | dist: | 7276.0 | amp: | 59228.9  | T: | 21.3 | MS = | 6.8 |
| WVT  | BZ | dist: | 7747.0 | amp: | 68435.1  | T: | 20.2 | MS = | 6.9 |
| WVT  | BZ | dist: | 7747.0 | amp: | 20200.8  | T: | 19.0 | MS = | 6.9 |
| DWPF | BZ | dist: | 8770.0 | amp: | 1956.5   | T: | 2.2  | mb = | 6.8 |
| DWPF | BZ | dist: | 8770.0 | amp: | 11188.1  | T: | 3.6  | mB = | 7.1 |
| DWPF | BZ | dist: | 8770.0 | amp: | 74362.7  | T: | 22.0 | MS = | 7.0 |
| DWPF | BZ | dist: | 8770.0 | amp: | 22267.2  | T: | 23.6 | MS = | 7.0 |
| TEIG | BZ | dist: | 9412.0 | amp: | 12224.9  | T: | 2.9  | mB = | 7.3 |
| TEIG | BZ | dist: | 9412.0 | amp: | 853.1    | T: | 1.4  | mb = | 6.8 |
| TEIG | BZ | dist: | 9412.0 | amp: | 18134.1  | T: | 21.4 | MS = | 7.0 |
| TEIG | BZ | dist: | 9412.0 | amp: | 58385.9  | T: | 20.6 | MS = | 7.0 |
| CTAO | BZ | dist: | 9733.0 | amp: | 686.2    | T: | 2.0  | mb = | 6.6 |
| CTAO | BZ | dist: | 9733.0 | amp: | 4437.9   | T: | 19.5 | MS = | 6.4 |
| CTAO | BZ | dist: | 9733.0 | amp: | 13972.4  | T: | 21.4 | MS = | 6.3 |



Map showing the epicenter (red star) and stations (blue triangles) used here.

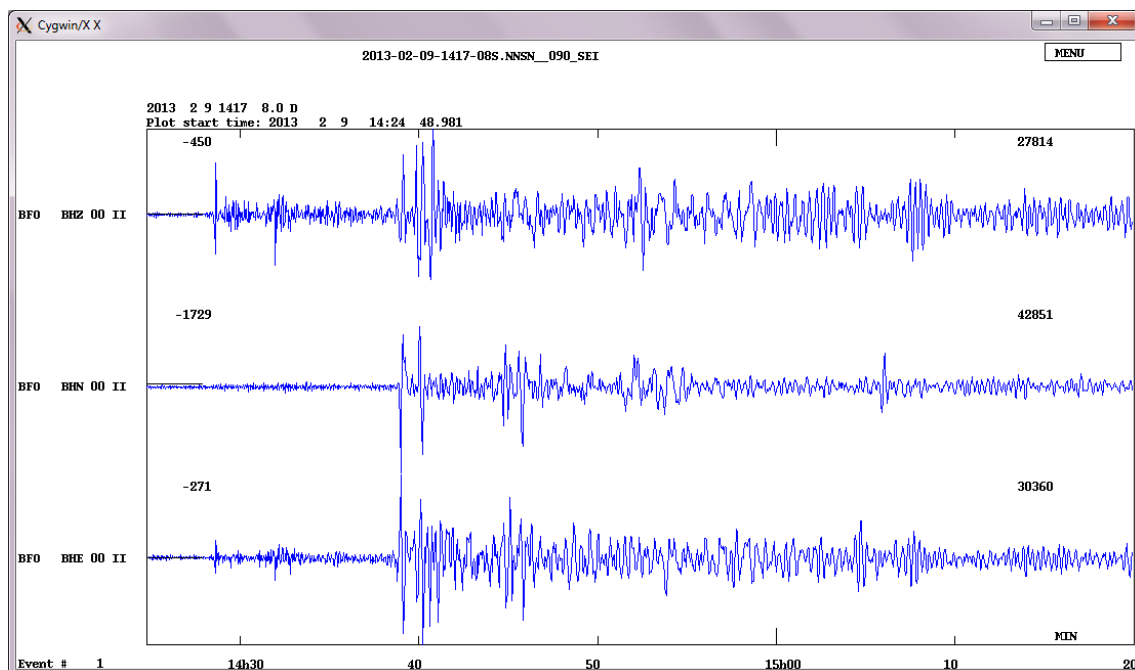
## 5 Analysis of an intermediate depth earthquake at teleseismic distances

The earthquake chosen here as example occurred in Colombia at an intermediate depth of 129 km (depth by USGS). We extracted data from IRIS that were recorded by the Global Seismograph Network (GSN). All data used here were recorded on broadband stations.

The earthquake parameters provided by the USGS are:

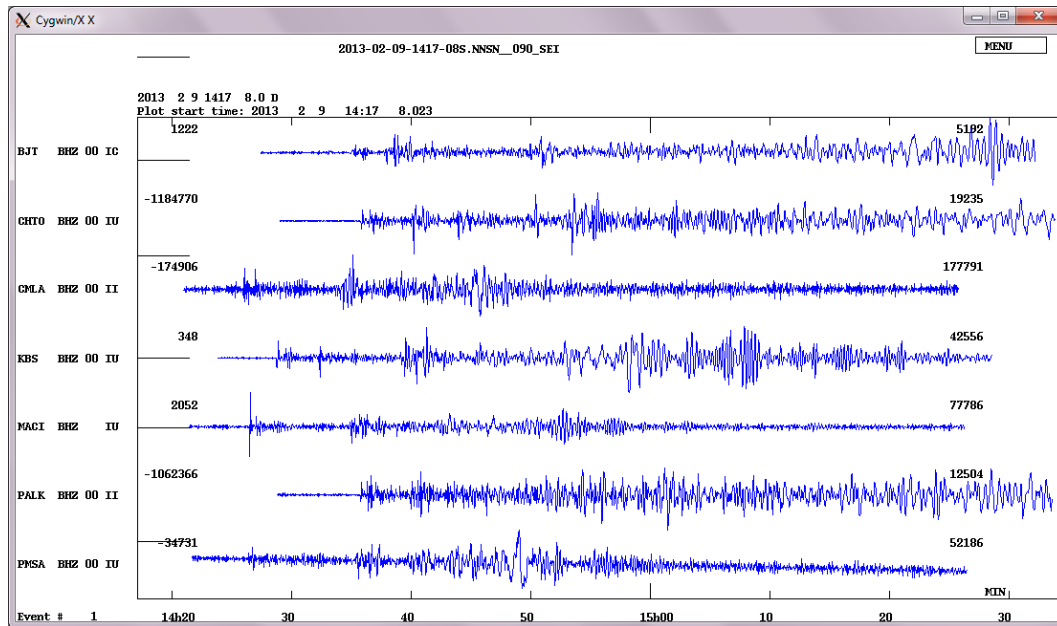
|              |                  |
|--------------|------------------|
| Date:        | 09/02/2013       |
| Origin time: | 14:16:06.3 (UTC) |
| Latitude:    | 1.14°N           |
| Longitude:   | 77.36°W          |
| Depth:       | 129 km           |
| Mw:          | 7.0              |

We now show the step-by-step procedure to analyze the earthquake.

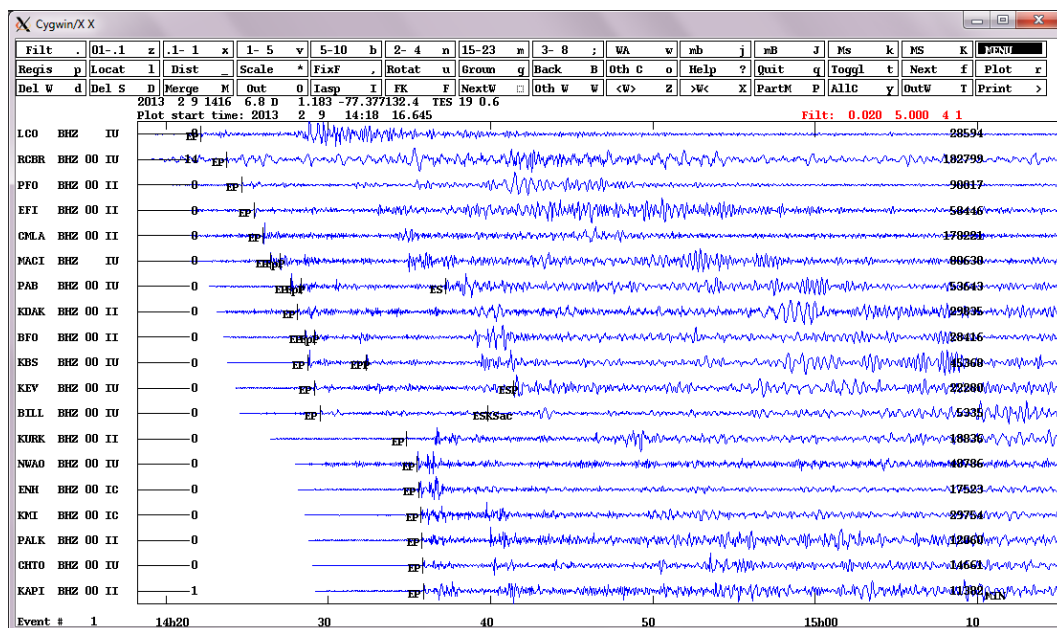


*Three-component trace plot of station BFO at  $\Delta = 86.4^\circ$ , unfiltered:* Looking at one of the stations, we can see that we are dealing with an earthquake at teleseismic distance as the record shown extends to almost one hour. The seismograms show very clear body-wave phase arrivals, which we will try to identify. The lack of more long-period and dispersed surface waves, however, points at a deep hypocenter. We will start to pick first P arrivals to get an initial estimate of the location, then we try to identify depth phases and other later phases for improved hypocenter determination and finally we use the 3-component record of station BFO only for a single-station event location.

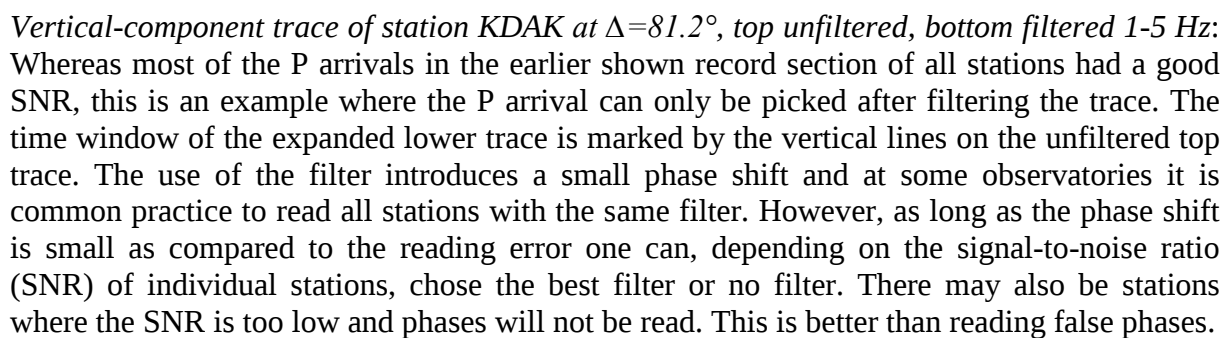




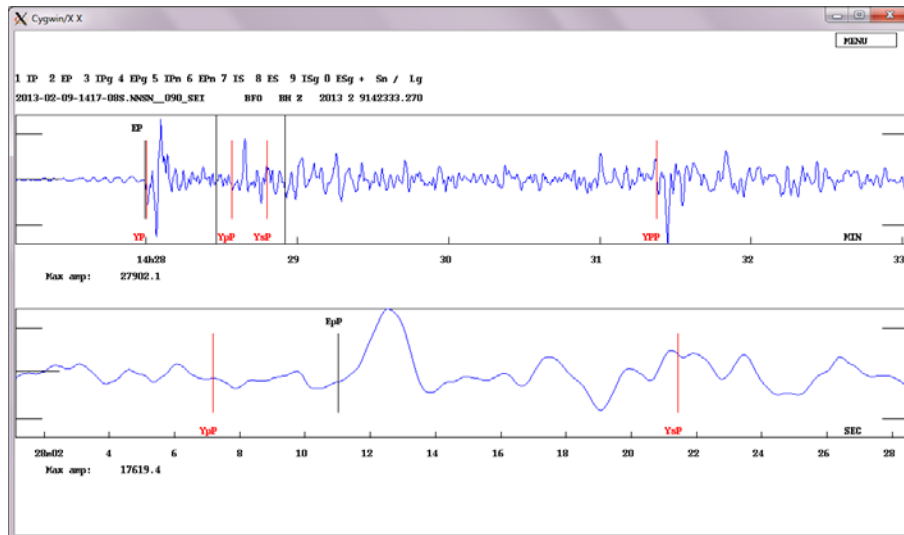
*Multi-trace plot of vertical components, unfiltered:* Shown is only a selection of the stations that are part of the sample data set for this event.



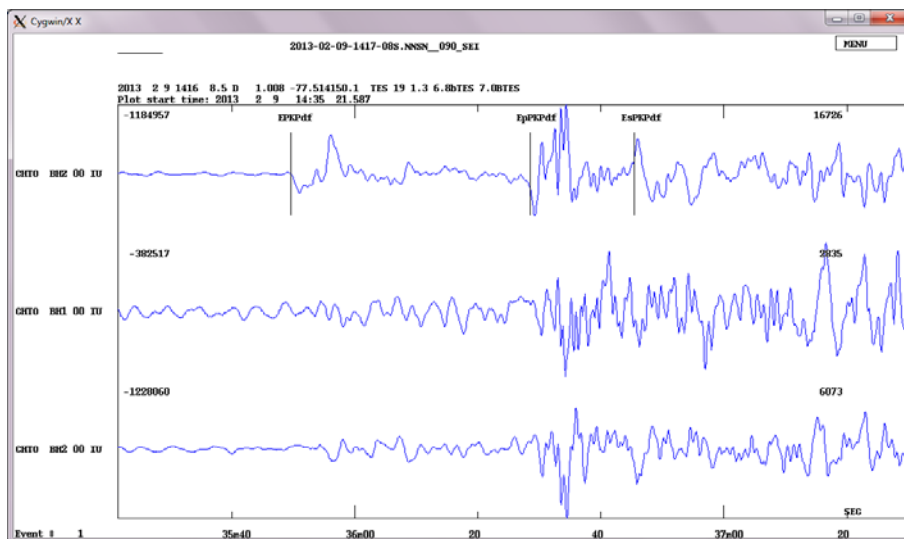
*Vertical-component traces for which phases are picked, filtered 0.02-5 Hz:* These traces are sorted by epicentral distance, ranging from  $30.7^\circ$  to  $162.7^\circ$ . Beside the P first arrival, at some of the traces also clear later arrivals have been marked and identified. Note that from station KURK downward the strong delay of the first arriving longitudinal waves is due to the Earth core shadow for direct P waves. In fact, the group of later arrivals, annotated here as P, relate to different branches of PKP core phases and likely also their depth phases. Also note several distinct onsets following within the first minute after the P onset and well before the arrival of PP. For a first event location we will only use the P and PKP first arrivals. Later, we will zoom into the first few minutes of records in order to find depth phases of P and PKP.



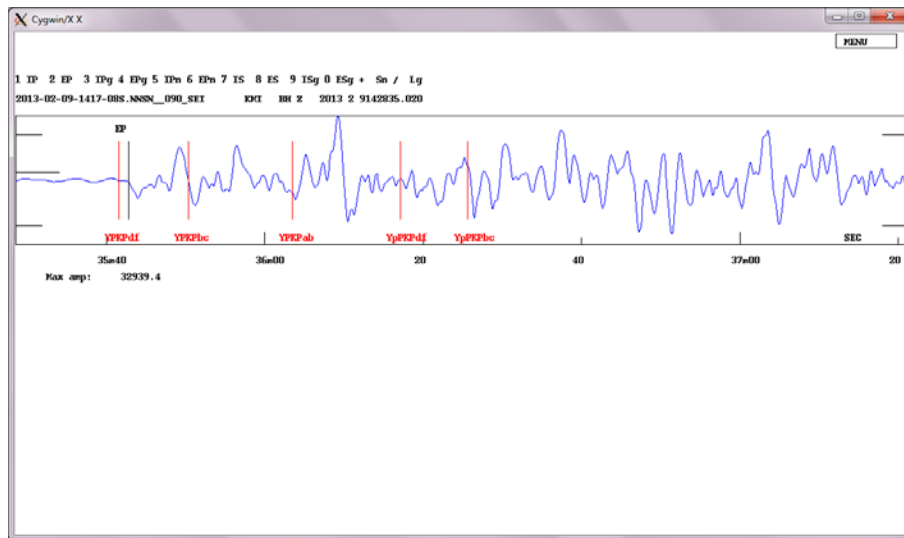
*Output from the location program on the previous page:* Locating with the first P arrivals only, we obtain a hypocenter location (1.183°N, 77.377°W and depth = 132.4 km) that is very close to the one given by the USGS. The depth estimate is perhaps better than expected in this case without the use of depth phases. However, we will now try to use the depth phases we noticed on a number of seismograms. We have so far read all first arrivals as P. The location program determines which phase arrives first. In our example we have P (until station KEV), Pdif (station BILL) and PKPdif (from station KURK onward).



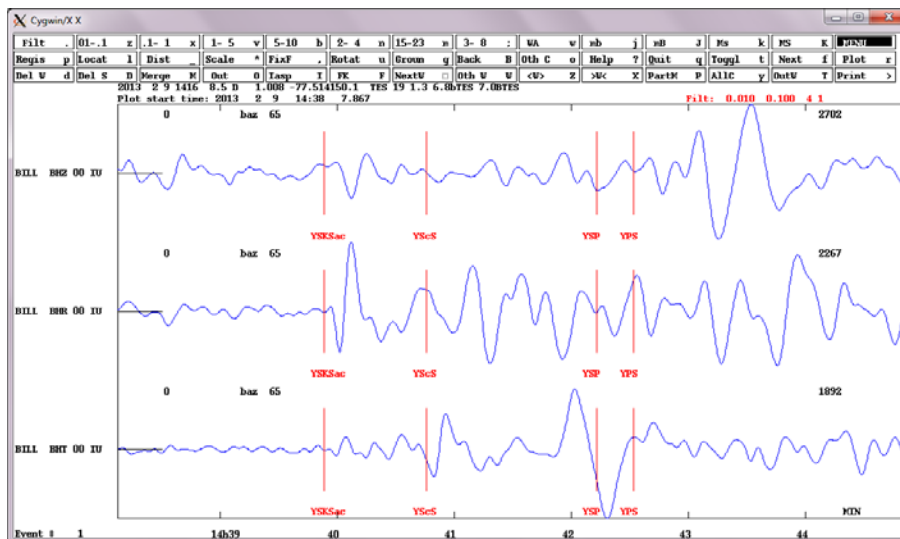
*Vertical-component trace for station BFO at  $\Delta = 86.4^\circ$ , unfiltered:* For this station we have computed synthetic arrival times based on our initial location, which are marked with red bars. We can see distinct phases arriving near the computed pP, sP and PP times. The bottom trace shows the picking of pP about 4 s after the predicted arrival, but there is no clear new energy arrival around the expected sP onset. Note, however, that sP may sometimes be even a stronger phase than pP. The amplitude relationship between these two depth phases depends on the radiation pattern of the source mechanism, which is different for P and S waves (see Chapter 3, section 3.4.1, Figs. 3.100-104), and on the different take-off angles of pP and sP from the source (see Chapter 2, section 2.6.3 with figures). From the measured travel-time difference of 38.2 s between pP and P we estimate a source depth of about 150 km, some 20 km deeper than the source depth calculated by the USGS.



*Three-component trace of station CHTO at  $\Delta = 160.0^\circ$ , unfiltered:* Knowing from the first location of the earthquake that the station is  $160^\circ$  away the first arrival at 35m50s is clearly not P, as originally marked, but PKPdf. The larger onset about 5 s later is a second phase of the same event, yet the much stronger sharp wave onset at about 36m28s is obviously the depth phase pPKPdf and the onset at about 36m45s the depth phase sPKPdf. Note that the travel-time difference pPKPdf-PKPdf of 38 s perfectly agrees with the time difference pP-P = 38.2 s measured at station BFO. Note that the differential times such as pP-P and sP-P should be near constant over a distance range.



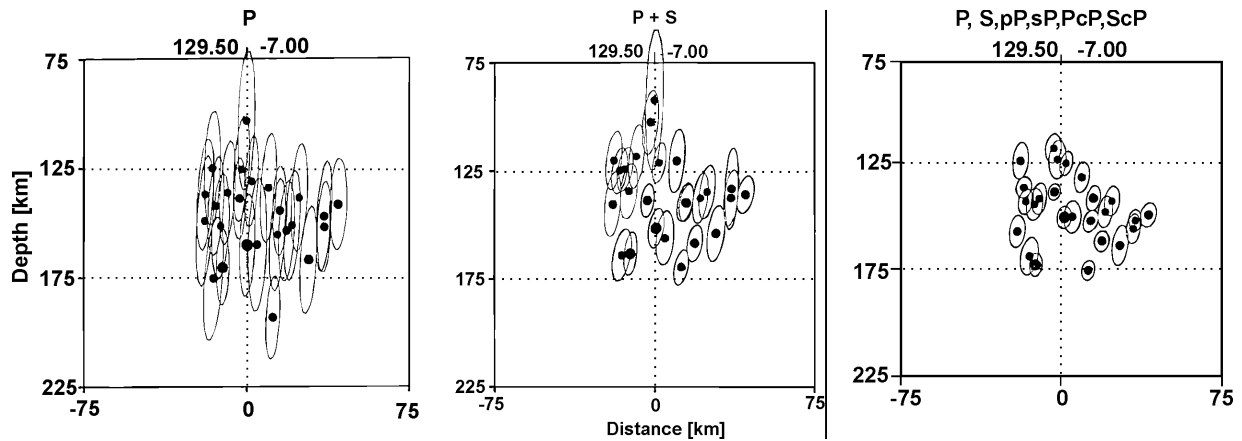
*Vertical-component trace for station KMI at  $\Delta=154.0^\circ$ , unfiltered:* In this example we can see onsets that relate to the different travel-time branches of PKP. The corresponding depth phase times are also shown and phases are visible. Core phases are often most clear in the frequency band 1-5 Hz.



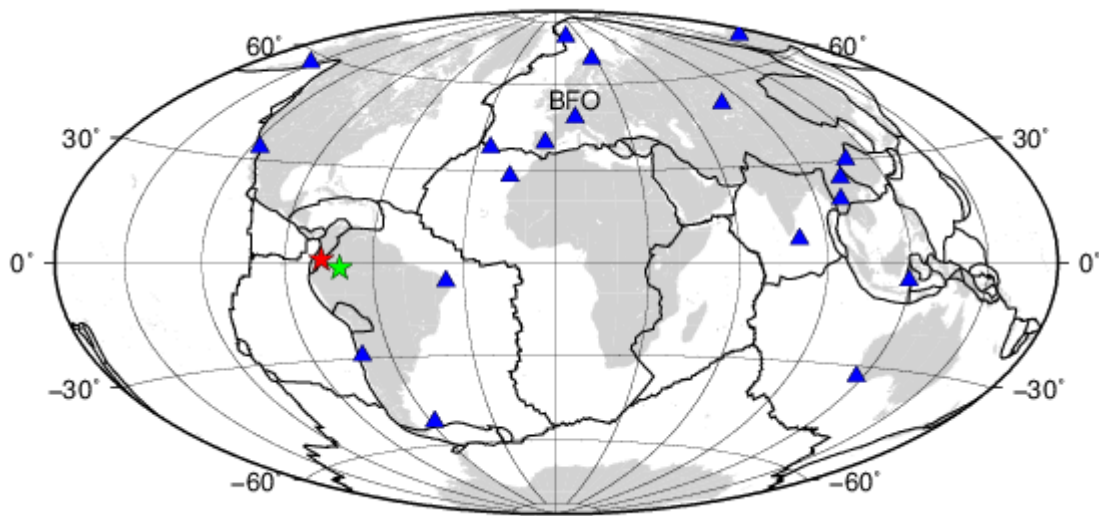
*Three-component traces for station BILL at  $\Delta = 98.6^\circ$ , rotated and filtered 0.01-0.1 Hz:* The two horizontal components have been rotated into radial and transverse components using a backazimuth of  $65^\circ$ . SKS has been back-converted from a P wave through the outer core, termed K, into an S wave through the mantle and is, therefore, polarised in the vertical plane of propagation. Accordingly, it has its largest amplitude in the radial component R (middle trace) and not on the transverse component T. It is also visible on the vertical component. This is not the rule for teleseismic S and illustrates that rotating seismograms can be useful in identifying seismic phases (see Figs. 11.16, 11.48 and 11.67 in Chapter 11 of NMSOP-2, Bormann et al., 2013). Having an array of stations, the different apparent velocities of seismic phases can also be used for identification (see, e.g., Chapter 9, and for the identification of core phases Fig. 11.46).

|        |       |                  |         |    |         |              |      |         |         |         |         |      |      |
|--------|-------|------------------|---------|----|---------|--------------|------|---------|---------|---------|---------|------|------|
| #      | 1     | 9 Feb 2013 14:16 | 8       | D  | 1.008   | -77.514150.1 | 1.3  | 6.8bTES | 19      | ? 1     |         |      |      |
| date   | hrmn  | sec              | lat     |    | long    | depth        | no   | m       | rms     | damp    | erln    | erlt | erdp |
| 13 2 9 | 1416  | 8.59             | 059.46N | 77 | 28.5W   | 149.5        | 28   | 3       | 1.16    | 0.000   | 21.6    | 17.2 | 5.7  |
| stn    | dist  | azm              | ain     | w  | phas    | calcp        | h    | tsec    | t-obs   | t-cal   | res     | wt   | di   |
| LCO    | 3398  | 168.3            | 40.0    | 0  | P       | P            | 1422 | 9.6     | 360.99  | 359.70  | 1.29    | 1.00 | 10   |
| RCBR   | 4682  | 99.8             | 36.3    | 0  | P       | P            | 1423 | 45.0    | 456.39  | 457.12  | -0.72   | 1.00 | 7    |
| PFO    | 5437  | 315.8            | 33.9    | 0  | P       | P            | 1424 | 39.4    | 510.86  | 510.52  | 0.35    | 1.00 | 7    |
| EFI    | 6121  | 165.4            | 31.7    | 0  | P       | P            | 1425 | 25.1    | 556.49  | 555.86  | 0.63    | 1.00 | 7    |
| EFI    | 6121  | 165.4            |         | 0  | IVmB_BB |              | 1425 | 28.3    | 559.7   |         |         |      |      |
| EFI    | 6121  | 165.4            |         | 0  | IAmb    |              | 1426 | 9.7     | 601.1   |         |         |      |      |
| CMLA   | 6687  | 46.1             | 29.9    | 0  | P       | P            | 1425 | 59.1    | 590.53  | 591.50  | -0.97   | 1.00 | 2    |
| CMLA   | 6687  | 46.1             |         | 0  | IVmB_BB |              | 1426 | 3.6     | 595.0   |         |         |      |      |
| CMLA   | 6687  | 46.1             |         | 0  | IAmb    |              | 1426 | 4.2     | 595.6   |         |         |      |      |
| MACI   | 7140  | 59.0             | 28.5    | 0  | P       | P            | 1426 | 27.2    | 618.59  | 618.76  | -0.17   | 1.00 | 2    |
| MACI   | 7140  | 59.0             | 151.0   | 0  | pP      | pP           | 1427 | 3.7     | 655.14  | 654.18  | 0.96    | 1.00 | 5    |
| PAB    | 8504  | 49.6             | 24.3    | 0  | P       | P            | 1427 | 40.6    | 691.97  | 693.14  | -1.17   | 1.00 | 1    |
| PAB    | 8504  | 49.6             | 155.3   | 0  | pP      | pP           | 1428 | 18.2    | 729.56  | 729.75  | -0.19   | 1.00 | 5    |
| PAB    | 8504  | 49.6             | 26.2    | 0  | S       | S            | 1437 | 15.4    | 1266.86 | 1266.68 | 0.18    | 1.00 | 8    |
| KDAK   | 9045  | 328.4            | 22.7    | 0  | P       | P            | 1428 | 6.6     | 717.97  | 719.55  | -1.57   | 1.00 | 3    |
| BFO    | 9626  | 41.8             | 20.8    | 0  | P       | P            | 1428 | 32.9    | 744.36  | 746.02  | -1.65   | 1.00 | 1    |
| BFO    | 9626  | 41.8             |         | 0  | IVmB_BB |              | 1428 | 37.2    | 748.6   |         |         |      |      |
| BFO    | 9626  | 41.8             |         | 0  | IAmb    |              | 1428 | 37.9    | 749.4   |         |         |      |      |
| BFO    | 9626  | 41.8             | 158.8   | 0  | pP      | pP           | 1429 | 11.1    | 782.48  | 783.45  | -0.96   | 1.00 | 4    |
| KBS    | 9899  | 11.1             | 20.0    | 0  | P       | P            | 1428 | 45.8    | 757.23  | 757.65  | -0.42   | 1.00 | 1    |
| KBS    | 9899  | 11.1             |         | 0  | IAmb    |              | 1428 | 49.9    | 761.4   |         |         |      |      |
| KBS    | 9899  | 11.1             |         | 0  | IVmB_BB |              | 1428 | 51.9    | 763.3   |         |         |      |      |
| KBS    | 9899  | 11.1             | 35.6    | 0  | PP      | PP           | 1432 | 20.5    | 971.87  | 968.68  | 3.20    | 1.00 | 2    |
| KEV    | 10472 | 19.7             | 19.5    | 0  | P       | P            | 1429 | 8.0     | 779.44  | 781.44  | -2.00   | 1.00 | 0    |
| KEV    | 10472 | 19.7             |         | 0  | IAmb    |              | 1429 | 11.9    | 783.3   |         |         |      |      |
| KEV    | 10472 | 19.7             |         | 0  | IVmB_BB |              | 1429 | 13.5    | 784.9   |         |         |      |      |
| KEV    | 10472 | 19.7             | 26.5    | 0  | SP      | SP           | 1441 | 25.8    | 1517.18 | 1514.82 | 2.36    | 1.00 | 6    |
| BILL   | 10982 | 340.0            | 18.9    | 0  | P       | Pdif         | 1429 | 29.2    | 800.57  | 801.82  | -1.25   | 1.00 | 2    |
| BILL   | 10982 | 340.0            |         | 0  | IVmB_BB |              | 1429 | 32.3    | 803.7   |         |         |      |      |
| BILL   | 10982 | 340.0            |         | 0  | IAmb    |              | 1429 | 35.1    | 806.5   |         |         |      |      |
| BILL   | 10982 | 340.0            | 11.7    | 0  | SKSac   | SKSac        | 1439 | 53.6    | 1425.02 | 1425.40 | -0.37   | 1.00 | 4    |
| KURK   | 13871 | 18.2             | 8.0     | 0  | P       | PKPdf        | 1434 | 49.3    | 1120.70 | 1121.34 | -0.64   | 1.00 | 0    |
| NWAO   | 16181 | 202.1            | 13.2    | 0  | P       | PKPbc        | 1435 | 28.6    | 1159.98 | 1160.26 | -0.28   | 1.00 | 3    |
| ENH    | 16499 | 348.5            | 6.8     | 0  | P       | PKPdf        | 1435 | 33.2    | 1164.59 | 1164.16 | 0.44    | 1.00 | 1    |
| KMI    | 17148 | 359.6            | 5.9     | 0  | P       | PKPdf        | 1435 | 42.8    | 1174.19 | 1173.26 | 0.94    | 1.00 | 1    |
| PALK   | 17448 | 69.0             | 5.4     | 0  | P       | PKPdf        | 1435 | 45.8    | 1177.26 | 1176.64 | 0.62    | 1.00 | 1    |
| CHTO   | 17812 | 10.0             | 4.8     | 0  | PKPdf   | PKPdf        | 1435 | 49.6    | 1181.04 | 1180.59 | 0.45    | 1.00 | 1    |
| CHTO   | 17812 | 10.0             | 175.2   | 0  | pPKPdf  | pPKPdf       | 1436 | 28.5    | 1219.87 | 1220.40 | -0.53   | 1.00 | 5    |
| CHTO   | 17812 | 10.0             | 177.3   | 0  | sPKPdf  | sPKPdf       | 1436 | 45.4    | 1236.84 | 1235.95 | 0.89    | 1.00 | 10   |
| KAPI   | 18072 | 256.6            | 4.2     | 0  | P       | PKPdf        | 1435 | 52.2    | 1183.62 | 1183.05 | 0.57    | 1.00 | 2    |
| 2013   | 2 9   | 1416             | 8.6     | D  | 0.991   | -77.475149.5 | TES  | 19      | 1.2     | 6.8bTES | 7.0BTES |      |      |

*Output from the location program:* Adding three pP phases and one pPKPdf and sPKPdf each, the hypocenter depth changes (shown in the third and last line) from 132 to 149 km. While this is now more different from the solution given by the USGS, we take this as the best solution determined from our data set, which may be even better than the USGS solution, because the latter commonly uses only P-wave first arrivals for their hypocenter location. Yet, the reading, proper identification and time picking of later arriving phases is crucial for significantly reducing the errors in hypocenter depth estimates (see Figure below).



**Left:** hypocenter locations using only P phases; **middle:** by adding S phases; **right:** by adding also depth phases and core reflections. (Copy of Figure 7 from IS 11.1, modified from Schöffel and Das, J. Geophys. Res., Vol. 104, No. B6, page 13, Figure 2; © 1999, by permission of American Geophysical Union).

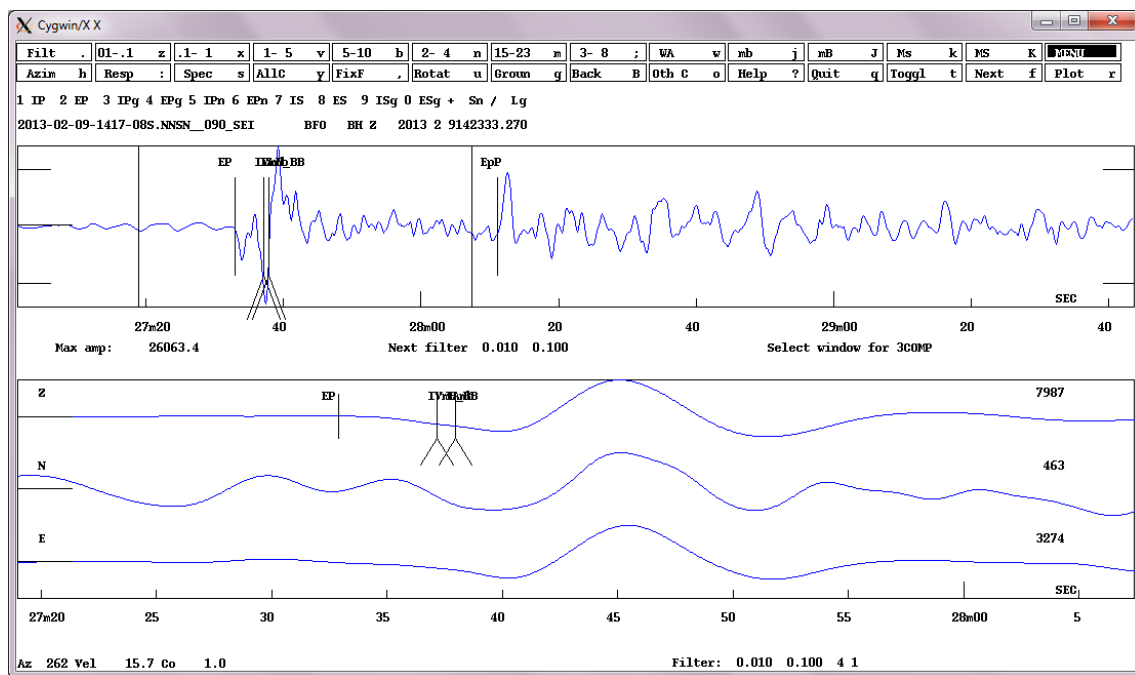


Map showing the epicenter of the Colombia earthquake (red star) determined by using a network of globally distributed stations (blue triangles). For comparison we plotted the epicenter from the single station (BFO, see label) 3-component location as a green star. For details of the BFO location see the next section.

## 6 Single station 3-component source location

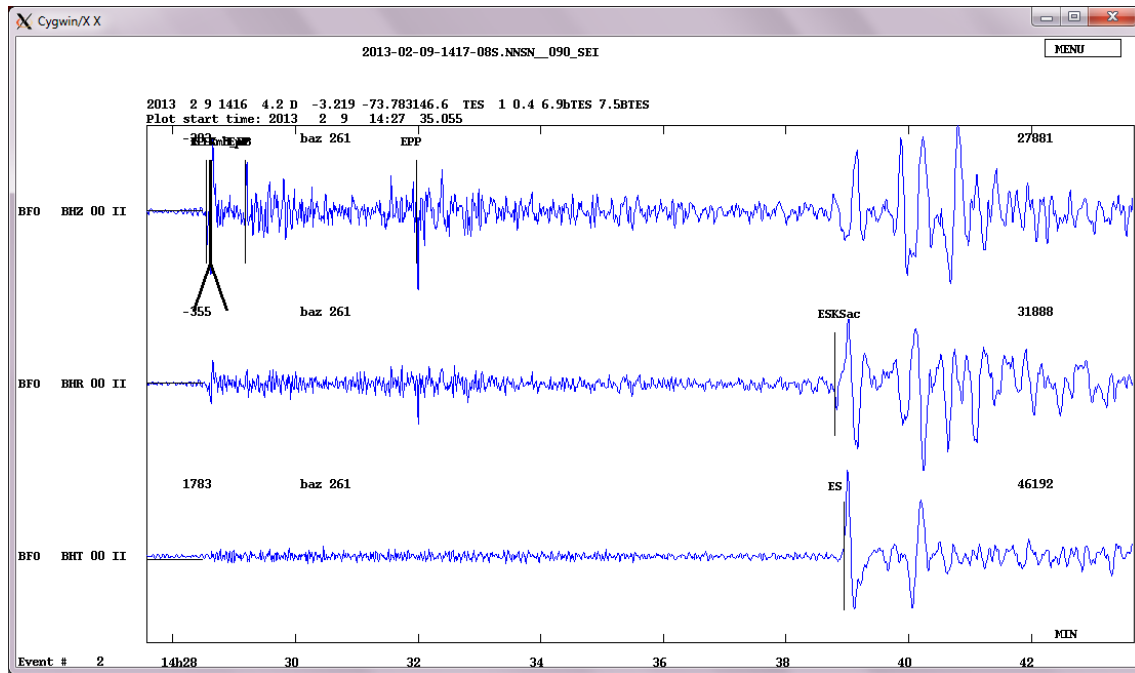
So far we have demonstrated how data from a global network can be processed in SEISAN. This is nowadays the common procedure as the data from a large number of stations are now freely accessible via Internet. However, it is also possible with SEISAN to estimate the event location by analyzing only the three-component records of a single station. The epicentral distance of the station can be determined best from the time difference S-P, but also using time differences of P to other later phases such as PP, SP or even SS (see Fig. 11.23 of Chapter 11 in Bormann2009, and EX 11.2). SKS-P is not suitable, because this time difference does not change much with distance. The backazimuth can be determined from the amplitude ratios of the three components (see Figure 1 and text in IS 11.1 and EX 11.2). For the backazimuth calculation the method of Roberts et al. (1989) is implemented in SEISAN.

We start with the determination of the back-azimuth from the P-wave first-motion amplitudes, as we need the backazimuth to rotate the seismograms into the ZRT system, which we will use to identify the phases.



*Three-component analysis of station BFO at  $\Delta = 86.4^\circ$ , filtered 0.01-0.1 Hz:* From the above trace plot we see a positive correlation between the vertical and the two horizontal components. By definition this means that first motion direction of P in the horizontal components shows toward the source, which is in the south-west. However, since the N-S amplitude is much smaller than the E-W amplitude (compare the numbers on the outermost right side of the traces), we can further conclude that the ray is coming more from the west than from the south thus matching for station BFO a backazimuth of  $262^\circ$ .

In the next step we estimate the epicentral distance via the time difference between P and identified later arriving phases.



*Three-component analysis of station BFO at  $\Delta = 86.4^\circ$ , unfiltered:* From the time difference between P with PP and S (here identified as SKS on the middle trace and as the slightly later arriving S on the bottom trace) we estimate the epicentral distance, assuming a source depth of 150 km according to the pP-P time (see Table below). This depth estimate agrees reasonably well with interpreting the strong onsets that follow SKS. S arrives about 13 s after SKS and the depth phase sS 74 s after S, corresponding at  $\Delta = 86.4^\circ$  to a computed source depth of 170 km according to the IASP91 tables. This gives some feeling for the possible influence of errors in the assumed velocity model, phase interpretation and/or onset-time piking on the estimation of source depth.

We can either read the distance for which we get the best match between our observations and the travel time curves of the global IASP91 velocity model, or equivalently estimate it with a grid search by computing travel times for a range of distances as done here. In our example we then get  $\Delta = 86.5^\circ$  (see Table below). This differs only  $0.15^\circ$  from our first multi-station P-wave location ( $86.35^\circ$ ) and even less from our improved multi-station and multi-phase location ( $86.56^\circ$ ) (see above location program outputs). Together with the backazimuth we can now locate the event on a suitable stereographic map with station BFO in the center (as in Figure 12 of EX 11.2 for an Ecuador earthquake and station CLL).

| Phase difference | Time difference $\delta t(s)$ | $\delta t(s)$ for $\Delta = 86^\circ$ | $\delta t(s)$ for $\Delta = 87^\circ$ |
|------------------|-------------------------------|---------------------------------------|---------------------------------------|
| PP-P             | 205.7                         | 202                                   | 205                                   |
| S-P              | 623.3                         | 622                                   | 626                                   |



```
#      2  9 Feb 2013 14:16  1  D                                     ? 1

date hrmn  sec      lat      long depth  no m      rms  damp erln erlt erdp
13 2 9 1416  4.23 312.63S 73 47.5W 146.6   5 2    0.37 0.000552.7488.0 0.0
stn  dist  azm  ain w phas  calcp hrmn tsec  t-obs  t-cal  res  wt di
BFO  9700 41.4 14.4 0 SKSac  SKSac 1438 47.31363.051362.11 0.94 0.98*12
BFO  9700 41.4 35.9 0 PP      PP    1431 58.6 954.40 954.80 -0.39 1.00*22
BFO  9700 41.4      0      1428 35.3 751.1
BFO  9700 41.4159.1 0 pP      pP    1429 11.1 786.84 786.42 0.42 1.00*22
BFO  9700 41.4      0 IAMB      1428 37.9 753.7
BFO  9700 41.4      0 IVMB_BB 1428 37.2 753.0
BFO  9700 41.4 20.4 0 P      P    1428 32.9 748.67 749.64 -0.97 0.97*11
BFO  9700 41.4      AZ      262.0 262.0 0.00 0.20 33

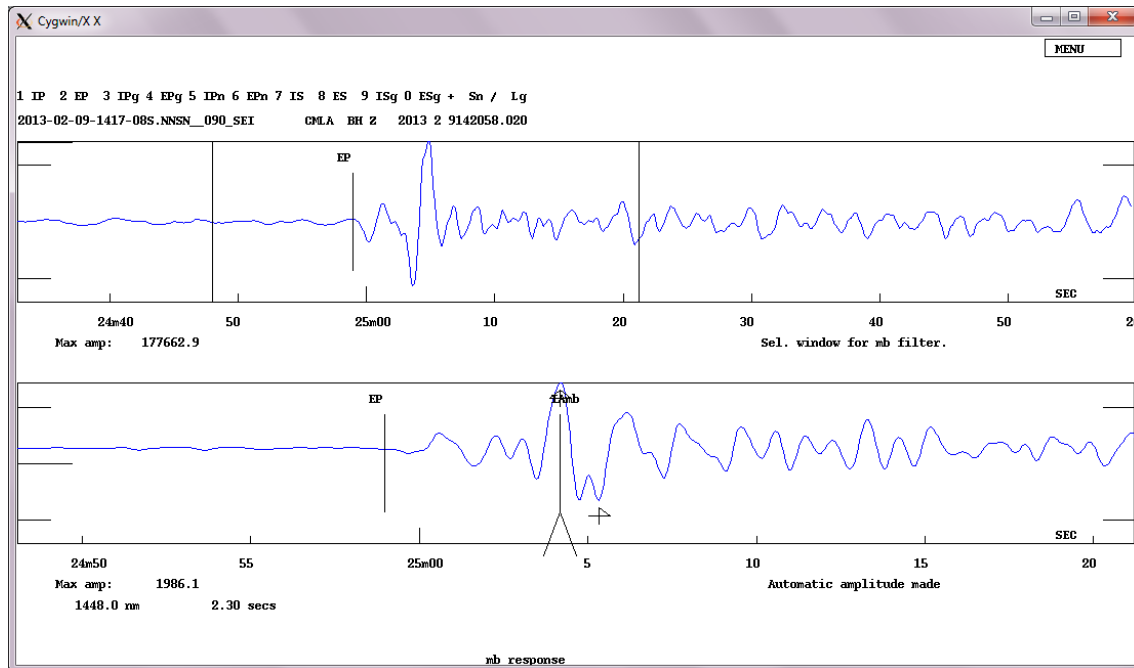
2013  2 9 1416  4.2 D  -3.210 -73.792146.6 TES  1 0.4
```

*Output from the location program:* We can obtain a single-station three-component location, which is within 500 km from the one obtained by the standard location program with several stations. While our distance estimate is very good, we would have to get a 5° larger back-azimuth reading to get a location very close to the multi-station location. This shows the approximate range of uncertainty of single station locations in the far teleseismic range. Wrong phase interpretation may increase single station location errors. E.g., when interpreting in this case the earlier arriving SKS as S, the distance estimate from S-P would be 0.2° shorter in the case of BFO but this underestimation would increase for any station further away within the distance range where SKS arrives before S (see differential travel-time curve in Figure 4 of EX 11.2). For shallow earthquakes at  $\Delta = 100^\circ$  the error  $\delta\Delta$  would reach already 0.9°. However, with proper S-P times, or when several other phases have also been correctly identified, the distance can generally be estimated, up to 160°, within 1-2° and the azimuth with P-wave first motion amplitudes, when measured with good signal-to-noise ratio on well calibrated 3-component seismometer components, within about 5-10°. This allows rather fast and reasonably good first location estimates even with single station data only (see Fig. 11.23 in Chapter 11 of NMSOP-1 in Bormann, 2009, and examples in EX 11.2).

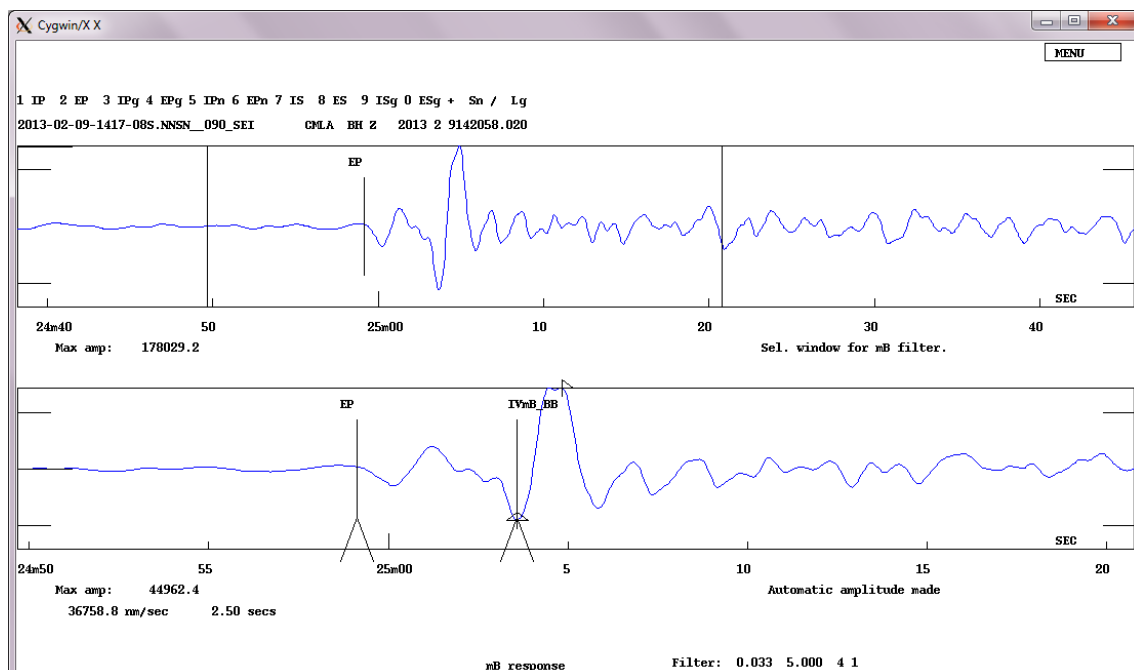
Finally, we determine the teleseismic magnitudes for this earthquake. Since this is a deep earthquake, only the two teleseismic standard body-wave magnitudes  $m_b$  and  $mB\_BB$  should be calculated. Note, however, that the IASPEI WG on magnitudes encourages stations to measure according to the standard procedures also IAMs\_20 and IVMs\_BB with their periods if surface-waves are still clearly visible in records of deeper earthquakes as this is the case also for our example (see, e.g., in section 5, the record trace for station KADAK in the plot: *Vertical-component traces for which phases are picked, filtered 0.02-5 Hz*). Such data are of great interest for investigating the depth and period dependence of surface waves and to derive improved calibration functions  $\sigma(\Delta, h, T)$ . Note that for 20 s surface-waves and thus  $M_s$ \_20 calculation such depth corrections have already been published by Herak et al. (2001):

$$\begin{aligned}
 \Delta M_s(h) &= 0 & \text{for } h < 20 \text{ km} \\
 \Delta M_s(h) &= 0.314 \log(h) - 0.409 & \text{for } 20 \text{ km} \leq h < 60 \text{ km} \\
 \Delta M_s(h) &= 1.351 \log(h) - 2.253 & \text{for } 60 \text{ km} \leq h < 100 \text{ km} \\
 \Delta M_s(h) &= 0.400 \log(h) - 0.350 & \text{for } 100 \text{ km} \leq h < 600 \text{ km.}
 \end{aligned}$$

For station KDAK, e.g., one would measure IAMs\_20 = 10078 nm at  $T = 19$  s and thus get a depth corrected  $M_s$ \_20 = 6.2 + 0.5 = 6.7 which compares rather well with our body-wave magnitudes  $m_b = 6.8$  and  $mB\_BB = 7.0$  determined in the following.



*Amplitude reading for mb magnitude at station CMLA at  $\Delta = 59.9^\circ$ :* As in the case of the regional event, the upper broadband trace is converted into a WWSSN short period trace (lower trace) on which the amplitude Iamb is read, here as 1148 nm with a period of 2.3 s.



*Amplitude reading for mB\_BB magnitude at station CMLA at  $\Delta = 59.9^\circ$ :* As in the case of the regional event the amplitude IVmB\_BB is read on the bottom trace that has been deconvolved into a broadband velocity trace (top trace shows uncorrected data, which are in the case of a broadband seismometer also proportional to velocity): here 36758 nm/s with a period of 2.5 s.

Computing the average magnitudes from all station amplitude observations we get the values below. In this case mB\_BB is equal to Mw, and mb is only slightly smaller. Since the period at which mb is read is almost the same as for mB\_BB the difference is mainly due to the larger bandwidth of the broadband record.

| Magnitude scale | Value |
|-----------------|-------|
| Mw (USGS)       | 7.0   |
| mb              | 6.8   |
| mB_BB           | 7.0   |

The following station amplitude readings were used in the event magnitude computation:

|      |    |       |         |      |         |    |     |      |     |
|------|----|-------|---------|------|---------|----|-----|------|-----|
| EFI  | BZ | dist: | 6121.0  | amp: | 13012.5 | T: | 2.7 | mB = | 6.9 |
| EFI  | BZ | dist: | 6121.0  | amp: | 788.3   | T: | 1.1 | mb = | 6.5 |
| CMLA | BZ | dist: | 6687.0  | amp: | 36756.6 | T: | 2.5 | mb = | 7.9 |
| CMLA | BZ | dist: | 6687.0  | amp: | 36758.8 | T: | 2.5 | mB = | 7.5 |
| CMLA | BZ | dist: | 6687.0  | amp: | 16414.7 | T: | 2.3 | mb = | 7.6 |
| KEV  | BZ | dist: | 10472.0 | amp: | 83.5    | T: | 0.8 | mb = | 6.0 |
| KEV  | BZ | dist: | 10472.0 | amp: | 2421.2  | T: | 5.5 | mB = | 6.6 |
| KBS  | BZ | dist: | 9899.0  | amp: | 500.9   | T: | 0.9 | mb = | 6.5 |
| KBS  | BZ | dist: | 9899.0  | amp: | 11474.4 | T: | 7.2 | mB = | 7.1 |
| BFO  | BZ | dist: | 9626.0  | amp: | 34897.1 | T: | 3.7 | mB = | 7.4 |
| BFO  | BZ | dist: | 9626.0  | amp: | 1747.1  | T: | 1.1 | mb = | 6.9 |
| BILL | BZ | dist: | 10982.0 | amp: | 1555.7  | T: | 6.0 | mB = | 6.7 |
| BILL | BZ | dist: | 10982.0 | amp: | 106.9   | T: | 1.2 | mb = | 6.3 |

## 7 Concluding remarks

The examples given here are meant to illustrate how the data processing is done using the SEISAN software and to illustrate some basic concepts of earthquake data processing. The examples should help newcomers to earthquake data processing to get started. It will be necessary to first learn SEISAN, but then the data used can be downloaded and the steps shown here followed. When analyzing seismograms, the number of phases that can be read depends on the earthquake size and thus the SNR of the recording, the epicentral distance, the source depth as well as on the bandwidth and period range of the seismic records (see Figs. 2.57-2.61 in Chapter 2). From a scientific perspective we wish to have as many good phases read as possible. However, there is often a limitation of how much can be done due to time constraints. We encourage the reader to read more than the first arriving P, and at least to identify clear additional later phases. A good exercise is to download data from the IRIS Wilber, take the data into SEISAN and start processing.

## Acknowledgments

Our thanks go to Jens Havskov, Won-Young Kim and James Dewey. Their review comments helped to improve this manuscript.

## References

- Alsaker, A., Kvamme, L. B., Hansen, R. A., Dahle, A., and Bungum, H. (1991). The  $M_L$  scale in Norway. *Bull. Seism. Soc. Am.*, **81**, 2, 379-389.
- Archangelskaya, V. M. (1959). The dispersion of surface waves in the earth's crust. *Izv. Akad. Nauk SSSR, Seriya Geofiz.*, **9** (in Russian).
- Bormann, P. (ed.) (2009). *New Manual of Seismological Observatory Practice*. Vol. 1 and Vol. 2, IASPEI, GFZ German Research Centre for Geosciences, Potsdam; electronic edition, accessible via <http://nmsop.gfz-potsdam.de>; DOI: 10.2312/GFZ.NMSOP\_r1
- Bormann, P. (2011). Earthquake magnitude. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 207-218; doi: 10.1007/978-90-481-8702-7.
- Bormann, P., and D. Di Giacomo (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Braunmiller, J., Deichmann, N., Giardini, D., Wiemer, St., and the SED Magnitude Working Group (2005). Homogeneous moment-magnitude calibration in Switzerland. *Bull. Seism. Soc. Am.*, **95**, 1, 58-74.
- Choy, G. L. (2012). IS 3.5: Stress conditions inferable from modern magnitudes: development of a model of fault maturity. In: Bormann, P. (Ed.) (2012). *New Manual of Seismological Observatory Practice (NMSOP-2)*, IASPEI, GFZ German Research Centre for Geosciences, Potsdam, 10 pp.; doi: 10.2312/GFZ.NMSOP-2\_IS\_35.
- Eshelby, J. (1957). The determination of the elastic field of an ellipsoidal inclusion and related problems. *Proc. R. Soc. London A*, 241, 376–396.
- Gorbunova, I. V., and Kondorskaya, N. V. (1977). Magnitudes in the seismological practice of the USSR. *Izv. Akad. Nauk SSSR, ser Fizika Zemli*, No. 2, Moscow (in Russian).
- Havskov, J. and Ottemöller, L. (2010). *Routine Data Processing in Earthquake Seismology, With Sample Data, Exercises and Software*. Springer, 380p.
- Herak, M., Panza, G., and Costa, G. (2001). Theoretical and observed depth corrections for  $M_s$ . *Pure Appl. Geophys.*, **158**, 1517-1530.
- Herak, M., Panza, G., and Costa, G. (2001). Theoretical and observed depth corrections for  $M_s$ . *Pure Appl. Geophys.*, **158**, 1517-1530.
- Hutton, L. K. and Boore, D. (1987). The  $M_L$  scale in Southern California. *Bull. Seismol. Soc. Am.*, 77:2074–2094.
- IASPEI (2013). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG_recommendations_20130327.pdf)
- Lienert, B. R. E., Berg, E., and Frazer, L. N. (1986). Hypocenter: An earthquake location method using centered, scaled, and adaptively damped least squares. *Bull. Seismol. Soc. Am.*, 76:771–783.
- Lienert, B. R. E. and Havskov, J. (1995). A computer program for locating earthquakes both locally and globally. *Seism. Res. Lett.*, 66:26–36.
- Richter, C. F. (1935). *Elementary seismology*, W. H. Freeman and Co., San Francisco, 578p.

Roberts, R. G., Christoffersson, A., and Cassidy, F. (1989). Real time events detection, phase identification and source location estimation using single station component seismic data and a small PC. *Geophys. J. Int.*, 97:471–480.

Schöffel, H.-J. and Das, S. (1999). Fine details of the Wadati-Benioff zone under Indonesia and its geodynamic implications. *J. Geophys. Res.*, **104**, 13101-13114.

Uhrhammer, R. A., and Collins, E. R. (1990). Synthesis of Wood-Anderson seismograms from broadband digital records. *Bull. Seism. Soc. Am.*, **80**, 702-716.

Utsu, T. (2002). Relationships between magnitude scales. In: *International Handbook of Earthquake and Engineering Seismology, Part A*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 733–746.

|                |                                                                                                                                                     |
|----------------|-----------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Topic</b>   | <b>HYPOSAT / HYPOMOD</b>                                                                                                                            |
| <b>Author</b>  | Johannes Schweitzer, NOR SAR, P.O.Box 53, N-2027 Kjeller<br>Fax: +47 63818719, E-mail: <a href="mailto:johannes@norsar.no">johannes@norsar.no</a>   |
| <b>Version</b> | HYPOSAT 4.4 and HYPOMOD 1.1 (as of October 2002); references and link to NMSOP-2 added in November 2011;<br><b>DOI: 10.2312/GFZ.NMSOP-2_PD_11.1</b> |

## User Manual for HYPOSAT (including HYPOMOD)

### 1 Introduction

**HYPOSAT** is a program package developed to locate seismic sources. It utilizes travel-time data, backazimuth (*i.e.*, station-to-event azimuth) values, and ray-parameter values. Phases considered are those included in IASP91-type tables, and reflections from the Conrad and from the Mohorovičić discontinuities, if local models are used. The program follows the phase name recommendations of the IASPEI Working Group on Standard Phase Names (see IS 2.1). Additionally, all possible travel-time differences between different onsets at individual stations are estimated and can be included in the location process (*e.g.*, PcP-P as an additional constraint for the source depth). If amplitude and period measurements for P onsets or surface waves are available, station magnitudes and an event magnitude can also be estimated. More details about the general features of the program can be found in Schweitzer (1997, 2001a).

The data files containing the global models (*e.g.*, **iasp91.tbl** and **iasp91.hed**), the list of up to now defined seismo-tectonic units (**REG\_L3.DAT**), the attenuation curves for magnitude estimations (**MB\_G-R.DAT** and **MB\_V-C.DAT**), and the ellipticity corrections (**elcordir.tbl**) must all be located in the same directory. The path to this directory must be set by the environment variable **HYPOSAT\_DATA**.

The program needs two input files in ASCII format. One file contains the general parameters to steer the inversion process (**hyposat-parameter**) and the other file contains the observed data for the event to be localized (**hyposat-in**). Contents and structure of these files will be explained in the following sections.

The program **HYPOMOD** uses the same input files as **HYPOSAT** but it only calculates all residuals for a given hypocenter without any inversion.

The newest versions of the programs **hyposat\_hypomod** (including source code, this manual, the PDF version of Schweitzer (1997), data files containing travel-time models and station parameters, and several examples) are located in six compressed tar-files (all versions in **hyposat.version.tar.Z** or **hyposat.version.tar.gz**, the UNIX version in **hyposat\_u.version.tar.Z** or **hyposat\_u.version.tar.gz**, and the LINUX version in **hyposat\_l.version.tar.Z** or **hyposat\_l.version.tar.gz**) for free download either directly from the program list in the NMSOP-2 cover page folder *Download Programs & Files* or from NOR SAR's anonymous ftp-address <ftp://ftp.norsar.no> under the directory `/pub/outgoing/johannes/` *hyposat*. The address when using your web-browser is: <http://ftp.norsar.no/pub/outgoing/johannes/hyposat>.

Questions related to program updates and maintenance should be directed to the author.

## 2 Getting started

This section describes how some simple examples for **HYPOSAT** can be started and executed. The simplest way to use the program for own locations is to start from one of the following examples and modify input data and parameters for your needs. The meaning and format of the input is described in the following sections.

### Installation of **HYPOSAT**:

1) Make a sub-directory for **HYPOSAT**, copy the compressed tar-file containing the *hyposat-software package* from the NORSAR's ftp site (*ftp.norsar.no*), decompress it, and run:

```
tar -xvf hyposat.version.tar or
tar -xvf hyposat_u.version.tar or
tar -xvf hyposat_l.version.tar,
depending on the module you have downloaded.
```

Then you will have a directory containing the following files and subdirectories in the UNIX Solaris) case:

```
bin/    data/    examples/ man/  README_u  src/
```

or in the LINUX:

```
bin_l/  data_l/  examples_l/ man/  README_l  src_l/
```

or all together if you had downloaded the full version.

The file *README\_u* (or *README\_l*) contains a complete list of all files following with the installed *hyposat-software package* and a explanation of these files.

2) If needed re-compile the software in the */src* (or */src\_l*) subdirectory by running:  
*make* and/or *make -f Makefile.hypomod*

### Executing **HYPOSAT**:

Change to the subdirectory *examples/* (or *examples\_l*).

Here you will find input file examples for four different cases: an event observed with a network of stations (*.net*), a single array (*.single\_array*), a set of local and regional stations (*.regional*), and a teleseismically observed event (*.tele*). **HYPOSAT** runs with two input files. To check your installation, try the following:

```
cp hyposat-in.net hyposat-in
cp hyposat-parameter.net hyposat-parameter
setenv HYPOSAT_DATA $path/hyposat/data
(or setenv HYPOSAT_DATA $path/hyposat/data_l)
(where $path is the actual path to the subdirectory hyposat)
and run:
```

```
../bin/hyposat (or ../bin_l/hyposat)
```

You will then get an output file *hyposat-out*, which should be identical to the file *hyposat-out.net* distributed with the *hyposat-software package*.



### 3 The File *hyposat-parameter*

The file containing the inversion-steering parameters must (!) have the name *hyposat-parameter* and must reside in the actual directory where the program is executed. The structure and contents of *hyposat-parameter* is as follows:

-----start of the example for a *hyposat-parameter* file -----

hyposat-parameter file for hyposat 4.4

```

GLOBAL MODEL                : ak135
GLOBAL MODEL 2              : iasp91
GLOBAL MODEL 3              : _
GLOBAL MODEL 4              : _

LOCAL OR REGIONAL MODEL     : _
PHASE INDEX FOR LOCAL MODEL : 0000

CRUST 5.1 PATH              : ./
CRUST 5.1                   : 0
OUTPUT OF REGIONAL MODEL (DEF 0) : 1

STATION FILE                : ../data/stations.dat
P-VELOCITY TO CORRECT ELEVATION : 4.5
S-VELOCITY TO CORRECT ELEVATION : 3.3
STATION CORRECTION FILE     : stations.cor

LG GROUP-VELOCITY (DEF 3.5 [km/s]) : 3.5752
RG GROUP-VELOCITY (DEF 2.5 [km/s]) : 2.5

LQ GROUP-VELOCITY (DEF 4.4 [km/s]) : 4.4
LR GROUP-VELOCITY (DEF 3.95 [km/s]) : 2.85

STARTING SOURCE LATITUDE [deg] : 999.
STARTING LATITUDE ERROR [deg]  : 10.

STARTING SOURCE LONGITUDE [deg] : 999.
STARTING LONGITUDE ERROR [deg]  : 10.

STARTING SOURCE DEPTH [km]      : 0.
STARTING DEPTH ERROR [km]       : 50.
DEPTH FLAG (f,b,d,F,B,D)       : b

STARTING SOURCE TIME (epochal time): 0.
STARTING TIME ERROR [s]         : 600.0

MAXIMUM # OF ITERATIONS        : 80
# TO SEARCH OSCILLATION (DEF 4) : 6

LOCATION ACCURACY [km] (DEFAULT 1.): 1.
CONSTRAIN SOLUTION (0/1)       : 1

CONFIDENCE LEVEL (68.3 - 99.99%) : 0.
EPICENTER ERROR ELLIPSE (DEF 1) : 1

SLOWNESS [S/DEG] (0 = APP. VEL) : 1

```

```

MAXIMUM AZIMUTH ERROR [deg]           : 30.
MAXIMUM SLOWNESS ERROR [s/deg]       : 5.

FLAG USING TRAVEL-TIME DIFFERENCES : 1

MAGNITUDE CALCULATION (DEF 0)        : 1
P-ATTENUATION MODEL (G-R or V-C)     : G-R
S-ATTENUATION MODEL (IASPEI or R-P)  : R-P

INPUT FILE NAME (DEF hyposat-in)     : _

OUTPUT SWITCH (YES = 1, DEFAULT)     : 1
OUTPUT FILE NAME (DEF hyposat-out)   : _
OUTPUT LEVEL                        : 4

```

-----end of the example-----

The order in which these parameters are set is arbitrary. The parameters must be identified with the above given description (bold-faced). The parameters must be written in the file in capital letters! The settings itself must follow after the 37th character of the line (*i.e.*, in this example two characters after the colon). Whenever a line does not comply with this rule, it will be ignored (*e.g.*, blank lines or lines starting with a '\*', ...). This file is read only once at the beginning of a location run. Each line can be repeated several times within the file with another setting. In this case, the last set value is used for the location process. For file names, the full path name can be given. In the following all the parameters are explained in more detail:

**GLOBAL MODEL:** Type of the reference model used to calculate all travel time related theoretical data. This package contains the following models:

|        |                                                           |
|--------|-----------------------------------------------------------|
| ak135  | AK135 model (Kennett <i>et al.</i> , 1995)                |
| iasp91 | IASP91 model (Kennett, 1991; Kennett & Engdahl, 1991)     |
| jb     | Jeffreys-Bullen model (Jeffreys & Bullen, 1940 and later) |
| prem   | PREM model (Dziewonski & Anderson, 1981)                  |
| sp6    | SP6 model (Morelli & Dziewonski, 1993)                    |

The directory where these travel-time tables reside must be specified with the environment variable **HYPOSAT\_DATA** before the program is started. The travel-time tables are based on the **libtau**-software package written by Ray Buland and distributed as IASP91-software. If you use an own version of the **libtau**-software, you will have to exchange the **libtau\_h.f** file (see **Makefile** in the source-code directory) and to exchange the corresponding data files (**\*.hed** and **\*.tbl**) because for the version included here some parameter and dimension settings were changed in the include file **ttlim.h**.

**GLOBAL MODEL 2:** Here one can give the name of any other second global model to be used for specific ray paths indicated in the data input file.

**GLOBAL MODEL 3:** Here one can give the name of any other third global model to be used for specific ray paths indicated in the data input file.

**GLOBAL MODEL 4:** Here one can give the name of any other fourth global model to be used for specific ray paths indicated in the data input file.

**LOCAL OR REGIONAL MODEL:** Name of the file with a local (or regional) velocity model. Travel times will be estimated for the following seismic phases (as far as they can be observed with respect to distance and source depth): Pg, Pb, Pn, P, pPg, pPb, pPn, pP,

PbP (i.e., in this program upper side reflection from the 'Conrad'), PmP, PgPg, PbPb, PnPn, PP, and the converted phases sPg, sPb, sPn, sP, SbP and SmP. The same phase set is used for S-type phases, respectively.

This parameter (file name) must only be set if a special local or regional model is to be used instead of the global one. The velocity model must contain the following information:

In the first line maxdis = maximum distance in [deg] for which this model shall be used. It is followed by the depth in [km], the P-phase velocity  $V_p$  in [km/s], and the S-phase velocity  $V_s$  in [km/s]. The model may contain layers with a constant velocity or with velocity gradients. First order discontinuities must be specified with two lines for the same depth. Additionally, the Conrad- and the Mohorovičić-discontinuities should be marked as shown in the following example. Otherwise, all calculated phases would be called Pg (or Sg).

```
-----example for a file containing a regional velocity-----
10.      | maxdis in free format
      0.000      5.400      3.100      | depth, vp, vs in format (3F10.3)
      10.000     5.800      3.200
      20.000     5.800      3.200CONR      | + mark for the 'Conrad'
      20.000     6.500      3.600      | in format (3F10.3,A4)
      30.000     6.800      3.900MOHO      | + mark for the 'Moho'
      30.000     8.100      4.500
      77.500     8.050      4.400
      120.000    8.100      4.500
-----end of example-----
```

**PHASE INDEX FOR LOCAL MODEL:** This parameter allows the user to specify the set of seismic phases, for which travel times and their partial derivatives will be calculated in the local/regional model:

The parameter is a 4 digit number. The position of a digit defines the phase type for which the value of the digit defines the action for this phase:

dxxx the digit (d) at this place is the flag for surface reflections (e.g., pP or sS)

xidx the digit (d) at this place is the flag for surface multiples (e.g., PP or SS)

xxdx the digit (d) at this place is the flag for reflections at the Conrad- or the Mohorovičić-discontinuity (e.g., PbP or SmS). Note that the here used name 'PbP', to indicate reflection from the Conrad discontinuity, is not a regular phase name as recommended by the IASPEI Working Group on Standard Phase Names (see IS 2.1).

xxxd the digit (d) at this place is the flag for converted phases (e.g., sP or PmS)

d itself can have the following values:

d = 1 only P-type onsets will be calculated

d = 2 only S-type onsets will be calculated

d = 3 both phase types (P and S) will be calculated

e.g.,: 1320 means: the phases pP, PP, SS, SbS and SmS will be calculated but no conversions. 000 or simply 0 means: none of these phases will be calculated.

The direct phases Pg, Pb, Pn, P (or the same for S) will always be calculated as long as the **PHASE INDEX FOR LOCAL MODEL** is not set to a negative value.

**CRUST 5.1 PATH:** The path to the directory where the CRUST 5.1 data files (Mooney *et al.*, 1998) reside.

**CRUST 5.1:** This parameter controls the usage of the model CRUST 5.1:

= 0 CRUST 5.1 is not used at all.

= 1 CRUST 5.1 is used to calculate station corrections with respect to the local crustal structure below the station.

- = 2 CRUST 5.1 is used to define a local/regional velocity model between the source and stations up to a distance of 6 deg.
- = 3 CRUST 5.1 is used to define a local/regional velocity and to correct for local crustal structures at the stations and at reflection points at the Earth's surface. If this parameter is set to any value larger than 0 and the model CRUST 5.1 is available, a time correction for the crustal structure at the reflection point of phases reflected at the Earth's surface will be calculated (*e.g.*, PnPn, sS, P'P', ...)

**OUTPUT OF REGIONAL MODEL:** This flag defines if the local/regional model used for the final inversion is to be printed out in the output file (**hyposat-out**). This option is particularly interesting whenever this velocity model was interpolated from CRUST 5.1:

- 0 no model output (default)
- 1 model output

**STATION FILE:** Name of the file with station coordinates either in NEIC or in CSS 3.0 file format. Only these two formats are currently supported! To get the location results faster, the usage of a file containing only your usually used stations is recommended.

**STATION CORRECTION FILE:** Name of the file for station corrections. This file must contain the station name and then the local velocities for P and S waves below this station to calculate the best elevation correction for this station. This value can also be used to correct for a known velocity anomaly below this station. The input is format free. If such information is not available, leave it blank. If one station is not in this list, the default values as defined by the input parameters **P-VELOCITY TO CORRECT ELEVATION** and **S-VELOCITY TO CORRECT ELEVATION** are used!

-----example for a file containing station corrections -----  
 GEC2 5.2 3.2 | in free format  
 -----end of example-----

**P-VELOCITY TO CORRECT ELEVATION:** Local P velocity ( $V_{pl}$ ) to correct for the station elevation (default 5.8 km/s) if this parameter is not set in the **STATION CORRECTION FILE**. If  $V_{pl} = 99$ , a station-elevation correction is not applied and the **STATION CORRECTION FILE** is not used.

**S-VELOCITY TO CORRECT ELEVATION:** Local S velocity ( $V_{sl}$ ) to correct for the station elevation (default  $V_{pl}/\sqrt{3}$ ), if not given in **STATION CORRECTION FILE**.

**LG GROUP-VELOCITY:** A group velocity for Lg can be defined; the default value is 3.5 [km/s].

**RG GROUP-VELOCITY:** A group velocity for Rg can be defined; the default value is 2.5 [km/s].

**LQ GROUP-VELOCITY:** A global group velocity for Love wave. (LQ) can be defined; the default value is 4.4 [km/s].

**LR GROUP-VELOCITY:** A global group velocity for Rayleigh waves (LR) can be defined; the default value is 3.95 [km/s].

**STARTING SOURCE LATITUDE:** Initial value for event latitude (no default value, an initial latitude will be estimated or chosen from the input data). Valid range: -90 deg ≤ value ≤ 90 deg. An initial solution must be set for both latitude and longitude!

**STARTING LATITUDE ERROR:** Its standard deviation (default 10 deg).

**STARTING SOURCE LONGITUDE:** Initial value for event longitude (no default value, a start longitude will be estimated or chosen from the given data). Valid range: -180 deg ≤ value ≤ 180 deg. An initial solution must be set for both latitude and longitude!

**STARTING LONGITUDE ERROR:** Its standard deviation (default 10 deg).

**STARTING SOURCE DEPTH:** Starting value for the event depth (default 0. km).

**STARTING DEPTH ERROR:** Its standard deviation (default 50 km).

**DEPTH FLAG:** Flag to handle the source depth:

f or F the hypocenters depth is fixed for this inversion, as defined by **STARTING SOURCE DEPTH**.

d or D the depth will be inverted from the beginning.

b or B means both: *i.e.*, the inversion begins with the fixed depth from **STARTING SOURCE DEPTH** and after reaching a stable solution, the routine also tries to invert for the source depth. Both solutions (fixed and free depth) will be listed in *hyposat-out* (see example).

**STARTING SOURCE TIME:** Initial value for source time in epochal time format. The initial source time can be given in three different formats: as epochal time (*i.e.*, the number of seconds after 01 January 1970 00:00:00), and in two human readable formats yyyy-doy:hh.mm.ss.sss (DOY = day-of-year) and yyyy-mo-dd:hh.mm.ss.sss. For example, the 1 October 2002 at 3 o'clock in the afternoon can be written as 1033484400.0 (epochal time) or as 2002-274:15.00.00.000 or as 2002-10-01:15.00.00.000.

If this value is not set, an initial source time will be estimated from travel-time differences between direct S type and direct P type observations by using Wadati's approach. For this, the program calculates a Vp/Vs relation for each phase type and estimate a source time, respectively. The initial source time is then the mean value of all estimated source times. In the case of only one S-P observation, Vp/Vs is set to sqrt(3.). If no S-P time observation is available, the source time is set to the earliest observed onset time.

**STARTING TIME ERROR:** its standard deviation (default 120 s).

**MAXIMUM # OF ITERATIONS:** To avoid indefinite iterations to find a solution, a maximum number of iterations must be defined (default 80).

**# TO SEARCH OSCILLATION:** Here the number of solutions from older iterations can be defined with which the newest solution will be compared to identify iterations between very similar solutions (oscillating solutions). The default value is 4 and the maximum number is 10.

**LOCATION ACCURACY:** If we calculate the distance between the solutions of two consecutive iterations in km, this value gives the maximum vector length to stop the iteration process. The default value is 1 km (also if **LOCATION ACCURACY** is set to 0.).

**CONSTRAIN SOLUTION:** If this flag is set to 1 (default value), all used observations are checked for their residuals and only the data with relatively small residuals are used for a final inversion.

0 no final restriction of data

1 final restriction of data (default).

**CONFIDENCE LEVEL:** Level of confidence for the output of uncertainties in percent, the default uncertainty is +/- one standard deviation (*i.e.*, ca. 68.3%).

**EPICENTER ERROR ELLIPSE:** The setting of this flag defines whether an error ellipse for the final solution will be calculated or not. The size of the error ellipse corresponds to the chosen confidence level.

0 no error ellipses

1 error ellipses will be calculated (default).

**SLOWNESS [S/DEG]:** The slowness of a seismic phase can be given as input value in two different units: apparent velocity or ray parameter. All slowness values must have the same unit in the data input file *hyposat-in*:

0 the slowness input values are apparent velocities in [km/s]

- 1 the slowness input values are ray parameters in [s/deg]
- MAXIMUM AZIMUTH ERROR:** Maximum value of a backazimuth-residual to use this observation as a defining phase in [deg].
- MAXIMUM SLOWNESS ERROR:** Maximum value of a slowness-residual to use this observation as a defining phase in [s/deg].
- FLAG USING TRAVEL-TIME DIFFERENCES:** By default, the program uses the travel-time differences between all phases observed at one station to estimate a hypocenter. This can be switched off:
- 0 travel-time differences are not used
- 1 travel-time differences are used (default)
- MAGNITUDE CALCULATION:** Flag if body wave (mb) or surface wave (Ms) magnitudes are calculated for this event.
- 0 magnitudes are not calculated (default)
- 1 magnitudes are calculated
- P-ATTENUATION MODEL:** With this parameter the attenuation model used for mb calculations can be chosen. The two possibilities are **G-R** for Gutenberg-Richter (Gutenberg and Richter, 1956a, b) or **V-C** for Veith-Clawson (Veith and Clawson, 1972). No default model is defined!
- S-ATTENUATION MODEL:** With this parameter the attenuation model used for Ms calculations can be chosen. The two possibilities are **IASPEI** for the IASPEI 1967 formula (often also called Praha formula) or **R-P** for the Rezapour and Pearce (1998) formula. No default model is defined!
- INPUT FILE NAME:** A file name can be defined at this place if not the standard input-file name (*hyposat-in*) should be used.
- OUTPUT SWITCH:** This flag determines whether any output file (see also **OUTPUT FILE NAME**) will be written:
- 0 no output file
- 1 output file will be written (default)
- OUTPUT FILE NAME:** A file name can be given at this place if not the standard output-file name (*hyposat-out*) should be used.
- OUTPUT LEVEL:** Verbosity level for output during the location process on screen (0 – 10) or on file (>10) during the inversion; the default value is 4. If **OUTPUT LEVEL** > 10, the output level for the screen is internally calculated. In addition, the resolution, covariance, correlation, and the information-density matrix will then be written out in a file called *hyposat-gmi.out*. This file contains always the named matrices for the last inversion. **OUTPUT LEVEL** can be set to the following values:

| Input   | Matrix Output                                                                     | Screen Output Level |
|---------|-----------------------------------------------------------------------------------|---------------------|
| 0 – 10  | None                                                                              | 0 – 10              |
| 11      | the resolution matrix will be written out.                                        | 0                   |
| 12      | the covariance matrix will be written out.                                        | 1                   |
| 13      | the correlation matrix will be written out.                                       | 3                   |
| 14      | all three matrices will be written out.                                           | 5                   |
| 15      | "                                                                                 | 7                   |
| 16      | "                                                                                 | 9                   |
| 17 – 19 | "                                                                                 | 10                  |
| 20      | ", plus the diagonal elements of information -density matrix will be written out. | 0                   |
| 21 – 29 | "                                                                                 | as for 11 – 19      |
| 30      | ", plus the whole information-density matrix                                      | 0                   |

|         |                      |                |
|---------|----------------------|----------------|
|         | will be written out. |                |
| 31 – 39 | "                    | as for 11 – 19 |

## 4 The File *hyposat-in*

**HYPOSAT** then needs a file with all the readings for a specific event: This file has by default the name *hyposat-in* but the name can be defined in the parameter file *hyposat-parameter*. The data-input file must have the following format:

1. line: any title for event identification of maximum 80 characters (used also in the output-file *hyposat-out*).
2. – (n+1)'th line with the n-observed onsets. This line must (!) be compatible with the following format (FORTRAN):

If we don't wish to calculate magnitudes (see **MAGNITUDE CALCULATION** in the *hyposat-parameter* file) this line can contain the following data with the following format:

```
(a5,1x,a8,1x,i4,4(1x,i2),1x,f6.3,1x,f5.3,1x,f6.2,3(1x,f5.2),1x,a6)
```

station name, phase name, year, month, day, hour, minute, second, standard deviation of the onset time, backazimuth, standard deviation of the backazimuth observation, either ray parameter [s/deg] or apparent velocity [km/s] (see **SLOWNESS [S/DEG]** in the *hyposat-parameter* file), standard deviation of the slowness observation in [s/deg] or [km/s], and a six character long combination of controlling flags. These steering flags (**123456**) have the following meanings and options:

- Position **1** the time reading of this onset can be used ('T' or 't') or not used ('\_') for the inversion.
- Position **2** the backazimuth reading of this onset can be used ('A' or 'a') or not used ('\_') for the inversion.
- Position **3** the slowness reading of this onset can be used ('S' or 's') or not used ('\_') for the inversion.
- Position **4** the time reading of this onset can be used ('D' or 'd') or not used ('\_') to calculate travel-time differences and use them in the inversion.
- Position **5** the onset-time reading of this onset will be corrected ('R' or 'r') or not corrected ('\_') for the crustal structure below a reflection point at the Earth's surface by calculating the travel-time difference for the crustal path between the used global model (as set with **GLOBAL MODEL** in *hyposat-parameter*) and CRUST 5.1.
- Position **6** if set to '2', '3', or '4', the other global Earth model (as set with **GLOBAL MODEL 2**, **GLOBAL MODEL 3**, or **GLOBAL MODEL 4** in *hyposat-parameter*) will be used to calculate the theoretical onset time, the ray parameter, and their partial derivatives for this onset. With any other character at this place the standard global Earth model (see **GLOBAL MODEL** in *hyposat-parameter*) will be used.

Keeping all positions 1 – 6 blank, the flag combination **TASDR\_** will be used as default value.

If one also wishes to calculate magnitudes (see **MAGNITUDE CALCULATION** in *hyposat-parameter*) this line can contain the following data in the following format:

(a5,1x,a8,1x,i4,4(1x,i2),1x,f6.3,1x,f5.3,1x,f6.2,3(1x,f5.2),1x,a6,1x,f6.3,1x,f12.2)

station name, phase name, year, month, day, hour, minute, second, standard deviation of the onset time, backazimuth, standard deviation of the backazimuth observation, either ray parameter [s/deg] or apparent velocity [km/s] (see **SLOWNESS [S/DEG]** in the *hyposat-parameter* file), standard deviation of the slowness observation in [s/deg] or [km/s], the six character long combination of controlling flags (see above), the period of the observed onset, and finally the amplitude of the signal in [nm].

S-type onsets must always be listed after the corresponding P-type onsets – if not, the travel-time difference between these two onsets (S-P) cannot be used for calculating a starting solution for source time and distance from the corresponding station. If it is unknown, of which type the P or S onsets are, you can choose the names P1 or S1 to tell the program that you know it is the first P or the first S onset at this station. Then the program itself chooses the right phase name depending on the distance of the observation.

Onsets with a station name starting with a \* and lines starting with a blank character are not used.

The values for backazimuth, slowness, period, and amplitude are optional. If backazimuth or slowness values are not available, they must be set to -999. or -1.; the amplitude/period information is only used if both values are larger than 0.

For each phase name not defined by the applied travel-time model(s), the program searches for the best fitting phase. However, onset time and ray parameter of such a phase are not used in the inversion, but eventually the backazimuth information!

When using the correct format, an input file can look like the following example:

```
----- example for a hyposat-in file -----
NORTHERN MOLUCCA SEA, 1996 29 June, from pIDC's Reviewed Event Bulletin (REB)
WRA P 1996 06 29 00 41 44.700 0.300 331.50 20.00 11.10 1.00 T ASD 0.300 4.50
WRA S 1996 06 29 00 45 48.7 0.600 338.0 20.0 17.0 2.00 T ASD 0.900 2.00
QIS P1 1996 06 29 00 42 12.8 0.300 -999. 0.0 -999. 0.00 T ASD
QIS PCP 1996 06 29 00 45 44.8 0.300 -999. 0.0 -999. 0.00 T ASD
ASAR P 1996 06 29 00 42 16.9 0.300 346.3 20.0 7.1 1.00 T ASD 0.500 3.40
ASAR PCP 1996 06 29 00 45 45.2 0.300 345.1 20.0 2.3 1.00 T ASD 0.500 2.20
ASAR S 1996 06 29 00 46 45.3 0.600 347.6 20.0 20.3 2.00 T ASD 0.800 3.90
WARB P 1996 06 29 00 42 31.2 0.300 339.4 30.0 8.2 2.00 T ASD 0.700 6.50
WARB PCP 1996 06 29 00 45 49.4 0.300 -999. 0.0 -999. 0.00 T ASD
MEEK P 1996 06 29 00 42 42.7 0.300 -999. 0.0 -999. 0.00 T ASD
CMAR P 1996 06 29 00 43 09.0 0.300 109.7 30.0 7.8 2.00 T ASD 0.400 0.60
CMAR LR 1996 06 29 00 57 48.7 50.000 110.0 30.0 39.5 2.00 A 19.360 188.80
FORT P 1996 06 29 00 43 11.3 0.300 -999. 0.0 -999. 0.00 T ASD
WOOL P 1996 06 29 00 43 15.6 0.300 7.9 30.0 9.9 2.00 T ASD 0.600 4.10
SHK P 1996 06 29 00 43 25.4 0.300 -999. 0.0 -999. 0.00 T ASD
STKA P 1996 06 29 00 43 46.8 0.300 323.2 20.0 9.0 1.00 T ASD 0.600 7.20
STKA PCP 1996 06 29 00 46 12.0 0.300 -999. 0.0 -999. 0.00 T ASD
KSAR P 1996 06 29 00 43 46.7 0.300 177.3 20.0 10.1 2.00 T ASD 0.700 1.70
KSAR LR 1996 06 29 01 04 25.0 50.000 160.0 20.0 46.0 2.00 A 19.860 40.10
MJAR P 1996 06 29 00 43 51.9 0.300 357.1 20.0 18.8 1.00 T ASD 0.550 2.00
PDY P 1996 06 29 00 46 47.0 0.300 111.2 30.0 6.6 2.00 T ASD 0.450 4.30
ZAL P 1996 06 29 00 47 09.0 0.300 -999. 0.0 -999. 0.00 T ASD
ABKT P 1996 06 29 00 48 09.5 0.300 -999. 0.0 -999. 0.00 T ASD
*
*NRI P 1996 06 29 00 48 08.7 0.300 195.9 30.0 3.9 2.00 T ASD 0.750 3.00
* Note, the following phase has an irregular phase name! Therefore we will not use its onset time
* or slowness for inversion.
*
NRI X 1996 06 29 00 48 08.7 0.300 195.9 30.0 3.9 2.00 A 0.750 3.00
MAW P 1996 06 29 00 49 01.1 0.300 -999. 0.0 -999. 0.00 T ASD
KVAR P 1996 06 29 00 49 17.2 0.300 -999. 0.0 -999. 0.00 T ASD
NPO P 1996 06 29 00 49 31.0 0.300 -999. 0.0 -999. 0.00 T ASD
ARCES P 1996 06 29 00 49 51.7 0.300 94.5 20.0 4.1 1.00 T ASD 0.550 0.80
SPITS P 1996 06 29 00 49 55.9 0.300 116.6 30.0 3.5 2.00 T ASD 0.900 3.60
FINES P 1996 06 29 00 50 00.0 0.300 111.2 20.0 5.9 1.00 T ASD 0.550 0.60
HFS P 1996 06 29 00 50 28.7 0.300 311.3 30.0 1.5 2.00 T ASD 0.750 1.10
TXAR PKPdf 1996 06 29 00 55 42.7 0.300 -999. 0.0 -999. 0.00 T ASD
SCHQ PKPdf 1996 06 29 00 55 41.2 0.300 -999. 0.0 -999. 0.00 T ASD
DBIC PKPdf 1996 06 29 00 55 57.1 0.300 -999. 0.0 -999. 0.00 T ASD
PLCA PKPdf 1996 06 29 00 56 09.1 0.300 311.9 30.0 6.4 2.00 T ASD 0.900 0.90
LPAZ PKPdf 1996 06 29 00 56 44.7 0.300 -999. 0.0 -999. 0.00 T ASD
```



----- end of the example -----

## 5 The File hyposat-out

With the above example for a *hyposat-in* file you will get the following output-file *hyposat-out*: This example was calculated on a UNIX system, the results on a LINUX system might be slightly different. Explanations are included in [ .... ]:

-----example for a *hyposat-out* file -----

HYPOSAT Version 4.4

NORTHERN MOLUCCA SEA, 1996 29 June, from pIDC's Reviewed Event Bulletin (REB)

Parameters of starting solution ( 1 standard deviation):

[ Not all backazimuth-observation pairs are used: if one station is more than 170 deg apart from the crossing point, this crossing point is skipped. ]  
 Mean epicenter calculated from 165 backazimuth observation pairs  
 Mean epicenter lat: 29.4748 42.0082 [deg]  
 Mean epicenter lon: 121.5921 6.0977 [deg]

[ type 1: S - P or S1 - P1 observation, type 2: Sg - Pg observation, type 3: Sn - Pn observation, type 4: Sb - Pb observation ]  
 S-P Travel-time difference type 1 with 2 observation(s)  
 Mean source time: 836008582.700 11.636 [s]  
 Mean vp/vs: 1.758 0.035

Iterations : 7  
 Number of defining: 58  
 First reference model : ak135  
 Second reference model : iasp91

The new source parameters

Confidence level of given uncertainties: 68.27 %

Source time : 1996 06 29 00 36 42.801 0.150 [s]  
 or 836008602.801 0.150 [s]  
 Epicenter lat: 1.3045 0.0250 [deg]  
 Epicenter lon: 126.3165 0.0611 [deg]  
 Source depth : 0.00 [km] Fixed

Epicenter error ellipse:  
 Major axes: 5.98 [km] Minor axes: 3.61 [km]  
 Azimuth: 62.0 [deg] Area: 67.82 [km\*2]

Flinn-Engdahl Region ( 266 ): Northern Molucca Sea

Magnitudes: 4.39 (mb, G-R) 3.57 (Ms, R-P)

| Stat                                                                                         | Delta   | Azi    | Phase | [used] | Onset time   | Res     | Baz    | Res     | Rayp  | Res   | Used | Amplitude | Period | MAG  |
|----------------------------------------------------------------------------------------------|---------|--------|-------|--------|--------------|---------|--------|---------|-------|-------|------|-----------|--------|------|
| WRA                                                                                          | 22.534  | 159.96 | P     |        | 00 41 44.700 | 0.167   | 331.50 | -7.15   | 11.10 | 0.47  | TASD | 4.50      | 0.300  | 4.43 |
| WRA                                                                                          | 22.534  | 159.96 | S     |        | 00 45 48.700 | -4.364  | 338.00 | -0.65   | 17.00 | -2.34 | TA D | 2.00      | 0.900  |      |
| QIS                                                                                          | 25.328  | 149.77 | P1    | P      | 00 42 12.800 | 1.244   |        |         |       |       | T D  |           |        |      |
| QIS                                                                                          | 25.328  | 149.77 | PcP   |        | 00 45 44.800 | 0.066   |        |         |       |       | T D  |           |        |      |
| ASAR                                                                                         | 25.895  | 163.91 | P     |        | 00 42 16.900 | 0.181   | 346.30 | 3.89    | 7.10  | -1.96 | TA D | 3.40      | 0.500  | 4.24 |
| ASAR                                                                                         | 25.895  | 163.91 | PcP   |        | 00 45 45.200 | -0.823  | 345.10 | 2.69    | 2.30  | 0.01  | TASD | 2.20      | 0.500  |      |
| ASAR                                                                                         | 25.895  | 163.91 | S     |        | 00 46 45.300 | -2.629  | 347.60 | 5.19    | 20.30 | 4.48  | TA D | 3.90      | 0.800  |      |
| WARB                                                                                         | 27.329  | 179.36 | P     |        | 00 42 31.200 | 1.588   | 339.40 | -19.89  | 8.20  | -0.78 | T SD | 6.50      | 0.700  | 4.50 |
| WARB                                                                                         | 27.329  | 179.36 | PcP   |        | 00 45 49.400 | 0.097   |        |         |       |       | T D  |           |        |      |
| MEEK                                                                                         | 28.756  | 194.44 | P     |        | 00 42 42.700 | 0.321   |        |         |       |       | T    |           |        |      |
| CMAR                                                                                         | 31.809  | 304.11 | P     |        | 00 43 9.000  | -0.476  | 109.70 | -9.60   | 7.80  | -0.99 | TAS  | 0.60      | 0.400  | 3.88 |
| [ The following LR onset was only used with its backazimuth observation. ]                   |         |        |       |        |              |         |        |         |       |       |      |           |        |      |
| CMAR                                                                                         | 31.809  | 304.11 | LR    |        | 00 57 48.700 | 23.611  | 110.00 | -9.30   | 39.50 | 0.45  | A    | 188.80    | 19.360 | 3.88 |
| FORT                                                                                         | 31.949  | 177.17 | P     |        | 00 43 11.300 | 0.807   |        |         |       |       | T    |           |        |      |
| WOOL                                                                                         | 32.500  | 187.42 | P     |        | 00 43 15.600 | 0.239   | 7.90   | -0.75   | 9.90  | 1.13  | TAS  | 4.10      | 0.600  | 4.54 |
| SHK                                                                                          | 33.587  | 9.52   | P     |        | 00 43 25.400 | 0.573   |        |         |       |       | T    |           |        |      |
| KSAR                                                                                         | 35.990  | 2.12   | P     |        | 00 43 46.700 | 1.176   | 177.30 | -5.37   | 10.10 | 1.53  | TA   | 1.70      | 0.700  | 3.99 |
| KSAR                                                                                         | 35.990  | 2.12   | LR    |        | 01 04 25.000 | 257.097 | 160.00 | -22.67  | 46.00 | 6.96  |      | 40.10     | 19.860 | 3.26 |
| STKA                                                                                         | 36.040  | 157.60 | P     |        | 00 43 46.800 | 0.740   | 323.20 | -10.20  | 9.00  | 0.43  | T SD | 7.20      | 0.600  | 4.68 |
| STKA                                                                                         | 36.040  | 157.60 | PcP   |        | 00 46 12.000 | -0.606  |        |         |       |       | T D  |           |        |      |
| MJAR                                                                                         | 36.750  | 16.10  | P     |        | 00 43 51.900 | -0.210  | 357.10 | 156.96  | 18.80 | 10.28 | T    | 2.00      | 0.550  | 4.09 |
| PDY                                                                                          | 59.127  | 351.99 | P     |        | 00 46 47.000 | 1.992   | 111.20 | -52.89  | 6.60  | -0.33 | T S  | 4.30      | 0.450  | 4.78 |
| ZAL                                                                                          | 62.559  | 333.79 | P     |        | 00 47 9.000  | 0.523   |        |         |       |       | T    |           |        |      |
| ABKT                                                                                         | 72.094  | 309.51 | P     |        | 00 48 9.500  | 0.098   |        |         |       |       | T    |           |        |      |
| [ The unknown phase x was associated as P and the corresponding residuals were calculated. ] |         |        |       |        |              |         |        |         |       |       |      |           |        |      |
| NRI                                                                                          | 72.621  | 346.75 | x     | P      | 00 48 8.700  | -3.059  | 195.90 | 56.31   | 3.90  | -2.05 |      | 3.00      | 0.750  | 4.50 |
| MAW                                                                                          | 81.351  | 200.29 | P     |        | 00 49 1.100  | 0.155   |        |         |       |       | T    |           |        |      |
| KVAR                                                                                         | 84.499  | 313.86 | P     |        | 00 49 17.200 | -0.871  |        |         |       |       | T    |           |        |      |
| NPO                                                                                          | 87.457  | 25.36  | P     |        | 00 49 31.000 | -1.079  |        |         |       |       | T    |           |        |      |
| [ This P onset does not at all fit in the location with a fixed depth at 0 km. ]             |         |        |       |        |              |         |        |         |       |       |      |           |        |      |
| ARCES                                                                                        | 92.567  | 339.77 | P     |        | 00 49 51.700 | -4.215  | 94.50  | 15.08   | 4.10  | -0.52 | S    | 0.80      | 0.550  | 4.32 |
| SPITS                                                                                        | 92.763  | 348.81 | P     |        | 00 49 55.900 | -0.784  | 116.60 | 46.40   | 3.50  | -1.11 | T S  | 3.60      | 0.900  | 4.78 |
| FINES                                                                                        | 93.759  | 331.71 | P     |        | 00 50 0.000  | -1.509  | 111.20 | 30.82   | 5.90  | 1.30  | T S  | 0.60      | 0.550  | 4.16 |
| HFS                                                                                          | 99.955  | 332.03 | P     | Pdif   | 00 50 28.700 | -1.003  | 311.30 | -118.24 | 1.50  | -2.95 | T    | 1.10      | 0.750  | 4.57 |
| SCHQ                                                                                         | 123.011 | 9.03   | PKPdf |        | 00 55 41.200 | -0.098  |        |         |       |       | T    |           |        |      |
| TXAR                                                                                         | 123.387 | 53.22  | PKPdf |        | 00 55 42.700 | -0.215  |        |         |       |       | T    |           |        |      |
| DBIC                                                                                         | 130.629 | 279.87 | PKPdf |        | 00 55 57.100 | 0.161   |        |         |       |       | T    |           |        |      |
| PLCA                                                                                         | 137.880 | 160.81 | PKPdf |        | 00 56 9.100  | -0.877  | 311.90 | 106.27  | 6.40  | 4.56  | T    | 0.90      | 0.900  |      |

LPAZ 159.401 137.08 PKPdf 00 56 44.700 -0.776 T

Defining travel-time differences:

| Stat | Delta  | Phases   | Observed | Res    |
|------|--------|----------|----------|--------|
| WRA  | 22.534 | S - P    | 244.000  | -4.530 |
| QIS  | 25.328 | PcP - P1 | 212.000  | -1.178 |
| ASAR | 25.895 | PcP - P  | 208.300  | -1.005 |
| ASAR | 25.895 | S - P    | 268.400  | -2.810 |
| ASAR | 25.895 | S - PcP  | 60.100   | -1.805 |
| WARB | 27.329 | PcP - P  | 198.200  | -1.490 |
| STKA | 36.040 | PcP - P  | 145.200  | -1.346 |

[ The azimuth range of the maximum gap without any observations is always given in clockwise direction. ]  
Maximum azimuthal gap of defining observations: 53.2 [deg] -> 137.1 [deg] = 83.9 [deg]

|                               |                              |                            |
|-------------------------------|------------------------------|----------------------------|
| $\sqrt{\frac{\sum Res^2}{N}}$ | $\frac{\sum  Res }{N}$       | $\frac{\sum Res}{N}$       |
| [ RMS is defined as ]         | [ MEAN-ERROR is defined as ] | [ and MEAN is defined as ] |
| Residuals of defining data    | RMS                          | MEAN-ERROR                 |
| 32 onset times                | : 1.199                      | 0.826                      |
| 9 azimuth values              | : 5.858                      | 4.955                      |
| 10 ray parameters             | : 0.812                      | 0.709                      |
| 7 travel-time differences     | : 2.333                      | 2.023                      |
|                               |                              | -0.194 [s]                 |
|                               |                              | -2.336 [deg]               |
|                               |                              | -0.040 [s/deg]             |
|                               |                              | -2.023 [s]                 |

|                                            |                                                                                                                                                                           |
|--------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| $\sqrt{\frac{\sum Res^2 \cdot w}{\sum w}}$ |                                                                                                                                                                           |
| [ The weighted RMS is here defined as ]    | with the listed residuals Res and the data weight w used for the inversion (i.e., here the standard deviations of the data from <i>hyposat-in</i> ) as used at the ISC. ] |
| Weighted RMS of onset times (ISC type):    | 1.455 [s]                                                                                                                                                                 |

|                                                            |                                               |
|------------------------------------------------------------|-----------------------------------------------|
| $\frac{\sum \frac{ Res }{w}}{N}$                           | $\frac{\sum \left(\frac{Res}{w}\right)^2}{N}$ |
| [ The weighted misfit is here defined for the L1-norm as ] | [ and for the L2-norm as ]                    |
| Weighted misfit of input data                              | L1 L2                                         |
| 33 onset times                                             | : 2.744 0.450                                 |
| 20 azimuth values                                          | : 1.351 2.282                                 |
| 17 ray parameters                                          | : 1.475 2.736                                 |
| 7 travel-time differences                                  | : 3.637 3.893                                 |
| 77 misfit over all                                         | : 2.183 2.114                                 |

T0 LAT LON Z VPVS DLAT DLON DZ DT0 DVPVS DEF RMS  
1996-06-29 00 36 42.801 1.305 126.317 0.00 1.76 0.0250 0.0611 Fixed 0.150 0.04 58 1.199

[ However, we have still a fixed depth. Let us now try to fit the data better with another depth (see DEPTH FLAG is set to b! )

Iterations : 4  
Number of defining: 58  
First reference model : ak135  
Second reference model : iasp91

The new source parameters

Confidence level of given uncertainties: 68.27 %

Source time : 1996 06 29 00 36 49.169 0.478 [s]  
or 836008609.169 0.478 [s]  
Epicenter lat: 1.3211 0.0147 [deg]  
Epicenter lon: 126.2965 0.0348 [deg]  
Source depth : 44.50 4.28 [km]

[ Note the now much smaller error ellipse. ]  
Epicenter error ellipse:  
Major axes: 3.40 [km] Minor axes: 1.94 [km]  
Azimuth: 71.2 [deg] Area: 20.75 [km\*\*2]

Flinn-Engdahl Region ( 266 ): Northern Molucca Sea

Magnitudes: 4.36 (mb, G-R) 3.57 (Ms, R-P)

| Stat | Delta  | Azi    | Phase | [used] | Onset time   | Res     | Baz    | Res    | Rayp  | Res   | Used | Amplitude | Period | MAG  |
|------|--------|--------|-------|--------|--------------|---------|--------|--------|-------|-------|------|-----------|--------|------|
| WRA  | 22.556 | 159.93 | P     |        | 00 41 44.700 | -1.002  | 331.50 | -7.11  | 11.10 | 0.51  | TASD | 4.50      | 0.300  | 4.36 |
| WRA  | 22.556 | 159.93 | S     |        | 00 45 48.700 | -1.916  | 338.00 | -0.61  | 17.00 | 0.65  | TASD | 2.00      | 0.900  |      |
| QIS  | 25.353 | 149.75 | P1    | P      | 00 42 12.800 | 0.540   |        |        |       |       | T D  |           |        |      |
| QIS  | 25.353 | 149.75 | PcP   |        | 00 45 44.800 | 0.532   |        |        |       |       | T D  |           |        |      |
| ASAR | 25.916 | 163.87 | P     |        | 00 42 16.900 | -0.488  | 346.30 | 3.93   | 7.10  | -1.95 | TA D | 3.40      | 0.500  | 4.19 |
| ASAR | 25.916 | 163.87 | PcP   |        | 00 45 45.200 | -0.354  | 345.10 | 2.73   | 2.30  | 0.00  | TASD | 2.20      | 0.500  |      |
| ASAR | 25.916 | 163.87 | S     |        | 00 46 45.300 | 0.595   | 347.60 | 5.23   | 20.30 | 4.49  | TA D | 3.90      | 0.800  |      |
| WARB | 27.346 | 179.32 | P     |        | 00 42 31.200 | 0.983   | 339.40 | -19.85 | 8.20  | -0.76 | T SD | 6.50      | 0.700  | 4.41 |
| WARB | 27.346 | 179.32 | PcP   |        | 00 45 49.400 | 0.570   |        |        |       |       | T D  |           |        |      |
| MEEK | 28.767 | 194.40 | P     |        | 00 42 42.700 | -0.210  |        |        |       |       | T    |           |        |      |
| CMAR | 31.783 | 304.10 | P     |        | 00 43 9.000  | -0.659  | 109.70 | -9.59  | 7.80  | -0.99 | TAS  | 0.60      | 0.400  | 3.79 |
| CMAR | 31.783 | 304.10 | LR    |        | 00 57 48.700 | 18.251  | 110.00 | -9.29  | 39.50 | 0.45  | A    | 188.80    | 19.360 | 3.88 |
| FORT | 31.966 | 177.13 | P     |        | 00 43 11.300 | 0.247   |        |        |       |       | T    |           |        |      |
| WOOL | 32.514 | 187.39 | P     |        | 00 43 15.600 | -0.284  | 7.90   | -0.71  | 9.90  | 1.14  | TAS  | 4.10      | 0.600  | 4.47 |
| SHK  | 33.574 | 9.55   | P     |        | 00 43 25.400 | 0.298   |        |        |       |       | T    |           |        |      |
| KSAR | 35.974 | 2.15   | P     |        | 00 43 46.700 | 0.958   | 177.30 | -5.40  | 10.10 | 1.54  | TA   | 1.70      | 0.700  | 4.03 |
| KSAR | 35.974 | 2.15   | LR    |        | 01 04 25.000 | 251.344 | 160.00 | -22.70 | 46.00 | 6.96  |      | 40.10     | 19.860 | 3.26 |
| STKA | 36.063 | 157.58 | P     |        | 00 43 46.800 | 0.192   | 323.20 | -10.18 | 9.00  | 0.45  | T SD | 7.20      | 0.600  | 4.72 |
| STKA | 36.063 | 157.58 | PcP   |        | 00 46 12.000 | -0.198  |        |        |       |       | T D  |           |        |      |
| MJAR | 36.740 | 16.13  | P     |        | 00 43 51.900 | -0.465  | 357.10 | 156.92 | 18.80 | 10.29 | T    | 2.00      | 0.550  | 4.17 |
| PDY  | 59.108 | 352.00 | P     |        | 00 46 47.000 | 2.113   | 111.20 | -52.91 | 6.60  | -0.32 | S    | 4.30      | 0.450  | 4.79 |

[ This P onset has now a larger residuum than in the first run and is therefore not longer defining for the solution. ]

```

ZAL 62.536 333.80 P 00 47 9.000 0.712 T
ABKT 72.068 309.50 P 00 48 9.500 0.395 T
NRI 72.600 346.76 x P 00 48 8.700 -2.787 195.90 56.30 3.90 -2.04 3.00 0.750 4.37
MAW 81.360 200.28 P 00 49 1.100 0.351 T
KVAR 84.474 313.86 P 00 49 17.200 -0.469 T
NPO 87.451 25.36 P 00 49 31.000 -0.748 T
[ Note that the following P onset has now a smaller residuum but was still too large to be used as defining. ]
ARCES 92.544 339.77 P 00 49 51.700 -3.787 94.50 15.07 4.10 -0.52 S 0.80 0.550 4.29
SPITS 92.743 348.81 P 00 49 55.900 -0.366 116.60 46.39 3.50 -1.11 T S 3.60 0.900 4.74
FINES 93.735 331.71 P 00 50 0.000 -1.073 111.20 30.81 5.90 1.30 T S 0.60 0.550 4.19
HFS 99.931 332.03 P Pdif 00 50 28.700 -0.564 311.30 -118.25 1.50 -2.95 T 1.10 0.750 4.57
SCHQ 122.998 9.02 PKPdf 00 55 41.200 0.464 T
TXAR 123.393 53.20 PKPdf 00 55 42.700 0.310 T
DBIC 130.607 279.88 PKPdf 00 55 57.100 0.742 T
PLCA 137.902 160.82 PKPdf 00 56 9.100 -0.377 311.90 106.29 6.40 4.56 T 0.90 0.900
LPAZ 159.427 137.09 PKPdf 00 56 44.700 -0.241 T

```

Defining travel-time differences:

| Stat | Delta  | Phases   | Observed | Res    |
|------|--------|----------|----------|--------|
| WRA  | 22.556 | S - P    | 244.000  | -0.914 |
| QIS  | 25.353 | PcP - P1 | 212.000  | -0.008 |
| ASAR | 25.916 | PcP - P  | 208.300  | 0.134  |
| ASAR | 25.916 | S - P    | 268.400  | 1.083  |
| ASAR | 25.916 | S - PcP  | 60.100   | 0.949  |
| WARB | 27.346 | PcP - P  | 198.200  | -0.413 |
| STKA | 36.063 | PcP - P  | 145.200  | -0.390 |

[ Here we get the number of all iterations *e.g.*, also including an earlier solution for fixed depth. ]  
Total number of iterations: 18

Maximum azimuthal gap of defining observations: 53.2 [deg] -> 137.1 [deg] = 83.9 [deg]

| Residuals of defining data | RMS   | MEAN-ERROR | MEAN          |
|----------------------------|-------|------------|---------------|
| 31 onset times             | 0.658 | 0.558      | -0.049 [s]    |
| 9 azimuth values           | 5.862 | 4.957      | -2.316 [deg]  |
| 11 ray parameters          | 0.801 | 0.706      | 0.033 [s/deg] |
| 7 travel-time differences  | 0.681 | 0.556      | 0.063 [s]     |

Weighted RMS of onset times (ISC type): 0.727 [s]

| Weighted misfit of input data | L1    | L2    |
|-------------------------------|-------|-------|
| 33 onset times                | 2.217 | 0.140 |
| 20 azimuth values             | 1.351 | 2.282 |
| 17 ray parameters             | 1.429 | 2.726 |
| 7 travel-time differences     | 0.946 | 1.093 |
| 77 misfit over all            | 1.702 | 1.763 |

| T0                      | LAT   | LON     | Z     | VPVS | DLAT   | DLON   | DZ   | DT0   | DVPVS | DEF | RMS   |
|-------------------------|-------|---------|-------|------|--------|--------|------|-------|-------|-----|-------|
| 1996-06-29 00 36 49.169 | 1.321 | 126.297 | 44.50 | 1.76 | 0.0147 | 0.0348 | 4.28 | 0.478 | 0.04  | 58  | 0.658 |

-----end of the example-----

## 6 The Program HYPOMOD and the File hypomod-out

For a given seismic hypocenter solution, the program **HYPOMOD** calculates the residuals for all observed data: the travel times, the backazimuths, and slowness values. With the example given here for a hypocenter inversion with **HYPOSAT**, one has only to modify slightly the *hyposat-parameter* file and then one can apply the program **HYPOMOD**. The modifications needed in *hyposat-parameter* are to set for the starting source parameters the inversion results. Then you will get an output-file called *hypomod-out*, which has in principle the same format as *hyposat-out*.

--- example for changes in *hyposat-parameter* -----

**STARTING SOURCE LATITUDE [deg] : 1.3211**

**STARTING SOURCE LONGITUDE [deg] : 126.2965**

**STARTING SOURCE DEPTH [km] : 44.50**

**STARTING SOURCE TIME (epochal time): 836008609.169**

or

**STARTING SOURCE TIME (DOY) : 1996-181:00.36.49.169**

or

**STARTING SOURCE TIME (HUMAN) : 1996-06-29:00.36.49.169**

Then run **HYPOMOD** and you will get an output file *hypomod-out* which will look like:

-----example for a *hypomod-out* file -----

HYPOMOD Version 1.1

NORTHERN MOLUCCA SEA, 1996 29 June, from pIDC's Reviewed Event Bulletin (REB)

The source parameters

Source time : 1996 06 29 00 36 49.169  
 or 836008609.169  
 Epicenter lat: 1.3211 [deg]  
 Epicenter lon: 126.2965 [deg]  
 Source depth : 44.50 [km]

Flinn-Engdahl Region ( 266 ): Northern Molucca Sea

Magnitudes: 4.21 (mb, V-C) 3.57 (Ms, R-P)

| Stat  | Delta   | Azi    | Phase | [used] | Onset time   | Res     | Baz    | Res     | Rayp  | Res   | Used | Amplitude | Period | MAG  |
|-------|---------|--------|-------|--------|--------------|---------|--------|---------|-------|-------|------|-----------|--------|------|
| WRA   | 22.556  | 159.93 | P     |        | 00 41 44.700 | -1.002  | 331.50 | -7.11   | 11.10 | 0.51  | TASD | 4.50      | 0.300  | 4.22 |
| WRA   | 22.556  | 159.93 | S     |        | 00 45 48.700 | -1.916  | 338.00 | -0.61   | 17.00 | 0.65  | TASD | 2.00      | 0.900  |      |
| QIS   | 25.353  | 149.75 | P1    | P      | 00 42 12.800 | 0.540   |        |         |       |       | T D  |           |        |      |
| QIS   | 25.353  | 149.75 | PcP   |        | 00 45 44.800 | 0.532   |        |         |       |       | T D  |           |        |      |
| ASAR  | 25.916  | 163.87 | P     |        | 00 42 16.900 | -0.488  | 346.30 | 3.93    | 7.10  | -1.95 | TASD | 3.40      | 0.500  | 4.18 |
| ASAR  | 25.916  | 163.87 | PcP   |        | 00 45 45.200 | -0.354  | 345.10 | 2.73    | 2.30  | 0.00  | TASD | 2.20      | 0.500  |      |
| ASAR  | 25.916  | 163.87 | S     |        | 00 46 45.300 | 0.595   | 347.60 | 5.23    | 20.30 | 4.49  | TASD | 3.90      | 0.800  |      |
| WARB  | 27.346  | 179.32 | P     |        | 00 42 31.200 | 0.983   | 339.40 | -19.85  | 8.20  | -0.76 | TASD | 6.50      | 0.700  | 4.41 |
| WARB  | 27.346  | 179.32 | PcP   |        | 00 45 49.400 | 0.570   |        |         |       |       | T D  |           |        |      |
| MEEK  | 28.767  | 194.40 | P     |        | 00 42 42.700 | -0.210  |        |         |       |       | T D  |           |        |      |
| CMAR  | 31.783  | 304.10 | P     |        | 00 43 9.000  | -0.659  | 109.70 | -9.59   | 7.80  | -0.99 | TASD | 0.60      | 0.400  | 3.66 |
| CMAR  | 31.783  | 304.10 | LR    |        | 00 57 48.700 | 18.252  | 110.00 | -9.29   | 39.50 | 0.45  | A    | 188.80    | 19.360 | 3.88 |
| FORT  | 31.966  | 177.13 | P     |        | 00 43 11.300 | 0.247   |        |         |       |       | T D  |           |        |      |
| WOOL  | 32.514  | 187.39 | P     |        | 00 43 15.600 | -0.284  | 7.90   | -0.71   | 9.90  | 1.14  | TASD | 4.10      | 0.600  | 4.31 |
| SHK   | 33.574  | 9.55   | P     |        | 00 43 25.400 | 0.298   |        |         |       |       | T D  |           |        |      |
| KSAR  | 35.974  | 2.15   | P     |        | 00 43 46.700 | 0.958   | 177.30 | -5.40   | 10.10 | 1.54  | TASD | 1.70      | 0.700  | 3.82 |
| KSAR  | 35.974  | 2.15   | LR    |        | 01 04 25.000 | 251.343 | 160.00 | -22.70  | 46.00 | 6.96  | A    | 40.10     | 19.860 | 3.26 |
| STKA  | 36.063  | 157.58 | P     |        | 00 43 46.800 | 0.192   | 323.20 | -10.18  | 9.00  | 0.45  | TASD | 7.20      | 0.600  | 4.51 |
| STKA  | 36.063  | 157.58 | PcP   |        | 00 46 12.000 | -0.198  |        |         |       |       | T D  |           |        |      |
| MJAR  | 36.740  | 16.13  | P     |        | 00 43 51.900 | -0.465  | 357.10 | 156.92  | 18.80 | 10.29 | TASD | 2.00      | 0.550  | 3.99 |
| PDY   | 59.108  | 352.00 | P     |        | 00 46 47.000 | 2.112   | 111.20 | -52.91  | 6.60  | -0.32 | TASD | 4.30      | 0.450  | 4.47 |
| ZAL   | 62.536  | 333.80 | P     |        | 00 47 9.000  | 0.712   |        |         |       |       | T D  |           |        |      |
| ABKT  | 72.068  | 309.50 | P     |        | 00 48 9.500  | 0.395   |        |         |       |       | T D  |           |        |      |
| NRI   | 72.600  | 346.76 | x     | P      | 00 48 8.700  | -2.787  | 195.90 | 56.30   | 3.90  | -2.04 | A    | 3.00      | 0.750  | 4.17 |
| MAW   | 81.360  | 200.28 | P     |        | 00 49 1.100  | 0.351   |        |         |       |       | T D  |           |        |      |
| KVAR  | 84.474  | 313.86 | P     |        | 00 49 17.200 | -0.469  |        |         |       |       | T D  |           |        |      |
| NPO   | 87.451  | 25.36  | P     |        | 00 49 31.000 | -0.748  |        |         |       |       | T D  |           |        |      |
| ARCES | 92.544  | 339.77 | P     |        | 00 49 51.700 | -3.787  | 94.50  | 15.07   | 4.10  | -0.52 | TASD | 0.80      | 0.550  | 4.10 |
| SPITS | 92.743  | 348.81 | P     |        | 00 49 55.900 | -0.370  | 116.60 | 46.39   | 3.50  | -1.11 | TASD | 3.60      | 0.900  | 4.55 |
| FINES | 93.735  | 331.71 | P     |        | 00 50 0.000  | -1.073  | 111.20 | 30.81   | 5.90  | 1.30  | TASD | 0.60      | 0.550  | 4.06 |
| HFSH  | 99.931  | 332.03 | P     | Pdif   | 00 50 28.700 | -0.564  | 311.30 | -118.25 | 1.50  | -2.95 | TASD | 1.10      | 0.750  | 4.68 |
| SCHQ  | 122.998 | 9.02   | PKPdf |        | 00 55 41.200 | 0.464   |        |         |       |       | T D  |           |        |      |
| TXAR  | 123.393 | 53.20  | PKPdf |        | 00 55 42.700 | 0.310   |        |         |       |       | T D  |           |        |      |
| DBIC  | 130.606 | 279.88 | PKPdf |        | 00 55 57.100 | 0.742   |        |         |       |       | T D  |           |        |      |
| PLCA  | 137.902 | 160.82 | PKPdf |        | 00 56 9.100  | -0.377  | 311.90 | 106.29  | 6.40  | 4.56  | TASD | 0.90      | 0.900  | 3.96 |
| LPZA  | 159.427 | 137.09 | PKPdf |        | 00 56 44.700 | -0.241  |        |         |       |       | T D  |           |        |      |

Travel-time differences:

| Stat | Delta  | Phases  | Observed | Res    |
|------|--------|---------|----------|--------|
| WRA  | 22.556 | S - P   | 244.000  | -0.914 |
| QIS  | 25.353 | PcP - P | 212.000  | -0.008 |
| ASAR | 25.916 | PcP - P | 208.300  | 0.134  |
| ASAR | 25.916 | S - P   | 268.400  | 1.083  |
| ASAR | 25.916 | S - PcP | 60.100   | 0.949  |
| WARB | 27.346 | PcP - P | 198.200  | -0.413 |
| STKA | 36.063 | PcP - P | 145.200  | -0.390 |

Maximum azimuthal gap of defining observations: 53.2 [deg] -> 137.1 [deg] = 83.9 [deg]

| Residuals of data         | RMS    | MEAN-ERROR | MEAN          |
|---------------------------|--------|------------|---------------|
| 33 onset times            | 0.988  | 0.703      | -0.097 [s]    |
| 20 backazimuth values     | 55.011 | 34.014     | 8.352 [deg]   |
| 17 ray parameters         | 3.151  | 1.973      | 0.963 [s/deg] |
| 7 travel-time differences | 0.681  | 0.556      | 0.063 [s]     |

Weighted RMS of onset times (ISC type): 1.018 [s]

| Weighted misfit of input data | L1    | L2    |
|-------------------------------|-------|-------|
| 33 onset times                | 2.217 | 0.140 |
| 20 backazimuth values         | 1.351 | 2.282 |
| 17 ray parameters             | 1.429 | 2.726 |
| 7 travel-time differences     | 0.946 | 1.093 |
| 77 misfit over all            | 1.703 | 1.763 |

-----end of the example-----

## References

- Dziewonski, A. M., and Anderson, D. L. (1981). Preliminary reference Earth model. *Phys. Earth Planet. Inter.*, **25**, 297-356.
- Gutenberg, B., and Richter, C. F. (1956a). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Gutenberg, B., and Richter, C. F. (1956b). Earthquake magnitude, intensity, energy and acceleration. *Bull. Seism. Soc. Am.*, **46**, 105-145.
- Jeffreys, H., and Bullen, K. E. (1940, 1948, 1958, 1967, and 1970). Seismological Tables. *British Association for the Advancement of Science, Gray Milne Trust*, London, 50 pp.
- Kennett, B. L. N. (Ed.) (1991). IASPEI 1991 Seismological Tables. *Research School of Earth Sciences, Australian National University*, 167 pp.
- Kennett, B. L. N., and Engdahl, E. R. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, **105**, 429-465.
- Kennett, B. L. N., Engdahl, E. R., and Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.*, **122**, 108-124.
- Mooney, W. D., Laske, G., and Masters, T. G. (1998), CRUST 5.1: a global crustal model at 5° x 5°. *J. Geophys. Res.*, **103**, 727-747.
- Morelli, A., and Dziewonski, A. M. (1993). Body wave traveltimes and a spherically symmetric P- and S-wave velocity model. *Geophys. J. Int.*, **112**, 178-194.
- Rezapour, M., and Pearce, R. G. (1998). Bias in surface-wave magnitude  $M_S$  due to inadequate distance corrections. *Bull. Seism. Soc. Am.*, **88**, 1, 43-61.
- Schweitzer, J. (1997): HYPOSAT – a new routine to locate seismic events. *NORSAR Scientific Report 1-97/98*, 94-102, NORSAR, Kjeller, Norway, November 1997.
- Schweitzer, J. (2001a). HYPOSAT – An enhanced routine to locate seismic events. *Pure and Appl. Geophys.*, **158**, 277-289.
- Veith, K. F., and Clawson, G. E. (1972). Magnitude from short-period P-wave data. *Bull. Seism. Soc. Am.*, **62**, 435-452.

|                |                                                                                                                                                                                |
|----------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Topic</b>   | <b>LAUFZE / LAUFPS</b>                                                                                                                                                         |
| <b>Author</b>  | <b>Johannes Schweitzer</b> , NORSAR, P.O.Box 53, N-2027 Kjeller<br>Fax: +47 63818719, E-mail: <a href="mailto:johannes.schweitzer@norsar.no">johannes.schweitzer@norsar.no</a> |
| <b>Version</b> | LAUFZE 6.2 and LAUFPS 3.2 (as of April 2011);<br><b>DOI: 10.2312/GFZ.NMSOP-2_PD_11.2</b>                                                                                       |

## User Manual for LAUFZE and LAUFPS

### 1 Introduction

The program **LAUFZE** calculates travel-time curves for a P- or an S-velocity model. This is a newer version of a routine, which was originally developed in the 1970s at the Institute of Geophysics in Karlsruhe, Germany, by the late Prof. Gerhard Müller and Dr. Christoph Gelbke. Since then, the author has extended the code to include many new features and options, in particular for calculating different types of teleseismic and of multiply reflected or refracted phases. These changes were made at the Institute for Meteorology and Geophysics, University of Frankfurt, Germany, the Institute of Geophysics, Ruhr-University Bochum, Germany, and, most recently, at NORSAR.

The travel-time curves can be calculated for horizontally layered or spherically symmetric models with or without reduced time scale. The velocity model is defined by input as a function of the depth  $z$ , the dominant signal period  $T_{sig}$ , and the depth-dependent quality factor  $Q$  given for a reference period  $T_{ref}$ . Then, with this as input information, the program calculates and then uses the **group velocity** for the different model depths as defined by:

$$u(z, T, Q) = v(z) \cdot \left(1 + \frac{1}{\pi \cdot Q(z)} \cdot \left(\ln\left(\frac{T_{ref}}{T_{sig}}\right) + 1\right)\right)$$

If no  $Q$ -structure is given, the program uses as default value  $Q(z) = 10^6$ . However,  $Q$  is always assumed to be frequency independent.

The source can be placed in any depth as well as the ‘receivers’. However, the ‘receivers’ are by default assumed to be at the Earth’s surface. In the case of a spherical Earth model the Earth radius used for the Earth-flattening transformation is 6371 km. However, the Earth’s centre cannot be reached! The velocities for depths, at which they are not explicitly given, are linearly interpolated.

The whole program is based on the ray approximation of seismic waves, which means that the different kinds of seismic phases must be separately defined by the input parameters. Travel-time curves for the following phase types can be calculated:

- I Direct waves from the source to the Earth’s surface (only in the case that the source is not at the surface).
- II Diving waves from the source radiated down in the Earth.

- III Reflections from any layer below the source, back to the Earth's surface (*e.g.*, PcP, ScP, PmP).
- IV Reflections of diving waves at any layer back down into the Earth (*e.g.*, PP, SS, PKKP, SKKS).
- V Multiple [reflections](#) between any two layers (*e.g.*, PPmP, a diving P wave from [the](#) source to the Earth's surface, reflected from there back into the Earth, and finally reflected at the Mohorovičić (Moho) discontinuity back to the surface).
- VI For the phases III – V, any number of multiple reflections can be calculated (*e.g.*, P3, P5KP, SmS3).
- VII If the source is not at the surface, for all phases of II – VI the corresponding surface reflections can be calculated (*e.g.*, pP, sScS, pP3KP).
- VIII The multiple reflection(s) as defined under V will automatically be calculated for all above defined phases. That means, *e.g.*, not only for the direct P phase a Moho reverberation PPmP will be calculated but also *e.g.*, for pPKP we will get a pPKPPmP phase.
- IX For the direct to the surface radiated wave (see I) reflections can be calculated from any layer between source and the Earth's surface down back into the Earth (*e.g.*, p450P, or smS, but also pP).

All parameters for steering the program must be given in a formatted ASCII file. The program asks for the name of the input file.

All results of the program are written in a ASCII file called **laufze-out**. This file can then easily be edited and the listed travel-time curves can then be plotted with any plotting routine or used as ASCII input for other programs.

**LAUFPS** is a program like **LAUFZE** but it calculates travel times not only for one model (P or S) but also for both models together in one step, including converted phases. However, this program cannot be used to place 'receivers' below the Earth's surface. All settings for **RECE** (see below) are ignored!

The newest versions of the programs **laufze\_laufps** (including source code, this manual, data files containing examples for input and output files) are located in two compressed tar-files; either in **laufze.version.tar.Z** or in **laufze.version.tar.gz**. They can be downloaded for free either directly from the program list in the NMSOP-2 cover page folder *Download Programs & Files* or from NORSAR's anonymous ftp-server [ftp.norsar.no](ftp://ftp.norsar.no) under the directory */pub/outgoing/johannes/lauf*. If using your web-browser, the address is: <ftp://ftp.norsar.no/pub/outgoing/johannes/lauf>. Questions related to program updates and maintenance should be directed to the author.

## 2 Getting Started

This section describes how the example for **LAUFZE** can be started and executed. The simplest way to use the program for own travel-time calculations is to use the following examples and to modify the input data and parameters for your needs. The meaning and format of the contents of the input file are described in the following sections.

Installation of **LAUFZE**:

1) Make a sub-directory for **LAUFZE**, download the compressed tar-file containing the **laufze-software package**, decompress it, and run:

```
tar -xvf laufze.version.tar
```

You will then have a directory containing the following files and subdirectories:

*bin/ bin\_l/ examples/ man/ README src/*

The file *README* contains a complete list of all files included in the **laufze-software package** and a short explanation of these files.

2) If needed, recompile the software in the *src/* subdirectory by running:

```
make -f Makefile.laufze
```

and/or

```
make -f Makefile.laufps
```

The software was tested under UNIX as well under LINUX and should therefore run on both platforms without any compatibility problems. In the case of a LINUX system please use the corresponding *Makefiles* with the extension *\_linux\_g77* or *\_linux\_gfortran*, depending on your installed FORTRAN compiler.

3) Executing **LAUFZE**:

Change to the subdirectory *examples/*.

Here you will find an example for an input file.

To check your **LAUFZE** installation, try the following:

```
../bin/laufze
```

The program needs one input file in ASCII format. You will be asked for the name of the input file and here you answer with **laufze-in**. This file contains all parameters to steer the travel-time calculations and the layered velocity model. All results of the program are written in a file called **laufze-out**. The output file you get should be identical to the file **laufze-out.test** distributed with the **laufze-software package**. Contents and structure of these files will be explained in the following sections.

The program **LAUFPS** uses the same input file format as **LAUFZE** but it needs three files, one containing the P-velocity model and ray definitions for the requested P phases, one file containing the S-velocity model and ray definitions for the requested S phases and one file containing additional parameters to steer the requested converted phases for both P-to-S and S-to-P types of phase conversions. After starting the program with

```
../bin/laufps
```

you will be asked for names of the files containing the P- and S-velocity models and the file



to steer the behaviour of all P-to-SV and/or SV-to-P conversions (*e.g.*, *laufp-dat*, *laufs-dat*, and *laufps-in*). If you use the files as delivered with the *laufze-software package*, the LAUFPS output file *laufps-out* should be identical to the file *laufps-out.test*.

### 3 The File *laufze-in*

The file *laufze-in* must contain the velocity model for P (or S) waves and all information about the seismic phases for which travel times shall be calculated. The program asks for the name of the file containing this information in the format below described. The user can of course use any file name.

----- example for a *laufze-in* file -----

For a source in 100 km, Jeffreys-Bullen Model

```

0.000 180.000 0.00 0 0
0.000 0.000 0 0 0
0.0 6.11 STRU 0.000 ! for pP
33.0 6.11
33.0 7.76
100. 7.95 SOUR
200. 8.26
300. 8.58
413. 8.97
600. 10.25
800. 11.00
1000. 11.42
1200. 11.71
1400. 11.99
1600. 12.26
1800. 12.53
2000. 12.79
2200. 13.03
2400. 13.27
2600. 13.50
2800. 13.64
2898. 13.64 REFL ! for PcP
2898. 8.10
3000. 8.22
3200. 8.47
3400. 8.76
3600. 9.04
3800. 9.28
4000. 9.51
4200. 9.70
4400. 9.88
4600. 10.06
4800. 10.25
4982. 10.44
5121. 10.44 REFL ! for PKiKP
5121. 11.16
5700. 11.26

! blank line, no more layers

1 1 20
1 1 ! for PP
1 2 ! for P3
21 1 ! for PKKP
```

```

21          2          ! for P3KP
                ! blank line: no more such mult. phases
1          ! for PcPPcP + PKiKPPKiKP
                ! blank line: no more such mult. phases
----- end of the example-----

```

The contents of the *laufze-in* file is as follows:

1. 1 line of maximum 80 characters with any explaining text as **TITLE**.
2. 1 line in FORMAT (3F10.3,3I5) containing the parameters: **RMIN**, **RMAX**, **VR**, **IELAS**, **NF11**.
 

**RMIN** is the beginning of the distance range from which onset times are calculated. Either measured in [deg] or in [km], see the definition of **VR**.

**RMAX** as **RMIN** but the end of the distance range used to print out the travel-time branches.

**VR** is the velocity or slowness to reduce the travel times.  
 If **VR** > 0, **RMIN** and **RMAX** are measured in [km] and **VR** is a reduction velocity [km/s].  
 If **VR** <= 0, **RMIN** and **RMAX** are measured in [deg] and **VR** is a reduction slowness [s/deg].

**IELAS** if this flag is set to 1,  $Q(z)$  is set to  $10^6$  in all depths (pure elastic case).

**NF11** if this flag is set to 1, no direct up-going rays from a source below the surface are calculated.
3. 1 line in FORMAT (2F10.3,3I5) containing the parameters: **PER**, **PERREF**, **LA1**, **LA2**, **NLA**.
 

**PER** is the dominant signal period  $T_{\text{sig}}$  [s].  
 If **PER** is set to 0 s, the default value of 1 s is used.

**PERREF** is the reference period  $T_{\text{ref}}$  [s].  
 If **PERREF** is set to 0 s, the default value of 1 s is used.

**LA1** is the number of the upper layer for the described reverberations (see V of the program options in the **Introduction**).

**LA2** as **LA1**, now the number of the lower layer.

**NLA** gives the number of reverberations.  
 The multiple phase travels **NLA** times more often through the depth range  $Z(\text{LA1}) \leq Z(I) < Z(\text{LA2})$  than the regular phase. If **LA1** = **LA2** or **NLA** = 0, no reverberations are calculated.
4. Now follows the model. The model is defined by one line for each depth with velocity information. All lines must fit in the FORMAT (2F10.3,A4,6x,F10.3) and contain the parameters **Z**, **V**, **AZ**, **QU**. The model can contain a maximum of 1000 layers. First

order discontinuities for one of the given parameters have to be defined by 2 lines in the same depth **Z**.

**Z** depth in [km] below the Earth's surface. The surface has the depth **Z**(1) = 0.

**V** seismic velocity in the depth **Z**.

**AZ** four characters long key words, with which one can define special actions in this depth:

= **SOUR** means that the seismic source is located in this depth of the model.

= **RECE** means that the 'receivers' are in this depth (to be set only if not at the surface, by default receivers are assumed to be at the surface).

= **SORE** means a combination of both **SOUR** and **RECE** in the same depth.

= **SURF** means that the seismic source is located in this depth and that for all (!) phases their corresponding surface reflections are additionally calculated (*e.g.*, pP, sScS, ...).

= **SURE** means a combination of both **SURF** and **RECE** in the same depth.

= **STRU** means that an up-going direct ray is reflected in this depth back down into the Earth (see IX of the program options in the **Introduction**).

If **STRU** is set at the surface, the classical surface reflection is calculated (*i.e.*, only pP or sS but not *e.g.*, pPP or sScS).

= **REFL** means that steep-angle reflections from this depth are calculated (see III of the program options in the **Introduction**).

**AZ** must be blank in all other cases.

**QU** is the quality factor for seismic waves in this depth; if **QU** = 0. the program sets it by default to **QU** = 1 000 000.

An empty line finishes the model input.

5. 1 line in FORMAT (3I5) with the three parameters **IS**, **IA**, **IB**.

**IS** = 0 the input model is assumed to be flat, *i.e.*, it consists of a set of horizontally flat layers.

= 1 the input model is spherical and has to be modified by the Earth-flattening transformation.

**IA** = I ; only the parts of the travel-time curves, which have their turning points below the I'th layer, are calculated. If **IA** = 1, all travel-time branches of all phases are calculated.

**IB** gives the number of rays, which will have their turning points between two given depth points of the model; an **IB** value of 10 – 20 usually gives a good approximation of the travel-time branch.

6. In the following line(s), the reflections of diving rays at any layer back down into the Earth can be defined (see IV of the program options in the **Introduction**). For each (!) such reflection 1 line in FORMAT (2I10) is needed with the parameters **IKMG** and **MULT**. If no (further) reflections of this type are to be calculated, one has to give a blank line.

**IKMG** is the number of the layer at which the ray is reflected; *e.g.*, for PP one has to set **IKMG** = 1.

**MULT** gives the number of reflections at this reflector (*e.g.*, for PP, SS, or PKKP one has to set **MULT** = 1, and for P3, S3, or S3KS one has to use **MULT** = 2).

7. Finally, multiple reflections for the steep-angle reflections as defined with **AZ** = **REFL** can be ordered with the following line(s) in FORMAT (I10) containing the parameter **MULTR**. No (further) multiples of this type have to be indicated by another blank line.

**MULTR** gives the number of additional multiples for each steep-angle reflection. **MULTR** = 1 will, for example, result in PmP2 (= PmPPmP) or ScS2 (= ScSScS) and **MULT** = 2 will give, *e.g.*, ScS3.

## 4 The File *laufze-out*

With the above listed example for a *laufze-in* file, you will obtain the output-file *laufze-out*. Please note that the output file has been truncated by numerous lines to reduce the number of pages in this manual. The ASCII listing of the travel times can easily be extracted and the user can plot them with any plotting program after some simple editing work. The original output file has 3117 lines and is included in the *laufze-software package*. Explanations are included in [ .... ]:

-----example for a *laufze-out* file -----

Travel times from LAUFZE 6.2

For a source in 100 km, Jeffreys-Bullen Model

Distance range

RMIN = 0.000 deg

RMAX = 180.000 deg

Ray parameter to reduce travel times P = 0.000 s/deg

The travel times are calculated for a group velocity  
at a reference period of 1.000 s

Model input (depth, velocity)

+ modified velocity-depth function after application  
of the Earth-flattening transformation

| Z       | V(Z)  | Q(Z)      | U(Z,Q,PER) | AZ(Z) | FLATT EARTH |       |
|---------|-------|-----------|------------|-------|-------------|-------|
|         |       |           |            |       | [ Z-FL      | U-FL] |
| 0.000   | 6.110 | 1000000.0 | 6.110      | STRU  | 0.000       | 6.110 |
| 33.000  | 6.110 | 1000000.0 | 6.110      |       | 33.086      | 6.142 |
| 33.000  | 7.760 | 1000000.0 | 7.760      |       | 33.086      | 7.800 |
| 100.000 | 7.950 | 1000000.0 | 7.950      | SOUR  | 100.793     | 8.077 |

|          |        |           |        |                |         |
|----------|--------|-----------|--------|----------------|---------|
| 200.000  | 8.260  | 1000000.0 | 8.260  | 203.207        | 8.528   |
| 300.000  | 8.580  | 1000000.0 | 8.580  | 307.293        | 9.004   |
| 413.000  | 8.970  | 1000000.0 | 8.970  | 426.995        | 9.592   |
| 600.000  | 10.250 | 1000000.0 | 10.250 | 630.162        | 11.316  |
| 800.000  | 11.000 | 1000000.0 | 11.000 | 854.873        | 12.580  |
| 1000.000 | 11.420 | 1000000.0 | 11.420 | 1087.799       | 13.546  |
| 1200.000 | 11.710 | 1000000.0 | 11.710 | 1329.566       | 14.427  |
| 1400.000 | 11.990 | 1000000.0 | 11.990 | 1580.871       | 15.367  |
| 1600.000 | 12.260 | 1000000.0 | 12.260 | 1842.496       | 16.372  |
| 1800.000 | 12.530 | 1000000.0 | 12.530 | 2115.328       | 17.464  |
| 2000.000 | 12.790 | 1000000.0 | 12.790 | 2400.367       | 18.642  |
| 2200.000 | 13.030 | 1000000.0 | 13.030 | 2698.760       | 19.903  |
| 2400.000 | 13.270 | 1000000.0 | 13.270 | 3011.817       | 21.290  |
| 2600.000 | 13.500 | 1000000.0 | 13.500 | 3341.056       | 22.808  |
| 2800.000 | 13.640 | 1000000.0 | 13.640 | 3688.241       | 24.335  |
| 2898.000 | 13.640 | 1000000.0 | 13.640 | REFL 3865.526  | 25.022  |
| 2898.000 | 8.100  | 1000000.0 | 8.100  | 3865.526       | 14.859  |
| 3000.000 | 8.220  | 1000000.0 | 8.220  | 4055.441       | 15.535  |
| 3200.000 | 8.470  | 1000000.0 | 8.470  | 4445.107       | 17.017  |
| 3400.000 | 8.760  | 1000000.0 | 8.760  | 4860.167       | 18.785  |
| 3600.000 | 9.040  | 1000000.0 | 9.040  | 5304.164       | 20.785  |
| 3800.000 | 9.280  | 1000000.0 | 9.280  | 5781.437       | 22.996  |
| 4000.000 | 9.510  | 1000000.0 | 9.510  | 6297.381       | 25.554  |
| 4200.000 | 9.700  | 1000000.0 | 9.700  | 6858.818       | 28.466  |
| 4400.000 | 9.880  | 1000000.0 | 9.880  | 7474.555       | 31.936  |
| 4600.000 | 10.060 | 1000000.0 | 10.060 | 8156.231       | 36.190  |
| 4800.000 | 10.250 | 1000000.0 | 10.250 | 8919.681       | 41.568  |
| 4982.000 | 10.440 | 1000000.0 | 10.440 | 9704.131       | 47.886  |
| 5121.000 | 10.440 | 1000000.0 | 10.440 | REFL 10375.893 | 53.211  |
| 5121.000 | 11.160 | 1000000.0 | 11.160 | 10375.893      | 56.880  |
| 5700.000 | 11.260 | 1000000.0 | 11.260 | 14339.481      | 106.911 |

Travel-time branch 1 and all the following ones were calculated,  
each travel-time branch consists of 20 rays

#### P H A S E :

[ If **NF11** is set to 1, this wave will not be calculated ]  
Directly upgoing wave

[ DELTA is the distance measured in [km] or in [deg], see input parameter **VR**, TT is the (eventually reduced) travel time, AIN is the radiation angle at the source,  $T^*$  is the travel time divided by the mean quality factor  $TT/Q$ , ATT is the amplitude attenuation due to  $T^*$ , i.e.,  $ATT = \exp(-2 \cdot \pi \cdot f_{ref} \cdot T^*)$ , and P is the ray parameter measured in [s/deg] ]

| DELTA | TT     | AIN     | $T^*$ | ATT   | P     |
|-------|--------|---------|-------|-------|-------|
| 0.000 | 13.931 | 180.000 | 0.000 | 1.000 | 0.000 |
| 0.033 | 13.940 | 177.722 | 0.000 | 1.000 | 0.547 |
| 0.066 | 13.967 | 175.443 | 0.000 | 1.000 | 1.094 |
| 0.099 | 14.012 | 173.165 | 0.000 | 1.000 | 1.639 |
| 0.132 | 14.075 | 170.886 | 0.000 | 1.000 | 2.181 |
| 0.166 | 14.158 | 168.608 | 0.000 | 1.000 | 2.719 |
| 0.200 | 14.260 | 166.329 | 0.000 | 1.000 | 3.254 |

|       |        |         |       |       |        |
|-------|--------|---------|-------|-------|--------|
| 0.234 | 14.381 | 164.051 | 0.000 | 1.000 | 3.783  |
| 0.269 | 14.524 | 161.772 | 0.000 | 1.000 | 4.306  |
| 0.305 | 14.688 | 159.494 | 0.000 | 1.000 | 4.823  |
| 0.342 | 14.874 | 157.215 | 0.000 | 1.000 | 5.332  |
| 0.380 | 15.085 | 154.937 | 0.000 | 1.000 | 5.832  |
| 0.419 | 15.321 | 152.658 | 0.000 | 1.000 | 6.323  |
| 0.459 | 15.584 | 150.380 | 0.000 | 1.000 | 6.804  |
| 0.500 | 15.877 | 148.101 | 0.000 | 1.000 | 7.275  |
| 0.543 | 16.201 | 145.823 | 0.000 | 1.000 | 7.734  |
| 0.588 | 16.559 | 143.544 | 0.000 | 1.000 | 8.181  |
| 0.635 | 16.954 | 141.266 | 0.000 | 1.000 | 8.614  |
| 0.685 | 17.389 | 138.987 | 0.000 | 1.000 | 9.034  |
| 0.737 | 17.870 | 136.709 | 0.000 | 1.000 | 9.440  |
| 0.792 | 18.401 | 134.430 | 0.000 | 1.000 | 9.831  |
| 0.850 | 18.988 | 132.152 | 0.000 | 1.000 | 10.207 |
| 0.913 | 19.638 | 129.873 | 0.000 | 1.000 | 10.566 |
| 0.980 | 20.359 | 127.595 | 0.000 | 1.000 | 10.908 |
| 1.053 | 21.162 | 125.316 | 0.000 | 1.000 | 11.234 |
| 1.131 | 22.059 | 123.038 | 0.000 | 1.000 | 11.541 |
| 1.218 | 23.067 | 120.759 | 0.000 | 1.000 | 11.830 |
| 1.313 | 24.204 | 118.481 | 0.000 | 1.000 | 12.101 |
| 1.418 | 25.494 | 116.203 | 0.000 | 1.000 | 12.353 |
| 1.536 | 26.969 | 113.924 | 0.000 | 1.000 | 12.584 |
| 1.670 | 28.667 | 111.646 | 0.000 | 1.000 | 12.796 |
| 1.823 | 30.640 | 109.367 | 0.000 | 1.000 | 12.988 |
| 2.000 | 32.953 | 107.089 | 0.000 | 1.000 | 13.159 |
| 2.207 | 35.688 | 104.810 | 0.000 | 1.000 | 13.310 |
| 2.451 | 38.953 | 102.532 | 0.000 | 1.000 | 13.439 |
| 2.742 | 42.881 | 100.253 | 0.000 | 1.000 | 13.547 |
| 3.091 | 47.631 | 97.975  | 0.000 | 1.000 | 13.634 |
| 3.512 | 53.386 | 95.696  | 0.000 | 1.000 | 13.699 |
| 4.018 | 60.330 | 93.418  | 0.000 | 1.000 | 13.743 |
| 4.621 | 68.624 | 91.139  | 0.000 | 1.000 | 13.765 |

P H A S E :

Diving wave

[ DELTA is the distance measured in [km] or in [deg], see input parameter **VR**, TT is the (eventually reduced) travel time, AIN is the radiation angle at the source,  $T^*$  is the travel time divided by the mean quality factor  $TT/Q$ , ATT is the amplitude attenuation due to  $T^*$ , i.e.,  $ATT = \exp(-2\pi f_{ref} T^*)$ , P is the ray parameter measured in [s/deg], and for diving waves also the depth [km] of the ray's turning point is given. ]

| DELTA  | TT      | AIN    | $T^*$ | ATT   | P      | ZS      |
|--------|---------|--------|-------|-------|--------|---------|
| 4.961  | 73.311  | 90.000 | 0.000 | 1.000 | 13.767 | 100.000 |
| 6.323  | 92.033  | 85.613 | 0.000 | 1.000 | 13.727 | 105.303 |
| 6.995  | 101.248 | 83.803 | 0.000 | 1.000 | 13.687 | 110.602 |
| 7.549  | 108.820 | 82.420 | 0.000 | 1.000 | 13.647 | 115.897 |
| 8.040  | 115.504 | 81.258 | 0.000 | 1.000 | 13.607 | 121.186 |
| 8.488  | 121.598 | 80.238 | 0.000 | 1.000 | 13.568 | 126.472 |
| 8.906  | 127.257 | 79.319 | 0.000 | 1.000 | 13.529 | 131.753 |
| 9.299  | 132.574 | 78.477 | 0.000 | 1.000 | 13.490 | 137.029 |
| 9.673  | 137.613 | 77.696 | 0.000 | 1.000 | 13.451 | 142.301 |
| 10.031 | 142.418 | 76.966 | 0.000 | 1.000 | 13.413 | 147.569 |
| 10.375 | 147.021 | 76.277 | 0.000 | 1.000 | 13.374 | 152.832 |

|        |         |        |       |       |        |         |
|--------|---------|--------|-------|-------|--------|---------|
| 10.706 | 151.448 | 75.624 | 0.000 | 1.000 | 13.336 | 158.091 |
| 11.027 | 155.720 | 75.003 | 0.000 | 1.000 | 13.298 | 163.345 |
| 11.338 | 159.853 | 74.409 | 0.000 | 0.999 | 13.261 | 168.595 |
| 11.641 | 163.860 | 73.840 | 0.000 | 0.999 | 13.223 | 173.840 |
| 11.936 | 167.752 | 73.292 | 0.000 | 0.999 | 13.186 | 179.081 |
| 12.223 | 171.539 | 72.765 | 0.000 | 0.999 | 13.149 | 184.317 |
| 12.504 | 175.230 | 72.255 | 0.000 | 0.999 | 13.112 | 189.549 |
| 12.779 | 178.832 | 71.762 | 0.000 | 0.999 | 13.076 | 194.777 |
| 13.049 | 182.351 | 71.284 | 0.000 | 0.999 | 13.039 | 200.000 |

[ Each layer of the model gives one block of rays (number of rays given as parameter **IB**) and forms one 'branch' of the travel-time curve. These rays are always written as one block in the listing. If one plans to plot the travel-time curves, one should plot each block as separate branch of the travel-time curve. Here all blocks with rays bottoming in the mantle were omitted! ]

[ The next possible rays are the phases bottoming in the Earth's core: here PKPab, PKPbc ]

|         |          |        |       |       |       |          |
|---------|----------|--------|-------|-------|-------|----------|
| 178.309 | 1310.499 | 18.832 | 0.001 | 0.996 | 4.444 | 3959.714 |
| 173.109 | 1287.400 | 18.810 | 0.001 | 0.996 | 4.439 | 3961.852 |

[ All rays bottoming in the outer core were deleted! ]

|         |          |       |       |       |       |          |
|---------|----------|-------|-------|-------|-------|----------|
| 154.616 | 1186.983 | 8.731 | 0.001 | 0.996 | 2.090 | 5121.000 |
|---------|----------|-------|-------|-------|-------|----------|

[ The inner core boundary (ICB) is a first order discontinuity with a positive velocity jump. The following rays 'bottoming' in the boundary build the over-critical part of the travel-time curve of PKiKP, the over-critical reflection from the ICB! ]

|         |          |       |       |       |       |          |
|---------|----------|-------|-------|-------|-------|----------|
| 154.616 | 1186.983 | 8.731 | 0.001 | 0.996 | 2.090 | 5121.000 |
| 144.603 | 1166.084 | 8.699 | 0.001 | 0.996 | 2.082 | 5121.000 |
| 140.610 | 1157.784 | 8.667 | 0.001 | 0.996 | 2.075 | 5121.000 |
| 137.605 | 1151.561 | 8.636 | 0.001 | 0.996 | 2.067 | 5121.000 |
| 135.111 | 1146.414 | 8.605 | 0.001 | 0.996 | 2.060 | 5121.000 |
| 132.943 | 1141.955 | 8.574 | 0.001 | 0.996 | 2.052 | 5121.000 |
| 131.006 | 1137.987 | 8.543 | 0.001 | 0.996 | 2.045 | 5121.000 |
| 129.244 | 1134.390 | 8.513 | 0.001 | 0.996 | 2.038 | 5121.000 |
| 127.621 | 1131.087 | 8.482 | 0.001 | 0.996 | 2.031 | 5121.000 |
| 126.111 | 1128.026 | 8.452 | 0.001 | 0.996 | 2.024 | 5121.000 |
| 124.695 | 1125.167 | 8.423 | 0.001 | 0.996 | 2.017 | 5121.000 |
| 123.361 | 1122.481 | 8.393 | 0.001 | 0.996 | 2.009 | 5121.000 |
| 122.096 | 1119.944 | 8.364 | 0.001 | 0.996 | 2.002 | 5121.000 |
| 120.893 | 1117.538 | 8.334 | 0.001 | 0.996 | 1.996 | 5121.000 |
| 119.744 | 1115.249 | 8.305 | 0.001 | 0.997 | 1.989 | 5121.000 |
| 118.644 | 1113.065 | 8.277 | 0.001 | 0.997 | 1.982 | 5121.000 |
| 117.587 | 1110.975 | 8.248 | 0.001 | 0.997 | 1.975 | 5121.000 |
| 116.571 | 1108.971 | 8.220 | 0.001 | 0.997 | 1.968 | 5121.000 |
| 115.590 | 1107.044 | 8.191 | 0.001 | 0.997 | 1.962 | 5121.000 |
| 114.643 | 1105.190 | 8.163 | 0.001 | 0.997 | 1.955 | 5121.000 |

[ Now followed the deleted PKPdf branch. ]

Surface reflection of the direct wave

[ See the parameters **STRU** and/or **SURF** in the input file. The ray output for pP, pPKP, and pPKiKP (over-critical part) was deleted. ]

P H A S E :

Diving wave      1-times reflected at the Earth's surface

[ See the parameters **IKMG** and **MULT** in the input file. The ray output for PP, P'P', and PKiKP2 (over-critical part) was deleted. ]

P H A S E :

Diving wave      2-times reflected at the Earth's surface

[ See the parameters **IKMG** and **MULT** in the input file. The ray output for P3, P'3, and PKiKP3 (over-critical part) was deleted. ]

P H A S E :

Diving wave      1-times reflected down at layer    21

[ See the parameters **IKMG** and **MULT** in the input file; layer 21 is here the core-mantle boundary. The ray output for PKKP and PKiKKiKP (over-critical part) was deleted. ]

P H A S E :

Diving wave      2-times reflected down at layer    21

[ See the parameters **IKMG** and **MULT** in the input file; layer 21 is here the core-mantle boundary (CMB). The ray output for P3KP and P3KiKP (over-critical part) was deleted. ]

P H A S E :

Steep-angle reflection from    2898.000 km

[ See the parameter **AZ = REFL** in the input file. Steep-angle reflection (*i.e.*, before the critical point) from the CMB; the ray output for PcP was deleted. ]

P H A S E :

Steep-angle reflection from    5121.000 km

[ See the parameter **AZ = REFL** in the input file. Steep-angle reflection (*i.e.*, before the critical point) from the ICB; the ray output for PKiKP was deleted. ]

P H A S E :

Multiple reflection (    1-times) for the  
Steep-angle reflection from    2898.000 km

[ See the parameters **AZ = REFL** and **MULTR** in the input file. Multiple steep-angle reflection (*i.e.*, before the critical point) from the CMB; the ray output for PcP2 was deleted. ]



P H A S E :

Multiple reflection ( 1-times) for the  
Steep-angle reflection from 5121.000 km

[ See the parameters **AZ** = **REFL** and **MULTR** in the input file. Multiple steep-angle reflection (*i.e.*, before the critical point) from the ICB; the ray output for PKiKP2 was deleted. ]

-----end of example for a *laufze-out* file -----

## 5 The Program LAUFPS and the File *laufps-in*

The program **LAUFPS** calculates travel-time curves for a given velocity model, as the program **LAUFZE** does, and was developed on the base of **LAUFZE**. In addition to **LAUFZE**, **LAUFPS** calculates P- and S-phase travel-time curves in one step and it can also calculate travel-time curves for converted phases. The input for **LAUFPS** consists of one steering and two model files:

One the model files contains the P-velocity model and one contains the S-velocity model. Both model files must have the same format as a **LAUFZE** input file, e.g., one file is then called *laufp-dat* and one *laufs-dat*. Both model files must sample the velocity models (*i.e.*, the P and the S model) at identical depths and the source must also be in the same depth. However, these input files can be extended at three points to inform the program in the case of multiple phases, how often these reverberations are eventually travelling through the model as converted phase. These additions are the following ones (I refer to the number of the format descriptions of the input-file for **LAUFZE**):

- 1 line in FORMAT (2F10.3,4I5) containing the parameters: **PER**, **PERREF**, **LA1**, **LA2**, **NLA**, **NLA2**.

**PER** is the dominant signal period  $T_{\text{sig}}$  [s].

If **PER** is set to 0 s, the default value of 1 s is used.

**PERREF** is the reference period  $T_{\text{ref}}$  [s].

If **PERREF** is set to 0 s, the default value of 1 s is used.

**LA1** is the number of the upper layer for the described reverberations (see V of the program options in the **Introduction**).

**LA2** as **LA1**, now the number of the lower layer.

**NLA** gives the number of reverberations.

The multiples travel through the depth range  $Z(\text{LA1}) \leq Z(I) < Z(\text{LA2})$  **NLA** times more than the regular phase. If **LA1** = **LA2** or **NLA** = 0, no reverberations are calculated.

**NLA2** gives how many of the **NLA** reverberations are travelling as converted phase (**NLA2** must be  $\leq \text{NLA}$ ).

6. In the following line(s) the reflections of diving rays at any layer back down into the Earth can be defined (see IV of the program options in the **Introduction**). For each (!) such reflection 1 line in FORMAT (3I10) is needed with the parameters **IKMG**, **MULT**, and **MULT2**. If no (further) reflections of this type shall be calculated one has to add a blank line.

**IKMG** is the number of the layer at which the ray is reflected; *e.g.*, for the surface reflection PP one has to set **IKMG** = 1.

**MULT** gives the number of reflections at this reflector (*e.g.*, for PP, SS, or PKKP one has to set **MULT** = 1, for P3, S3, or S3KS one has to use **MULT** = 2).

**MULT2** gives how many of the **MULT** reverberations are travelling as a converted phase (**MULT2** must be  $\leq$  **MULT**), *e.g.*, the travel-time curve of PPS will need to set **IKMG** = 1, **MULT** = 2, and **MULT2** = 1.

8. Finally, multiple reflections for the steep-angle reflections as defined with **AZ** = **REFL**, can be ordered with the following line(s) in FORMAT (2I10) containing the parameter **MULTR** and **MULTR2**. No (further) multiples of this type have to be indicated by another blank line.

**MULTR** give the number of multiples for each order steep-angle reflection. **MULTR** = 1 will *e.g.*, result in PmP2 or ScS2 and **MULT** = 2 will give *e.g.*, ScS3.

**MULTR2** gives how many of the **MULTR** reverberations are travelling as converted phase (**MULTR2** must be  $\leq$  **MULTR**), *e.g.*, the travel-time curve of PcPScS will need to set **MULTR** = 2, and **MULTR2** = 1.

The directory *examples/* contains files with one P- (*laufp-dat*) and one S-velocity model (*laufs-dat*) applying some of the mentioned settings.

In addition to these two model related files, the program **LAUFPS** needs one file containing steering parameters for each travel-time curve to define further phase conversions. It is recommended to run the program in a first step with setting the parameter **KONSAR** to 0 and then editing the steering file again. This steering file (here *e.g.*, called *laufps-in*) must contain the following information in the described format:

- a) In the first line the parameter **KONSOR** in FORMAT (I5). **KONSOR** steers the general behaviour of **LAUFPS**:

= 0 ; no conversions are calculated (not even the ones defined above!).  
 = 1 ; only conversions from P to SV type phases are calculated.  
 = 2 ; only conversions from SV to P type phases are calculated.  
 = 3 ; all types of conversions are calculated.

- b) For each (!) phase, as calculated in a first test run after setting **KONSOR** = 0 and then listed in *laufps-out* with an own phase header, one has to add one line with the parameters **KON**, **KON1**, **NDISK1**, **KON2**, **NDISK2** in FORMAT (5I5). For the test run, one can just add a large number of empty lines.

**KON** = 0 ; no conversion for this phase.

- = 1 ; this is a P phase and we get a P to SV conversion.
- = 2 ; this is an S phase and we get a SV to P conversion.
- = 9 ; the next entry is for the same phase but another conversion type.

**KON1** = 1 ; the first conversion happens on the ray path down.

- = 2 ; the first conversion happens on the ray path up.

**NDISK1** gives the discontinuity number for the first conversion: see the listing in the table in the *laufps-out* file.

**KON2** = 1 ; a possible second conversion happens on the ray path down.

- = 2 ; a possible second conversion happens on the ray path up.

**NDISK2** gives the discontinuity number for the second conversion.

```

-----example for a laufps-in file -----
00003
  0    0    0    0    0    0    Pdirect -->
  0    0    0    0    0    0    Sdirect -->
  1    2    3    0    0    0    P/PKP   -->   PKS
  1    2    2    0    0    0    P/PKP   -->   P/PKP at Moho to SV
  0                                     pP    -->
  0                                     PP    -->
  0                                     PPS   -->
  0                                     S/SKS -->
  0                                     sS    -->
  0                                     SS    -->
  2    2    3                         S3KP   -->
  1    2    3                         PcP    -->   PcS
  0                                     PcPScS -->
  2    2    3                         ScS    -->   ScP
-----end of example for a laufps-in file -----

```

The output from program **LAUFPS** can become very complex and long. However, the principle listing looks very like the output for **LAUFZE** and no example of an output file had been added here. For an example of a *laufps-out* file please apply the here listed *laufps-in* file and see also the files in the directory *examples/*.

# Chapter 12

## Intensity and Intensity Scales

(Version June 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch12)

Roger M.W. Musson<sup>1)</sup> and Ina Cčić<sup>2)</sup>

- <sup>1)</sup> British Geological Survey, West Mains Road, Edinburgh, EH9 3LA, UK;  
E-mail: [rmwm@bgs.ac.uk](mailto:rmwm@bgs.ac.uk), Phone: +44 (131) 650-0205, Fax: +44 (131) 667-1877
- <sup>2)</sup> Environmental Agency of the Republic of Slovenia, Urad za seizmologija i geologija,  
Dunajska 47/VII, SI-Ljubljana, Slovenia; Fax: +386 1 4787295, E-mail : [ina.cecic@gov.si](mailto:ina.cecic@gov.si)

|                                                                           | page |
|---------------------------------------------------------------------------|------|
| <b>12.1 Intensity and the history of intensity scales</b>                 | 2    |
| <b>12.2 Modern intensity scales</b>                                       | 3    |
| 12.2.1 European Macroseismic Scale (EMS)                                  | 4    |
| 12.2.2 Modified Mercalli (MM) Scale                                       | 7    |
| 12.2.3 Japan Meteorological Agency (JMA) scale                            | 9    |
| <b>12.3 Collection of macroseismic data</b>                               | 12   |
| 12.3.1 Macroseismic questionnaires, pre-internet                          | 12   |
| 12.3.2 Macroseismic questionnaires, post-internet                         | 14   |
| 12.3.3 Sample questionnaire                                               | 15   |
| 12.3.4 Field investigations                                               | 17   |
| 12.3.4.1 Preparation – before the earthquake occurs                       | 17   |
| 12.3.4.2 Preparation – after the earthquake occurs                        | 18   |
| 12.3.4.3 Equipment                                                        | 19   |
| 12.3.4.4 In the field                                                     | 20   |
| 12.3.4.5 Later (post-earthquake) activities                               | 22   |
| <b>12.4 Processing of macroseismic data</b>                               | 22   |
| 12.4.1 Traditional intensity assessment                                   | 22   |
| 12.4.2 Automatic intensity assessment                                     | 23   |
| 12.4.3 Accuracy assessment                                                | 26   |
| 12.4.4 Equivalence between scales                                         | 27   |
| 12.4.5 Presentation of intensity data                                     | 28   |
| 12.4.6 Exchange of macroseismic data                                      | 32   |
| <b>12.5 Determination of earthquake parameters from macroseismic data</b> | 32   |
| 12.5.1 Epicentral intensity                                               | 32   |
| 12.5.2 IDPs and isoseismals                                               | 33   |
| 12.5.3 Macroseismic epicenter                                             | 34   |
| 12.5.4 Macroseismic magnitude                                             | 35   |
| 12.5.5 Estimation of focal depth                                          | 36   |
| 12.5.6 Intensity attenuation                                              | 37   |
| <b>Acknowledgments</b>                                                    | 38   |
| <b>References</b>                                                         | 38   |

## 12.1 Intensity and the history of intensity scales

Intensity can be defined as a classification of the strength of shaking at any place during an earthquake, in terms of its observed effects. The fact that it is essentially a classification, akin to the Beaufort Scale of wind speed, rather than a physical parameter, leads to some special conditions on its use. Principal among these is its being an integer quantity when assigned from observed data. Traditionally, Roman numerals have been used to represent intensity values to emphasise this point (it is hard to write "VII½"). Nowadays the use of Roman numerals is largely a matter of taste, and most seismologists find Arabic numerals easier to process by computer.

The word “macroseismic” is used to denote those effects of an earthquake that can be determined without the use of instruments. This includes intensity data but is not restricted to it. A list of places with associated intensity values is macroseismic data; so is the information that an earthquake was felt at a place, or caused damage there. The phrase “intensity data point” (IDP) refers to a datum specifying latitude, longitude and intensity (and usually location name). One also encounters “macroseismic data point” (MDP) in which the numerical intensity may be replaced by some code to indicate “felt”, “damage”, “heavy damage”, etc. The study of the felt effects of an earthquake is termed *macroseismology*. A contour line enclosing places where the intensity was predominantly equal to or higher than a certain value is called an *isoseismal*.

The use of intensity scales is historically important because no instrumentation is necessary, and useful measurements of an earthquake can be made by an unequipped observer. The earliest recognisable use of intensity was by Egen in 1828, although simple quantifications of damage had been made in the previous century by Schiantarelli in 1783 (Sarconi 1784), and some earlier Italian examples are said to exist. However, it was only in the last quarter of the 19th century that the use of intensity became widespread; the first scale to be used internationally was the ten-degree Rossi-Forel Scale of 1883. The early history of intensity scales can be found in Davison (1900, 1921, 1933), a later study can be found in Medvedev (1962), and see also the discussion on the evolution of scales in Musson et al (2010).

The scale of Sieberg (1912,1923) became the foundation of all modern twelve-degree scales. A later version of it became known as the Mercalli-Cancani-Sieberg Scale, or MCS Scale (Sieberg 1932), still in use in Southern Europe. The 1923 version was translated into English by Wood and Neumann (1931), becoming the inappropriately named Modified Mercalli Scale (MM Scale). This was completely overhauled in 1956 by Richter (1958) who refrained from adding his name to the new version in case of further confusion with "Richter Scale" magnitudes. Richter's version became instead the "Modified Mercalli Scale of 1956" (MM56) despite the fact that the link to Mercalli was now extremely remote. Local modifications of Richter's MM56 scale have been used in Australia and New Zealand. Later attempts to radically modernise the MM scale, such as that of Brazee (1978), did not catch on. Stover and Coffman's (1993) version of the scale consists of modifications to Wood and Neumann's (1931) version, bypassing Richter (1958). This version is now the most widely used, thanks to it being the basis of the popular “Did You Feel It?” system maintained by the USGS.

In 1964 the first version of the MSK Scale was published by Medvedev, Sponheuer and Karník (Sponheuer and Karník 1964). This new scale was based on MCS, MM56 and previous work by Medvedev in Russia, and greatly developed the quantitative aspect to make the scale more powerful. This scale became widely used in Europe, and received minor modifications in the mid 1970s and in 1981 (Ad-hoc Panel 1981). In 1988 the European

Seismological Commission agreed to initiate a thorough revision of the MSK Scale. The result of this work (undertaken by a large international Working Group under the chairmanship of Gottfried Grünthal, Potsdam) was published in draft form in 1993, with the final version released (after a period of testing and revision) in 1998 (Grünthal 1998). Although this new scale is more or less compatible with the old MSK Scale, the organisation of it is so different that it was renamed the European Macroseismic Scale (EMS). Since its publication it has been widely adopted inside and also outside Europe.

The one important intensity scale that does *not* have twelve degrees (now that the Rossi-Forel Scale is no longer much in use) is the seven-degree Japanese Meteorological Agency Scale (JMA Scale). This is based on the work of Omori, and is the scale generally used in Japan (but nowhere else). A recent modification to the JMA scale subdivides degrees 5 and 6 into upper and lower, and explicitly describes a degree 0, resulting in a ten-degree scale (JMA 1996, 2009). Also, the scale used in Taiwan up to 2000 ran from degree 0 to degree 6, but now includes a degree 7, making eight degrees in total (Wu et al, 2003).

A point to be stressed is that intensity in the modern understanding of the word is a measure of the strength of shaking of an earthquake, not just a list of things that may accompany an earthquake. Thus, for instance, tsunami observations bear no relation to intensity. It may be that in places where the intensity is high, there is also a strong tsunami – but one can also observe strong tsunami effects at great distances where the earthquake is not even felt, and an earthquake that does not displace the seabed will not produce a tsunami no matter how strong the shaking. The same stricture applies to secondary effects due to surface faulting.

To some extent the middle years of the 20th century saw a decline in interest in macroseismic investigation, with the improvements in instrumental monitoring. However, since the middle 1970s there has been a revival of interest in the subject since macroseismic data are essential for the revision of historical seismicity and of great importance in seismic hazard assessments. Macroseismic studies of modern earthquakes are vital for (i) calibrating studies of historical earthquakes; (ii) studying local attenuation, and (iii) investigations of vulnerability, seismic hazard and seismic risk.

The use of the internet as a medium for gathering data has proved an extremely powerful tool, and in developed countries a vast amount of data can be gathered and processed very rapidly, giving immediate information on the impact of an earthquake; information that formerly would not be available after the processing of many paper questionnaires, a work that could occupy months. The possibility of showing, via the internet, the distribution of effects of an earthquake within a matter of minutes or hours, is an excellent opportunity for communicating science to the public, and has attracted a lot of interest. The wealth of macroseismic data now readily available has renewed interest in intensity as an analogue for physical ground motion parameters, especially velocity (e.g. Ebel and Wald 2003, Kaka and Atkinson 2004, Boatwright et al 2006, Atkinson and Wald 2007).

## **12.2 Modern intensity scales**

The three most important intensity scales in current use are the European Macroseismic Scale (EMS-98), the Modified Mercalli Scale (MM or MMI) and the JMA scale, although the latter is not widely used in practice any longer to assign intensities. Each is given in turn here.

### 12.2.1 European Macroseismic Scale (EMS)

The complete EMS-98 scale is too long to reproduce in its entirety, being a small book in length. This is because, while historically intensity scales have been presented simply as a list of classes and diagnostics for the user to make of what he will, the EMS-98 scale comes with extensive support material, including guidelines, illustrations and worked examples. Even the traditional "core" part of the scale contains tabular and graphical material explaining the classification of buildings and quantities used.

Some intensity scales in the past, such as the Modified Mercalli scale (in its 1956 incarnation, Richter 1958) have attempted to distinguish between the effects of earthquake shaking on buildings of different construction types, using type as an analogue of strength. An essential feature of the EMS is that it employs a series of six vulnerability classes which represent strength directly, and involve construction type, but also other factors such as workmanship and condition. These vulnerability classes allow a flexible and robust approach to assessing intensity from damage. The system is also adaptable to new or different building types, and includes consideration of engineered structures with earthquake resistant design. Damage is handled in a way that discriminates between structural and non-structural damage, and the different forms damage takes in buildings of different types. A system of five damage grades is used: negligible to slight; moderate; substantial to heavy; very heavy, and destruction. These are not only defined but also illustrated pictorially (Figure 12.1).

The probabilistic nature of intensity is stressed by the use of numerically-defined expressions of quantity. For any intensity degree, it is expected that for buildings of equivalent strength there will be a modal level of damage that will be most frequently encountered, and decreasing proportions of the building stock of equivalent strength will show lesser or greater degrees of damage. This relates closely to real experience from damage surveys.

Although natural phenomena such as landslips, rockfalls, cracks in ground, etc., have been used in intensity scales for a long time, more recent experience has shown that the occurrence of these is very strongly influenced by other factors than the severity of earthquake shaking - especially pre-existing hydrological conditions. In the EMS, although these effects are not deleted entirely, they are relegated to an Annexe rather than being included in the core scale; they are treated in a graphical table which shows the ranges of intensities over which such phenomena are commonly (and exceptionally) encountered.

The full scale is published as Grünthal (1998). It can also be obtained in full at the following WWW address: <http://tinyurl.com/cp7dc2>

| Classification of damage to masonry buildings                                       |                                                                                                                                                                                                                                                                                                 |
|-------------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|    | <b>Grade 1: Negligible to slight damage</b><br><b>(no structural damage, slight non-structural damage)</b><br>Hair-line cracks in very few walls.<br>Fall of small pieces of plaster only.<br>Fall of loose stones from upper parts of buildings in very few cases.                             |
|    | <b>Grade 2: Moderate damage</b><br><b>(slight structural damage, moderate non-structural damage)</b><br>Cracks in many walls.<br>Fall of fairly large pieces of plaster.<br>Partial collapse of chimneys.                                                                                       |
|   | <b>Grade 3: Substantial to heavy damage</b><br><b>(moderate structural damage, heavy non-structural damage)</b><br>Large and extensive cracks in most walls.<br>Roof tiles detach. Chimneys fracture at the roof line; failure of individual non-structural elements (partitions, gable walls). |
|  | <b>Grade 4: Very heavy damage</b><br><b>(heavy structural damage, very heavy non-structural damage)</b><br>Serious failure of walls; partial structural failure of roofs and floors.                                                                                                            |
|  | <b>Grade 5: Destruction</b><br><b>(very heavy structural damage)</b><br>Total or near total collapse.                                                                                                                                                                                           |

**Fig. 12.1** An example of the use of pictures in classifying damage grades in the European Macroseismic Scale

Despite the name of the scale (which reflects the fact that it was developed at the instigation of the European Seismological Commission), the scale is equally suitable for use outside of Europe, and has been used successfully for assessing modern earthquakes in many parts of the world. It is possible that the scale will be officially renamed in the near future to reflect this.



Here the short form (section 8 of the published scale) is reproduced. *This is not suitable for, and not intended for, use in assigning intensities.* It gives the character of each degree in a very simplified and generalised form for educational purposes.

### EMS Scale of 1998 (abstracted)

|             |                               |                                                                                                                                                                                                                                                                                                                                |
|-------------|-------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>I</b>    | <b>Not felt</b>               | Not felt.                                                                                                                                                                                                                                                                                                                      |
| <b>II</b>   | <b>Scarcely felt</b>          | Felt only by very few individual people at rest in houses.                                                                                                                                                                                                                                                                     |
| <b>III</b>  | <b>Weak</b>                   | Felt indoors by a few people. People at rest feel a swaying or light trembling.                                                                                                                                                                                                                                                |
| <b>IV</b>   | <b>Largely observed</b>       | Felt indoors by many people, outdoors by very few. A few people are awakened. Windows, doors and dishes rattle.                                                                                                                                                                                                                |
| <b>V</b>    | <b>Strong</b>                 | Felt indoors by most, outdoors by few. Many sleeping people awake. A few are frightened. Buildings tremble throughout. Hanging objects swing considerably. Small objects are shifted. Doors and windows swing open or shut.                                                                                                    |
| <b>VI</b>   | <b>Slightly damaging</b>      | Many people are frightened and run outdoors. Some objects fall. Some houses suffer slight non-structural damage like hair-line cracks and fall of small pieces of plaster.                                                                                                                                                     |
| <b>VII</b>  | <b>Damaging</b>               | Most people are frightened and run outdoors. Furniture is shifted and objects fall from shelves in large numbers. Many well built ordinary buildings suffer moderate damage: small cracks in walls, fall of plaster, parts of chimneys fall down; older buildings may show large cracks in walls and failure of fill-in walls. |
| <b>VIII</b> | <b>Heavily damaging</b>       | Many people find it difficult to stand. Many houses have large cracks in walls. A few well built ordinary buildings show serious failure of walls, while weak older structures may collapse.                                                                                                                                   |
| <b>IX</b>   | <b>Destructive</b>            | General panic. Many weak constructions collapse. Even well built ordinary buildings show very heavy damage: serious failure of walls and partial structural failure.                                                                                                                                                           |
| <b>X</b>    | <b>Very destructive</b>       | Many ordinary well built buildings collapse.                                                                                                                                                                                                                                                                                   |
| <b>XI</b>   | <b>Devastating</b>            | Most ordinary well built buildings collapse, even some with good earthquake resistant design are destroyed.                                                                                                                                                                                                                    |
| <b>XII</b>  | <b>Completely devastating</b> | Almost all buildings are destroyed.                                                                                                                                                                                                                                                                                            |

## 12.2.2 Modified Mercalli (MM) Scale

As stated, the situation regarding the MMI scale is confused. Most of the versions in circulation (e.g. on internet pages) are either the short form of the 1931 version, or some arrangement of the 1956 version. While papers frequently cite Stover and Coffman (1993) as the currently preferred version, one searches in vain for a copy of the scale according to Stover and Coffman (1993), since even Stover and Coffman do not present it in a unified form. Instead, they give Wood and Neumann's longer version, and a separate section in which they describe the modifications they apply in practice. Since this is the version of the scale that underlies the "Did You Feel It?" system (Wald et al 1999) and most of the recent studies on intensity and physical ground motion measures, the Stover and Coffman version is important, and it is strange that, hitherto, it seems to exist nowhere in print as an integrated scale.

Presented below is an attempt to give Stover and Coffman's 1993 version of MMI in one piece. This is not straightforward as it is not always clear precisely what Stover and Coffman's meaning is for some diagnostics, so a certain amount of editing has been necessary. Notable changes introduced by Stover and Coffman are that effects on people (other than loss of balance) are entirely removed from the scale above intensity 4. A number of other diagnostics are removed, and some are downgraded. The general result is that intensities assessed with this version are likely to be lower than before, especially in the range 4 to 7.

### MMI Scale of 1993 (Stover and Coffman, after Wood and Neumann, edited Musson)

- I** Not felt - or, rarely under especially favourable circumstances. Under certain conditions, at and outside the boundary of the area in which a great shock is felt: sometimes birds, animals, reported uneasy or disturbed; sometimes dizziness or nausea experienced; sometimes trees, structures, liquids, bodies of water, may sway - doors may swing very slowly.
- II** Felt indoors by few, especially on upper floors, or by sensitive, or nervous persons. Also, as in grade I, but often more noticeably: sometimes hanging objects may swing, especially when delicately suspended; sometimes trees, structures, liquids, bodies of water may sway; doors may swing very slowly; sometimes birds, animals reported uneasy or disturbed; sometimes dizziness or nausea experienced.
- III** Felt indoors by several, motion usually rapid vibration. Sometimes not recognized to be an earthquake at first. Duration estimated in some cases. Vibration like that due to passing of light, or lightly loaded trucks, or heavy trucks some distance away. Hanging objects may swing slightly. Movements may be appreciable on upper levels of tall structures. Rocked standing motor cars slightly.
- IV** Felt by many to all. Trees and bushes shaken slightly. Buildings shook moderately to strongly. Walls creaked loudly. Observer described the shaking as "strong." Awakened few, especially light sleepers. Frightened no one, unless apprehensive from previous experience. Vibration like that due to passing of heavy or heavily loaded trucks. Sensation like heavy body striking building or falling of heavy objects inside. Rattling of dishes, windows, doors; glassware and crockery clink and clash. Hanging objects

swung, in numerous instances. Disturbed liquids in open vessels slightly. Rocked standing motor cars noticeably.

- V** Buildings trembled throughout. Broke dishes, glassware, to some extent. Cracked windows - in some cases, but not generally. Overturned vases, small or unstable objects, in many instances, with occasional fall. Hanging pictures fall. Opened, or closed, doors, shutters, abruptly. Pendulum clocks stopped, started or ran fast, or slow. Moved small objects, furnishings, the latter to slight extent. Trees, bushes, shaken moderately to strongly. People have difficulty standing or walking. Felt moderately by people in moving vehicles.
- VI** Damage slight in poorly built buildings. Fall of plaster in small amount. Cracked plaster somewhat, especially fine cracks in chimneys in some instances. Broke dishes, glassware, in considerable quantity, also some windows. Fall of knickknacks, books, pictures. Overturned furniture in many instances. Moved furnishings of moderately heavy kind. Small bells rang - church, chapel, school, etc. The intensity can only be assessed as VI if damage to buildings is observed, unless many small objects fall from shelves or many glasses or dishes are broken.
- VII** Damage negligible in buildings of good design and construction, slight to moderate in well-built ordinary buildings, considerable in poorly built or badly designed buildings, adobe houses, old walls (especially where laid up without mortar), spires, etc. Cracked chimneys to considerable extent, walls to some extent. Fall of plaster in considerable to large amount, also some stucco. Broke numerous windows, furniture to some extent. Shook down loosened brickwork and tiles. Broke weak chimneys at the roofline (sometimes damaging roofs). Fall of cornices from towers and high buildings. Dislodged bricks and stones.
- VIII** Damage slight in structures (brick) built especially to withstand earthquakes. Considerable in ordinary substantial buildings, partial collapse: racked, tumbled down, wooden houses in some cases; threw out panel walls in frame structures, broke off decayed piling. Fall of walls. Cracked, broke, solid stone walls seriously. Twisting, fall, of chimneys, columns, monuments, also factory stacks, towers. Moved conspicuously, overturned, very heavy furniture. Trees shaken strongly - branches, trunks, broken off, especially palm trees.
- IX** Damage considerable in (masonry) structures built especially to withstand earthquakes: threw out of plumb some wood-frame houses built especially to withstand earthquakes; great in substantial (masonry) buildings, some collapse in large part; or wholly shifted frame buildings off foundations, racked frames.
- X** Damage serious to dams, dykes, embankments. Severe to well-built wooden structures and bridges, some destroyed. Developed dangerous cracks in excellent brick walls. Destroyed most masonry and frame structures, also their foundations.
- XI** Damage severe to wood-frame structures, especially near shock centres. Great to dams, dikes, embankments often for long distances. Few, if any (masonry) structures remained standing. Destroyed large well-built bridges by the wrecking of supporting piers, or pillars. Affected yielding wooden bridges less.
- XII** Damage total - practically all works of construction damaged greatly or destroyed.

**Notes:** It is ambiguous in the text of Stover and Coffman (1993) whether the section of diagnostics for intensity 4 beginning “Awakened few ...” and ending “Rocked standing motor cars noticeably” is intended to be included in the scale or not. USGS practice is to include them (Dewey 2009 *pers. comm.*). According to some sources, e.g. Boatwright and Bundon (2005), the revisions of Stover and Coffman (1993) include treating the fall of few chimneys as intensity 6 and the fall of half of all chimneys or more as 7. This is not actually found anywhere in Stover and Coffman (1993).

### 12.2.3 Japan Meteorological Agency (JMA) Scale

The JMA scale has evolved over the years from the seven-degree scale of Omori (1900). The revision of 1996 added two new degrees, but to try and maintain consistency with older data sets, the new scale divided degree 5 into “5 Lower” and “5 Upper”, and similarly for degree 6. The latest revision in 2009 maintains this system, but the scale is reworded and revised throughout, and is generally greatly improved in clarity and cohesiveness.

However, routine practice in Japan seems to be largely to rely on estimates of intensity made by customised strong-motion instruments. The scale is therefore a guide to what might be expected to happen in a place where the local intensity instrument registers a given value; it is not a means of recording what actually was observed. Thus the scale notes that at intensities above 4, railway services may be suspended for safety checks. No-one would use that information to assign intensity, but it is useful information for people affected by an earthquake.

#### JMA Scale of 2009 (re-arranged)

|                  |                                                                                                             |
|------------------|-------------------------------------------------------------------------------------------------------------|
| <b>0</b>         | <i>Effects on</i>                                                                                           |
| <i>People:</i>   | Imperceptible to people.                                                                                    |
| <b>1</b>         | <i>Effects on</i>                                                                                           |
| <i>People:</i>   | Felt slightly by some people keeping quiet in buildings.                                                    |
| <b>2</b>         | <i>Effects on</i>                                                                                           |
| <i>People:</i>   | Felt by many people keeping quiet in buildings. Some people may be awoken.                                  |
| <i>Indoors:</i>  | Hanging objects such as lamps swing slightly.                                                               |
| <b>3</b>         | <i>Effects on</i>                                                                                           |
| <i>People:</i>   | Felt by most people in buildings. Felt by some people walking. Many people are awoken.                      |
| <i>Indoors:</i>  | Dishes in cupboards may rattle.                                                                             |
| <i>Outdoors:</i> | Electric power lines swing slightly.                                                                        |
| <b>4</b>         | <i>Effects on</i>                                                                                           |
| <i>People:</i>   | Most people are startled. Felt by most people walking. Most people are awoken.                              |
| <i>Indoors:</i>  | Hanging objects such as lamps swing significantly, dishes in cupboards rattle. Unstable ornaments may fall. |

*Outdoors:* Electric wires swing significantly. Those driving vehicles may notice the tremor.

**5 Lower** *Effects on*

*People:* Many people are frightened and feel need to hold onto something stable.

*Indoors:* Hanging objects such as lamps swing violently. Dishes in cupboards and items on bookshelves may fall. Many unstable ornaments fall. Unsecured furniture may move, and unstable furniture may topple over.

*Outdoors:* In some cases, windows may break and fall. People notice electricity poles moving. Roads may sustain damage.

*Wooden houses (low earthquake resistance):* Slight cracks may form in walls.

*Ground and slopes:* Small cracks in ground may form and liquefaction may occur. Liquefaction may be seen in areas with a high groundwater level and loose sand deposits. Damage observed as a result of liquefaction includes spouts of muddy water from the ground, outbreaks of subsidence in riverbanks and quays, elevation of sewage pipes and manholes, and leaning or destruction of building foundations. Rock falls and landslips may occur.

*Utilities and infrastructure:* In the event of shaking with a seismic intensity of about 5 Lower or more, gas meter with safety devices are tripped, stopping the supply of gas. In the event of stronger shaking, the gas supply may stop for entire local blocks. Suspension of water supply and electrical blackouts may occur in regions experiencing shaking with a seismic intensity of about 5 Lower or more. In the event of shaking with a seismic intensity of about 5 Lower or more, elevators with earthquake control devices will stop automatically for safety reasons.

**5 Upper** *Effects on*

*People:* Many people find it hard to move; walking is difficult without holding onto something stable.

*Indoors:* Dishes in cupboards and items on bookshelves are more likely to fall. TVs may fall from their stands, and unsecured furniture may topple over.

*Outdoors:* Windows may break and fall, unreinforced concrete-block walls may collapse, poorly installed vending machines may topple over, automobiles may stop due to the difficulty of continued movement.

*Wooden houses (low earthquake resistance):* Cracks may form in walls.

*Reinforced concrete buildings (low earthquake resistance):* Cracks may form in walls, crossbeams and pillars.

*Ground and slopes:* As 5 Lower.

**6 Lower** *Effects on*

*People:* It is difficult to remain standing.

*Indoors:* Many unsecured furniture moves and may topple over. Doors may become wedged shut.

*Outdoors:* Wall tiles and windows may sustain damage and fall.

*Wooden houses (low earthquake resistance):* Cracks are more likely to form in walls. Large cracks may form in walls. Tiles may fall, and buildings may lean or collapse.

*Wooden houses (high earthquake resistance):* Slight cracks may form in walls.

*Reinforced concrete buildings (low earthquake resistance):* Cracks are more likely to form in walls, crossbeams and pillars.

*Reinforced concrete buildings (high earthquake resistance):* Cracks may form in walls, crossbeams and pillars.

*Ground and slopes:* Cracks in ground may form. Landslips and landslides may occur.

## **6 Upper**

### *Effects on*

*People:* It is impossible to remain standing or move without crawling. People may be thrown through the air.

*Indoors:* Most unsecured furniture moves, and is more likely to topple over.

*Outdoors:* Wall tiles and windows are more likely to break and fall. Most unreinforced concrete-block walls collapse.

*Wooden houses (low earthquake resistance):* Large cracks are more likely to form in walls. Buildings are more likely to lean or collapse.

*Wooden houses (high earthquake resistance):* Cracks may form in walls.

*Reinforced concrete buildings (low earthquake resistance):* Slippage and X-shaped cracks may be seen in walls, crossbeams and pillars. Pillars at ground level or on intermediate floors may disintegrate, and buildings may collapse.

*Reinforced concrete buildings (high earthquake resistance):* Cracks are more likely to form in walls, crossbeams and pillars.

*Ground and slopes:* Large cracks in ground may form. Landslips are more likely to occur; large landslides and massif collapses may be seen. When large landslides and massif collapses occur, dams may form depending on geographical features, and debris flow may occur due to the large quantities of sediment produced.

*Utilities and infrastructure:* In the event of shaking with a seismic intensity of 6 Upper or more, gas, water and electric supplies may stop over wide areas.

## **7**

### *Effects on*

*People:* It is impossible to remain standing or move without crawling. People may be thrown through the air.

*Indoors:* Most unsecured furniture moves and topples over, or may even be thrown through the air.

*Outdoors:* Wall tiles and windows are even more likely to break and fall. Reinforced concrete-block walls may collapse.

*Wooden houses (low earthquake resistance):* Buildings are even more likely to lean or collapse.

*Wooden houses (high earthquake resistance):* Cracks are more likely to form in walls. Buildings may lean in some cases.

*Reinforced concrete buildings (low earthquake resistance):* Slippage and X-shaped cracks are more likely to be seen in walls, crossbeams and pillars. Pillars at ground level or on intermediate floors are more likely to disintegrate, and buildings are more likely to collapse.

*Reinforced concrete buildings (high earthquake resistance):* Cracks are even more likely to form in walls, crossbeams and pillars. Ground level or intermediate floors may sustain significant damage. Buildings may lean in some cases.

*Basic infrastructure:* Electrical, gas, and water supplies are interrupted over a large area.

*Ground and slopes:* As 6 Upper.

## **Notes:**

Wooden houses and RC buildings are classified into two categories according to their earthquake resistance, which tends to be higher for newer structures. Earthquake resistance tends to be low for structures built up to 1981, and high for those built since 1982. However,

to maintain a certain range of earthquake resistance according to differences in structure and wall arrangement, resistance is not necessarily determined only by building age. The earthquake resistance of existing buildings can be ascertained through quakeproofing diagnosis.

The walls of wooden houses are assumed to be made of mud and/or mortar. Mortar in a wall with a weak base can easily break off and fall, even under conditions of low deformation. Damage to wooden houses depends on the period and duration of seismic waves. In some cases (such as the Iwate-Miyagi Nairiku Earthquake of 2008), few buildings sustain damage in relation to the level of seismic intensity observed.

## **12.3 Collection of macroseismic data**

Collection of macroseismic data from current earthquakes is derived principally from two sources - questionnaire surveys and field investigations, either or both of which may be required for a particular earthquake. As a general rule, questionnaire surveys are used for assessing intensities in the range of 2 to 6, while for 7 and above field investigations are necessary. Questionnaires used to be distributed predominantly as paper documents, to be filled in and returned to the investigator. Since about 2000, this has been progressively replaced by internet collection of data.

One thing in common with both questionnaire surveys and field investigations is the desirability of rapid response - evidence of earthquake damage is patched up within days or even hours, and human memory of details (and interest in the subject) also wanes rapidly. There is a third source - documentary material, for instance, newspaper accounts - which is the principal source of macroseismic data for historical earthquakes. The treatment of this is a separate subject involving the techniques of the professional historian as well as the seismologist, and is not dealt with here.

A general point common to both paper and internet questionnaires is the problem of geo-location. One cannot ask people to give their location in terms of latitude and longitude, and interpreting addresses from a large number of questionnaires would be too onerous. So the usual practice is to rely on postcodes (zip codes). The problem here is that different countries use different systems with different resolutions, so in one country the postcode may be convertible to a co-ordinate location with accuracy of tens of metres, while in another, the accuracy may not be better than tens of kilometres.

In either case, someone who fills in a questionnaire for an earthquake he felt other than at home, may not know the postcode of the location where he was – and may simply put in his home postcode, a potential source of error in the data.

### **12.3.1 Macroseismic questionnaires, pre-internet**

When dealing with traditional macroseismology, when surveys of earthquake effects had to be conducted using paper forms and the postal service, one could distinguish two basic types of macroseismic questionnaire, dependant on the intended recipient. The first is the questionnaire to be answered by an individual citizen recounting his personal experiences of the earthquake. The second is the questionnaire designed to be answered by someone with knowledge of the experiences of the entire community. Which of these two approaches is

used will shape the macroseismic investigation as a whole; the choice may well be forced on the investigator by circumstances. For example, in some countries there will be found an official in each town or rural community whose job includes completing such requests for data, and who can be relied on to fill in any questionnaire submitted. In some other countries such officials are not to be found, and this means of investigation is therefore not possible. Some institutes may have the resources to post out thousands of questionnaires as mailing shots; others may not be able to afford such a technique.

One may discern four basic types of person who may fill in a macroseismic questionnaire, two in each of the classes outlined above.

- (i) *The unselected individual* - questionnaires may be distributed haphazardly and in great bulk by publication in newspapers, dissemination at libraries, etc. This guarantees a large response, but probably biases the results in favour of positive responses.
- (ii) *The randomly selected individual* - there exists a methodology, highly developed in the social sciences, for disseminating questionnaires in such a way as to maximise the statistical validity of the results, using random selection procedures based on electoral rolls and direct mailing, often with some incentive to return the questionnaire (such as a prize draw). This is the best method in terms of the reliability of the results, since a random sample enables one to make statistically valid estimations of the characteristics of the whole population. The drawbacks are that such a response may be difficult to organise rapidly after an earthquake, and is likely to be relatively expensive. It should not be forgotten, that the art of questionnaire design and methodology has been studied in detail by social scientists for many years, and the expertise accumulated should not be ignored by seismologists whose background usually lies in the physical sciences.
- (iii) *The public official* - it is very convenient to be able to send a single questionnaire to the local burgomaster/postal officer/police superintendent's office and have it filled in with the details of the effects of the earthquake in the whole of the community under the official's jurisdiction. What the seismologist cannot be certain of is how conscientiously the questionnaire is filled in. Does the official make detailed enquiries, or does he jot down the first thing that comes into his head?
- (iv) *The volunteer* - some seismological institutes arranged networks of local volunteers with some standing in the community (schoolteachers, clergymen) and enthusiasm for the task of supplying useful data. Such volunteers can be given a stack of blank questionnaires in advance and can be relied upon to fill one in after an earthquake occurs with dependable data on the effects in the locality. Such a system is very effective, but can be laborious to set up and maintain. An advantage of such a system is that one stands a good chance of getting reliable negative information.

A further division of questionnaire design is that between the free-form questionnaire and the multiple choice style. The first style gives open-ended questions to which the respondent can answer in his own words ("What sort of shaking did you experience?") while the second gives a series of boxes to tick ("The shaking was A - weak; B - moderate; C - strong"). The second style is easier to process, but runs the risk of losing information that doesn't easily fit the predefined categories. A combination of both styles is also possible.

Length of questionnaire is also important. Too long or difficult a questionnaire will discourage people from filling it in, as will asking questions that are too hard for most people



to answer - for example, how many people can accurately describe their local geology? One should guard against asking questions that are not strictly necessary (such as personal details). From the above discussion it will be seen that questionnaire design is somewhat of an art, and that what will work for one country won't work for another.

A sample of a typical paper questionnaire for distribution after an earthquake is given in section 12.3.3 below.

### **12.3.2 Macroseismic questionnaires, post-internet**

The use of the internet as a tool for collecting macroseismic data has transformed the subject over the first decade of the 21<sup>st</sup> century, at least in developed countries where the use of the internet in ordinary homes is widespread. Not only can data be collected in very large volumes very rapidly, it can also be processed in real time and the results displayed immediately. The adoption of such a system answers definitively the question discussed in the previous section as to who is expected to provide the response – internet questionnaires will be completed by ordinary persons, who visit the institute's web site and choose to record their experience. The internet questionnaire will therefore be designed to capture this type of individual experience, from persons lacking technical expertise. It does mean that the sample is always self-selecting, and therefore will tend to be biased in favour of positive responses. Automatic processing routines favour the collection of data where answers are predefined between several options, so a check-box approach will usually be most appropriate, though space should be provided to allow for extra comments to be written in; these will need to be read individually by the investigator.

It was difficult in the past to produce a standard macroseismic questionnaire that could be used internationally, because of the different types of respondent in different countries. However, since all internet questionnaires are directed at ordinary individuals, there is a real possibility that a single standard questionnaire could be developed. Indeed, both the USGS and EMSC have collected data for earthquakes in many different countries, according to standard questionnaire formats.

The main obstacle to a single unified questionnaire is that internet questionnaires are to some extent shaped by the intensity scale in use and the system that will be used to process the data. This is likely to vary from country to country still, due to a desire to maintain compatibility with past data sets. The establishment of a global standard questionnaire, from which national institutes can select subsets of questions as desired, is still a realisable goal. Currently, a draft standard internet questionnaire has been drawn up by the ESC Working Group on Internet Macroseismology.

Examples of online questionnaires can be seen on the USGS (<http://earthquake.usgs.gov/earthquakes/dyfi/>) and EMSC (<http://www.emsc-csem.org/Earthquake/felt.php>) websites, among others.

### 12.3.3 Sample questionnaire

#### EARTHQUAKE QUESTIONNAIRE

The following questionnaire is part of a study of the effects of the \_\_\_\_\_ (name) earthquake, which occurred on \_\_\_\_\_ (date) at \_\_\_\_\_ (time). You are invited to use it to record what you experienced.

If you did NOT feel the earthquake, or notice it at all, please tick here [ ] and complete questions 1, 2 and 5, below. This information will still be useful for our study. Please send completed questionnaires to this address:

.....  
.....  
.....

#### SECTION A – WHERE YOU WERE

1. At the time of the earthquake, where were you?

Address (including post code).....

Outdoors [ ]                      Ground floor [ ]                      Upper floor [ ] ; If so, which floor?

Stationary vehicle [ ]                      Moving vehicle [ ]                      Other \_\_\_\_\_

If indoors, please describe the type of building:

Function (house, school, church, etc) .....

Height (number of stories) .....

Construction (brick, stone, wood, etc) .....

2. What were you doing?

Walking [ ]                      Standing [ ]                      Sitting [ ]                      Kneeling [ ]

Lying down [ ]                      Sleeping [ ]

#### SECTION B – EARTHQUAKE SHAKING AND SOUND

3. What best describes the shaking you felt?

No shaking [ ]                      Trembling [ ]                      Swaying [ ]                      Jerky motion [ ]

Impact [ ]                      Rolling motion [ ]                      Other [ ] .....

It was ...    Weak [ ]                      Moderate [ ]                      Severe [ ]

4. What best describes any sound you heard?

No sound [ ]                      Rumbling [ ]                      Roaring [ ]                      Explosion [ ]

Other [ ] .....

It was ...    Faint [ ]                      Moderate [ ]                      Loud [ ]

#### SECTION C – EFFECTS ON PEOPLE AND ANIMALS

5. Which best describes what happened where you were (your house, neighbours)?

Nobody noticed it [ ]                      Only one or two people noticed it [ ]  
 Some people noticed it, but not many [ ] Many people noticed it [ ]  
 Most people noticed it [ ] Everyone noticed it [ ]  
 People indoors noticed it, but not those outside [ ]  
 People upstairs noticed it, but not those on the ground floor [ ]  
 I don't know whether other people noticed it or not [ ]

6. (Only for earthquakes that happened at night) Did the earthquake wake you?  
 No [ ]      Yes [ ]              I wasn't asleep [ ]  
 Were other people where you were woken up?  
 No [ ]      Yes, a few [ ]    Yes, many [ ]    Yes, most/all [ ]                      Don't know [ ]
7. Were you frightened?  
 No [ ]      Yes [ ]  
 Where you were, did anybody run outdoors in fright?  
 No [ ]      Yes, a few [ ]    Yes, many [ ]    Yes, most/all [ ]                      Don't know [ ]
8. Was it hard to stand or walk?  
 No [ ]      Yes, a bit [ ]    Yes, very [ ]

#### SECTION D – EFFECTS ON OBJECTS, BUILDINGS, ETC

9. Did any of the following things happen?
- |                                  | Yes | No  | Don't know |
|----------------------------------|-----|-----|------------|
| Windows/doors rattled            | [ ] | [ ] | [ ]        |
| Crockery, etc rattled            | [ ] | [ ] | [ ]        |
| Hanging objects swung            | [ ] | [ ] | [ ]        |
| Pictures moved askew or fell     | [ ] | [ ] | [ ]        |
| Small objects shifted or fell    | [ ] | [ ] | [ ]        |
| Books or similar shifted or fell | [ ] | [ ] | [ ]        |
| Furniture shook visibly          | [ ] | [ ] | [ ]        |
| Furniture shifted out of place   | [ ] | [ ] | [ ]        |
| Furniture toppled over           | [ ] | [ ] | [ ]        |
| Liquids splashed or spilled      | [ ] | [ ] | [ ]        |
- Please give details, or note any other things that you noticed:.....  
 .....
10. Was there any damage to buildings where you were?  
 No [ ]      Yes [ ]              Don't know [ ]  
 If yes, please describe the damage .....  
 .....  
 .....
11. Were there any effects on natural surroundings where you were, for example, landslips, cracks in ground, effects on ponds or streams, etc?  
 No [ ]      Yes [ ]              Don't know [ ]  
 If yes, please describe the effects .....  
 .....  
 .....  
 .....
12. Have you any other comments about the effects of the earthquake that might be useful?

### **12.3.4 Field investigations**

Earthquakes can't be predicted; but one can know, in any region, the events that are likely to occur at some time or other, and one can at least partially prepare in advance. As always in macroseismology, there is no fixed rule how to do things, because how one proceeds will heavily depend on the size of the earthquake, its location and the importance of the data for the overall knowledge of the seismicity of the area. For example, an earthquake of intensity 6 will not surprise anyone in Italy or Greece, and will probably be surveyed only briefly, but such event in Finland or Denmark might be one of the strongest felt in a very long period, and therefore would be studied very much in detail.

There are quite a few distinctions in procedure in dealing with an event with only minor (non-structural) damage and a more severely damaging one.

Field work in macroseismology in general can be divided in three main phases: preparation, the field activities themselves, and finally later activities. The following issues need to be discussed: arranging teams of investigators, providing equipment, arranging travel and accommodation, methods of investigation, and ensuring safe working practices.

#### **12.3.4.1 Preparation – before the earthquake occurs**

Some preparation should be done on a contingency basis, before any earthquake has occurred; this minimises the amount of time needed arranging things between the occurrence of the earthquake and the setting out of the field investigation team.

The first issue is building a roster of potential participants in field investigations. These may be observatory staff (seismologists) but may also include other seismologists, either from other institutes in the same country or from abroad (especially neighbouring countries; cross-border earthquakes may need to be considered). In case of an earthquake strong enough to cause structural damage, it is essential to be able to call on the help of one or more civil engineers with experience in earthquake engineering. The help of a geotechnical engineer or engineering geologist will also be needed in cases of high-intensity earthquakes. In cases that it is expected that international team members will be called in who do not speak the local language, suitable translators should be available. Names and phone numbers of potential team members should be compiled in advance.

In some cases it may be useful to include team members with less experience, for training purposes, though this is not recommended for earthquakes that are humanitarian disasters.

Equipment (discussed in detail below) needs to be collected in advance and stored for use when needed, and checked at regular intervals to make sure that things like batteries are in a state of readiness.

Procedures for arranging travel can also be made in advance, for fast issuing of tickets, travel insurance and other travel documents. If the observatory does not have its own vehicles, arrangements should be in place for quick availability of rented vehicles (and in extreme cases of very strong earthquakes, a helicopter will probably be needed).

Additional items to consider: a list of passport/identity cards/driving licenses expiry dates of all team members, to be checked regularly so that the documents can be updated in time; a list of passwords for remote connections, for laptops, and pin numbers of mobile phones.

#### **12.3.4.2 Preparation – after the earthquake occurs**

Immediately after the earthquake has occurred, the actual field investigation can be prepared, in the light of the severity of the event and its actual location. In the case of earthquakes causing only minor damage, a quick arrival into the field is essential, as damage will swiftly be made good and debris swept away. In the case of major earthquakes causing severe damage and many casualties, rapid response is less essential (ruined buildings will not be cleared so quickly) and may even be a bad idea, as priority must be given to rescue work. Scientific investigation should not be allowed to hinder or compromise humanitarian relief.

For an earthquake of intensity up to 6 it is usually enough that two people spend a few days in the field. If the earthquake was felt in a large town, an additional day spent collecting data in schools can be useful. Of course, it depends how important the earthquake is for the knowledge of the seismicity of the region. In countries with low seismicity such earthquake would definitely justify more work than elsewhere. For a severe earthquake, one should expect that around six people (two groups of three) should work for four or five days, though this can vary according to the situation and resources.

Travel arrangements vary very much depending on the size of the event, the level of destruction and the distance from the epicentral area to the starting point of the team. In the majority of cases, a car or several cars will be enough. In case of a major earthquake, one might consider aerial survey (plane, helicopter) to get an idea of the situation in the most damaged area. Public transport might be not available in the affected area, so one cannot rely on it. If the nearby airfields are closed, the teams might need to use humanitarian or rescue flights; for this, it would be advisable to establish contacts in advance and keep the contact data updated.

Regarding accommodation, a centrally placed hotel is ideal when dealing with relatively minor earthquakes, but in cases of major events with strong aftershock sequences, one should choose a hotel well away from the epicentral area. In extreme cases, for instance if the damaged area is difficult to access by road, or the area of destruction is very large, it may be necessary to camp in tents.

When dealing with a strongly damaging earthquake it will be essential to liaise with the civil authorities in advance of arriving in the damage zone, to arrange authorisation and access. After the 2010 Haiti earthquake, remote sensing data was made available rapidly, from which it was possible to identify the worst damaged areas, and locations of landslides and other major ground failures. This sort of data is likely to continue to be available in the future, at least for major disasters, and may be extremely useful in planning locations to visit.

Lastly, point of contact arrangements should be made with observatory staff back at base, with the team making contact regularly (usually once a day, in the evening) to report that all are safe.

#### 12.3.4.3 Equipment

The equipment that will be needed can be divided into two: items that are always needed, and items needed when investigating areas with structural damage.

The basic items are the following:

- Maps, road navigation devices/GPS, cameras, extra sets of batteries, laptop, notebooks, portable voice recorder, tape measure, copy of the intensity scale.
- Optionally, one may wish to carry printed questionnaires (especially if one intends to visit some schools) when dealing with earthquakes with limited damage. Also educational material (especially guidelines to earthquake safety) could be carried, and again, this is particularly useful in case of visits to schools.
- Clothes and shoes should be chosen depending on the weather conditions and the type of terrain. If the investigation will last several days, and conditions are likely to be difficult (e.g. weather is bad) then special care should be taken in choosing clothing. Clothes should be comfortable to wear and not complicated to take care of. Sturdy shoes, materials, that behave well in rain (jeans can be impractical when wet), jackets, hats and mittens, and multilayered clothing is recommended. Basics like sunglasses, insect repellent, sunscreen should not be forgotten.
- When in a damaged area, safety clothing is essential. This means a helmet, high-visibility jacket, and steel-tipped boots. The jacket should preferably be marked clearly with the name of the institute or field investigation team. It can be very difficult to access the epicentral area without very visible identification (Figure 12.2).



**Fig. 12.2** Italian seismologists making a macroseismic field investigation after the 2009 L'Aquila earthquake: note the safety helmets and high-visibility jackets clearly marked with the institute name.

- In general, clothes with the institute logo are useful, and IDs with photographs should be carried at all times. Residents are sometimes suspicious of people going round photographing houses.
- In case of a high intensity earthquake it might be necessary for a team to use a tent; so camping equipment needs to be available, and regularly checked/renewed as well: tents, sleeping bags, cooking equipment, etc. Lamps, flashlights, batteries and first aid kits should be ready, as well as dry food and water for a few days. Disaster zones can be insanitary, so water purification, surgical masks, and gloves may be needed.

#### **12.3.4.4 In the field**

In dealing with a relatively low-intensity event, a visit to epicentral area should include a short stay in each (or at least in a representative number) of the damaged localities. Each locality should be surveyed on foot (if small) or by car, to get the overview of the situation and to note the extreme cases of damage. Contact with local authorities is recommended; information about damage can be found from the emergency services, insurance agencies and just talking to people. Schools are a good source of information – if there are enough team members, one can spend a few hours visiting the local school, especially if it is in a rural setting, where children are being brought from a wider area. In villages it is often enough to stop in the shop or in the pub, these being the points where many people come and talk to each other. One can consider leaving some printed questionnaires with some person willing to conduct a small survey and send back the answers, or at a local library or post office.

The team will usually split up into pairs or trios, one pair per car. While one drives, the other one can deal with the collected information, study maps, take phone calls etc. In the ideal case the pairing should consist of one seismologist and one civil engineer. In case of geotechnical issues (liquefaction, etc), an engineering geologist or geotechnical engineer should accompany them. In practice, a team member can be anyone working in seismology, with working knowledge of the scale and some basic training in recognising the building type and grade of damage. For low intensity earthquakes, it is not expected that there would be language barriers between the team and the inhabitants of the epicentral area, as the investigation would be made only by members of local institutes; in case of language differences, it is recommendable to have an interpreter along as well. For large earthquakes, the size of the team can vary during the time of the survey, as more participants can arrive later, if necessary, and the team members may be changed according to circumstances.

As a rule, if all the damage is non-structural, a team consisting only of seismologists is sufficient. Once structural damage needs to be registered, the assistance of a civil engineer becomes essential.

In the field, it is necessary to combine both detailed and general surveys of structural behaviour. Structures need to be surveyed in terms of: the distribution of different types; the overall vulnerability (resistance or lack of resistance to earthquake shaking) of typical structures while noting deviations in terms of good or bad examples; and the distribution of different grades of damage within each building type. Care should be taken over making accurate records of the location of all structures studied or photographed. Data should be gathered as written notes and photographs.

For engineered structures, a detailed study should be carried out to identify both good and bad performance in a sample of both damaged and undamaged structures. Both external and internal damage should be recorded if safe to do so (bearing in mind that entering damaged buildings may be too dangerous), identifying typical modes of failure. In order to be able to relate the damage to the intensity scale, information on the strength of the building is required: strengths and weaknesses in the construction techniques, special points of vulnerability, or high resistance, irregularity or symmetry in the building design, the quality of the materials used, and so on. It is a good idea to collect information on what earthquake-resistant design regulations were in force before the earthquake, and also, where possible, to investigate to what extent these regulations were followed in the buildings examined.

The case is similar for non-engineered structures; these are likely to be less individual, so the task becomes one of identifying the main characteristic structural forms, their age and condition. Again, the extent and types of damage, both interior and exterior, need to be recorded.

Detailed photographic surveys can be made of individual streets or districts to record the percentages of various types of buildings that were damaged to a lesser or greater degree. These surveys should be supplemented with internal records from at least a sample of the buildings examined.

The overall spatial distribution of damage can be recorded over a large area by the use of general surveys employing proper sampling techniques to generate statistically consistent data. Distinctions between different construction types, usage, height and age and quality of construction should always be made wherever possible.

Geotechnical aspects should also be investigated. Any relationship between local geology and damage distribution should be investigated. (This does not entail “correcting” intensities for local conditions, but does explain local variations in observed intensity). Data should be gathered on groundwater and hydrological conditions before the earthquake. The following topics also need to be considered: types of foundation and their performance; effects on embankments, cuttings and river banks; liquefaction and other ground effects like cracking; landslides and rockfalls. Negative data as well as positive data should be collected.

Special studies may be needed of individual industrial or civil facilities. Effects on factories can include damage to pipework and ducting, pumps and valves, cabling systems, tanks, machinery, electrical controls, computers and cranes. The effects on dams, bridges, port facilities, tunnels and irrigation systems should be recorded. The effects on lifelines (services, transport) also merit attention: underground provision of water, gas, electricity and telecommunications; railways, roads etc. These sorts of data are not generally suitable for intensity assessment per se, but are important to record, particularly when making an assessment of the economic impact of the earthquake, or looking at lessons to be learnt from an engineering perspective.

One should not forget that for intensity evaluation the data be statistically representative, and in larger settlements this is difficult to achieve. Therefore, the teams should not lose too much time on a single building, but try to optimize the time. Teams of engineers will study building by building, and their data can be used to check some cases, if necessary.

When dealing with a small settlement (village or a fraction of the village) it is more or less simple to go around and see all the buildings. When dealing with a larger settlement, it is



reasonable to divide it into parts and assign several teams to do the job block by block. All the collected data should be geo-referenced in situ, if possible; backup copies of photographs, videos and files produced during the field work should be made at least once a day. To be safe, it is good to have the ability to email or ftp the data regularly to another location, to prevent losses.

While entering areas with high intensity effects, team members must wear safety gear and be aware of the danger. In many cases a special permit should be obtained from the civil authorities, who then assign specialists (firemen, policemen, soldiers etc) to accompany teams and take care of safety issues. One should not underestimate the danger of aftershocks and additional collapses they might cause. The safety of the team members comes above the data. In case of two or more field investigation teams working in the same area, one should as much as possible liaise with these, with a view to avoiding duplication of effort and sharing experience.

As a general rule, one should also be careful to maintain good relations with the local community in the area being investigated. In areas that have suffered a major earthquake, residents may still be traumatised or shocked, and one needs to deal with them with tact and care. Even in cases of low-intensity earthquakes, one should take care to respect local cultural conventions (e.g. with respect to appropriate clothing).

#### **12.3.4.5 Later (post-earthquake) activities**

On return to base team members should take care to label carefully and archive all the collected data; all photographs, videos and other material should have in titles or captions clear data about the place where they were taken, names of the streets, buildings etc. All the data should be copied several times and stored in different archives, to prevent losses. It may be appropriate to continue liaison with other groups involved with the earthquake with a view to pooling and sharing data and experience.

### **12.4 Processing of macroseismic data**

The way in which intensity is assessed varies enormously according to whether one is proceeding from traditional paper questionnaires, in which case the process is essentially a subjective matching of the data to descriptions in the scale, or whether one is using an automatic algorithm to process data from the internet. These will be discussed separately. Field observation data can be considered “traditional” in this respect.

#### **12.4.1 Traditional intensity assessment**

Although the conversion of descriptive information to numerical intensity data by use of an intensity scale is fundamental to macroseismic studies, the process was in general rather poorly documented in the past. This led to considerable variations in practice from worker to worker, resulting in serious inconsistencies in results. It is widely recognised that assessing intensity in the traditional way is to some extent a subjective exercise, and that some variations between workers will always occur, but it is better if these are minimised through common methodology as much as possible.

The following points apply to most common intensity scales:

- Data should be grouped by place prior to assessing intensity. By "place" is meant a village or town or part of a city. Places should not be too big (like a county) or too small (like a single house). When assessing intensity for a place, all the data relating to that place should be considered together. If there are fifteen reports from one village, a single intensity should be assigned to those fifteen jointly, rather than making fifteen assessments and combining them. It is common for intensity scales to use phrases like "in many cases", or "in few cases", and this cannot be assessed on the basis of one questionnaire, hence the need to combine them.
- Make sure there are sufficient data for a reasonable assessment. If there are too few reports, or the reports are too lacking in detail, it is better to record merely that the earthquake was felt rather than forcing an intensity value on inadequate data. In some cases it will be possible to make a range assessment, e.g. 4-5, >6, (4 or 5, more than 6) etc. A single report may be very unrepresentative of the general experience.
- For each place, compare the picture of earthquake effects provided by the data with the idealised pictures provided by each description of an intensity degree in the intensity scale, in order to look for the best overall fit. The match will seldom be perfect, so it is necessary to look for the most coherent, general comparison. It should be remembered that given the very variable nature of intensity, in any place individual effects may be observed that are higher or lower than those to be expected from the general (modal) intensity level. It is important not to give these too much attention. For example, if most of the data for a place are suggesting intensity 4, but there is a single exceptional report that a chimney fell, this chimney does not invalidate an assessment of intensity 4 for the place.
- When using a quantitative intensity scale (MSK, EMS) then the comparison of the data with the scale will usually be a question of making a best fit of the percentages of a particular observation that were recorded and the percentage ranges expected for each degree of the scale. For EMS-98, the procedures for assessing intensity are discussed in detail in the scale support material (Grünthal 1998).
- The absence of reports of a particular phenomenon may or may not be evidence that it did not occur, depending on the nature and quality of the data. It cannot automatically be inferred that, for instance, an absence of reports of damage indicates no damage occurred, although this will often be the case. A positive statement along the lines of "there was no damage" is more reliable.
- To make inferences from a particular source of data requires an understanding of the nature and limitations of that data source. For instance, newspapers often pluralise things for effect, so a newspaper report that says "pictures fell from walls" may mean one picture fell from one wall. Where the sources are historical documents the advice of a professional historian in understanding the nature of the documents should be taken.
- Effects on nature (landslides, ground water changes, etc) should only be used with caution, since their frequency is strongly influenced by local hydrological conditions and other factors not related to intensity. It is therefore very difficult to arrive at reliable intensity values for remote, largely uninhabited, rural areas. (This point has been found unpalatable by some, but unfortunately is realistic).

#### **12.4.2 Automatic intensity assessment**

The collection of macroseismic data rapidly from internet questionnaires lends itself to automatic data processing. If intensity can be assigned algorithmically, the results of the

macroseismic survey can be continually updated and displayed in real time, a huge advantage. This also does away with any subjectivity in assessment, since the results are defined in terms of the algorithm, and are totally objective. The downside is that the correspondence between the algorithm and the actual intensity scale that it is intended to mimic is uncertain.

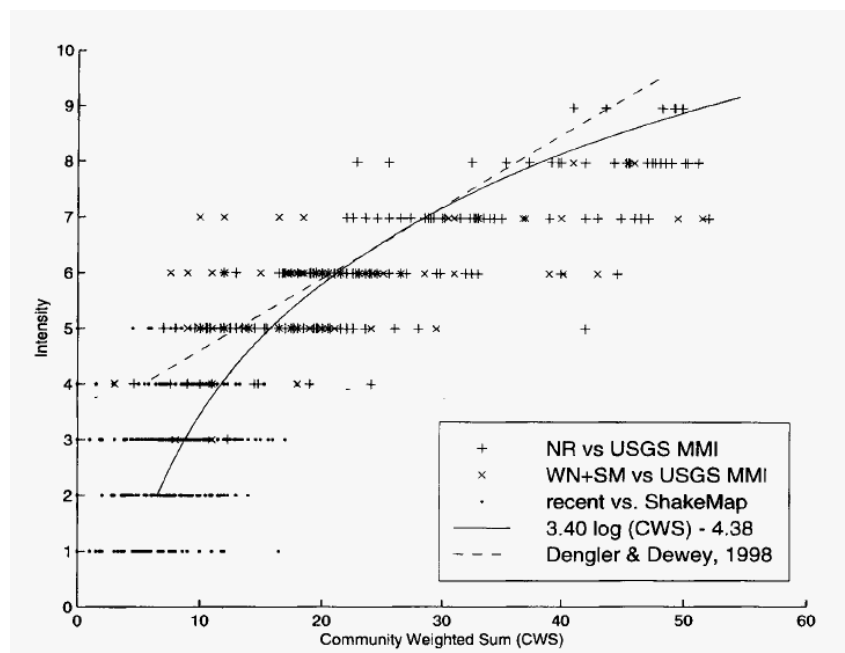
There are several systems in current use, but all can be divided into one of two approaches: those based on regression, and those based on an expert-systems approach.

The regression approach is typified by the “DYFI?” system, the basis of which is fully described in Wald et al (1999). In this system, each questionnaire is assessed as it is received, and scores are given based on answers to key questions. For each location from which responses are received, the scores from each question are tallied. The final score for the location is the weighted sum of the average scores for each question.

When the system was set up, a training set was compiled of scores for a set of locations, together with human-assigned intensities for those locations. From that training set, a regression equation was developed between location score and intensity (Figure 12.3). For a current earthquake, this equation gives the intensity for any location by simply converting the final score value, or “community weighted sum” (CWS). Intensity is calculated from the CWS according to the equation

$$I = 3.4 \ln \text{CWS} - 4.38 \quad (12.1)$$

This allows decimal intensities, in that, for instance, a CWS of 15 gives an intensity of 4.8 MM – but this should be tempered by the standard error of equation (12.1), which is unfortunately not given by Wald et al. (1999), but appears by eye to be between  $\pm 0.5$  and  $\pm 1.0$  degrees. The lower the CWS the greater the scatter in the regression, though this is probably mainly due to poorly constrained values in the training data set. The inherent uncertainty in computed values is uncertain, but likely to be higher for locations with few questionnaires.



**Fig. 12.3** Regression of community weighted sum against intensity; reproduced from Figure 1 in Wald et al. (1999). NR = Northridge, WN = Whittier Narrows, SM = Sierra Madre. For further information, see the original paper.

The system does not impose a minimum number of questionnaires before assigning an intensity, but the number received may be useful supplementary information in estimating uncertainty.

The expert systems approach is quite different, and seeks to emulate the thought processes of a seismologist using an intensity scale in the traditional way. Such a system is described by Musson (2006) for the EMS-98 scale. The basis of it is an assessment of the totality of reports received from a location compared to the requirements of the scale. For this purpose, for key effects, the proportion of observers at a location who reported the effect is computed. This might be, for instance, that 35% of questionnaires from a place answered yes to “did windows or doors rattle?”. Such ratios cannot meaningfully be established from single or very few questionnaires, so for places with very few replies, only “felt” will be returned.

Once the ratios have been computed, the effects at each place are considered by a process of elimination. This runs as follows:

- First, was the earthquake felt at all? If all the reports are negative, the intensity is 1 EMS and processing stops. Second, are there many reports of substantial damage? If so, the intensity is 7 or 8 EMS (8 being the upper limit of the system) and more detailed tests are applied to choose which.
- If the intensity was not 1, 7 or 8, the next thing to look at is whether there is a high proportion of “strong” effects like objects falling, people running out, minor damage, and so on. If this is the case, the intensity is 5 or 6 EMS and more detailed tests are applied to choose which. If this is not the case, only intensities 2, 3 and 4 are left. If the preponderance of data suggests that many or most people felt the earthquake and there were some physical manifestations like rattling doors, the intensity is assessed as 4 EMS. If it is found that only people at rest on upper floors felt it, the intensity is 2 EMS. If neither of these is the case, the intensity is finally 3 EMS.

Both approaches have positive and negative features. Regression-based systems can be calibrated to past national data sets to maintain consistency; but they then inherit whatever subjective practices informed the intensity assessments of the calibration data set. The expert systems approach, in contrast, is tied to the interpretation of the actual intensity scale, and not any previous data sets (and therefore may not be precisely compatible with them).

The regression approach will always return a numerical intensity, where the expert systems approach will not. On the other hand, the uncertainty in intensity estimates from a regression-based system may be large, especially if based on few questionnaires. The expert systems approach effectively redefines the scale as a series of specific tests, from which it follows that there is no uncertainty at all in any assigned intensity value – at least, with respect to the data on which the assignment is made.

In both approaches, there is still the problem that a self-selected sample from a large population (the total number of inhabitants of the location) is likely to be unrepresentative. Therefore, besides any inherent uncertainty in assessment, there is an additional unquantifiable uncertainty as to the relation between the sample (the questionnaires received) and the total population.

Note that this uncertainty would be quantifiable if the sample was randomly drawn from the population, instead of being self-selecting. As well as a probable bias towards positive

responses, there may also be a demographic bias, in that some categories of people are less likely to be internet users. These people are also likely to live in poorer housing, and there may in some cases be other influences on the data, for example in places where wealthy suburbs are built on high ground and poorer suburbs on reclaimed land.

More details on these two systems, and the actual software required to implement them, are available from USGS and BGS respectively, on request.

### **12.4.3 Accuracy of assessment**

The previous paragraphs lead naturally into a consideration of what degree of accuracy in intensity assessment can be expected in general.

Given a certain strength of shaking, it is to be expected that buildings of equivalent strength will not respond in a completely uniform way. Rather, there should be a modal level of damage observed, with some buildings suffering less and others more. The net effect approximates to a normal distribution (as has often been seen in damage surveys). Thus, for any particular level of shaking, it is expected to be found that different percentages of the building stock of a given strength will suffer different degrees of damage. In assessing intensity (and this is true of the lower degrees as well as the damaging ones) one is usually dealing with a sample or estimate of the percentages that were observed, and attempting to match these to the expected ranges for one of the intensity degrees. In most cases, given a degree of robustness, an adequate fit can be found without much problem.

Difficulties can occasionally arise when this task is compounded, as it sometimes is, by one or other of two factors: (i) that the effects of an earthquake vary considerably over very short distances, due to a combination of local conditions and the complexity of earthquake ground motion; and (ii) information is often not complete. These two factors can have an effect on the level of accuracy that can be expected in intensity assessments. The variability of earthquake effects is well-known, as in cases where, of two identical houses side-by-side, one is heavily damaged and the other nearly intact. This may give a misleading impression of difficulty in assessing intensity which might actually disappear once a larger sample of houses was assessed. The difficulty with respect to information is that one is often working from an uncontrolled, possibly unrepresentative, sample of the whole data population (particularly when not working with data derived from a field investigation), and there may also be uncertainty about the condition of buildings before the earthquake. This can cause problems where the real percentage distribution of effects is obscured by the limited data. Also, even when the amount of data is good, there can be cases where the reported effects do not match unambiguously any of the "pen pictures" presented by the classes in the intensity scale.

In cases where one cannot determine intensities to a resolution of one degree, two degrees can be bracketed together to show the probable range. This can be particularly the case for lower intensity degrees; there are often cases where it is hard to be sure between intensity 2 or 3, or between 3 or 4, or between 4 and 5. In such cases one may write 4-5, meaning either intensity 4 or intensity 5.

One guideline that can be suggested is that the description of each degree should be considered the minimum case. For example, in a case in which the data satisfy the requirements for intensity 4, but certainly do not adequately satisfy the criteria for intensity 5,

then the correct assessment is 4, even if the effects seem stronger than the basic intensity 4 description.

Intensity scales are designed to include the necessary degree of robustness to make identification of the different degrees as practical as possible. The number of degrees in a scale is controlled by the number of different levels that can be distinguished in normal use without too much difficulty. Experience shows that it is very unlikely that one could ever meaningfully discriminate intensities to a resolution of less than one degree of a twelve-degree scale. If one could state accurately in some case that the intensity was, for example,  $6\frac{1}{2}$ , this would imply that a 23 degree intensity scale could be written, which is doubtful.

This reservation does not apply to algorithmically computed intensities; as shown above, a “DYFI?” intensity of 4.8 has a precise meaning – that the community weighted sum for the locality from the questionnaires received was 15. While it might not seem that “ $4.8 \pm 0.5$ ” is really more meaningful than “5” or even “4-5”, it is reported that the fractional intensities computed this way do tend to show a smooth attenuation with distance (Atkinson and Wald 2007).

It may be no coincidence that modern intensity scales are converging on ten degrees as the optimal amount of discretisation possible. The JMA scale, with its degree 0 and split 5 and 6 degrees has ten degrees in total, and the major twelve-degree scales are effectively ten-degree scales, since in modern practice, intensity 11 and 12 are not used. At very high intensities, scales tend to saturate once all buildings have been destroyed. In areas where the building stock is very poor, this can be as low as 8 or 9 MMI-MSK-EMS. EMS-98 attempts to discriminate intensities 11 and 12 by extrapolating damage distributions to building stock with modern anti-seismic design, but this is largely untested in practice. Some other scales define intensity 11 and 12 in terms of secondary effects that are actually due to surface faulting, slope failure or liquefaction, and are not truly indicators of intensity.

#### **12.4.4 Equivalence between scales**

The translation of values from one scale to another has been discussed in detail by Musson et al. (2010), whose conclusions can only be summarised here.

It has often been the practice to attempt to express the equivalence between different intensity scales by way of a chart that compares different degrees of a scale by a series of rectangles overlapping to a smaller or larger extent. Such charts are unhelpful, partly because they imply that, for instance, intensity 5 in one scale may be “slightly higher” than intensity 5 in another scale, when in integer degrees there is no “slightly higher”. Secondly, any set of intensities is a combination of the scale employed and the working practices used in the assessment. One worker may make the assumption that if there is any damage, the intensity must be at least 6, another worker may not, and the result will be consistent variation even if both are using the same intensity scale.

For practical purposes, therefore, it is much preferable to revisit the original data and make a fresh intensity assignment with the desired intensity scale. However, this may not always be possible.

Comparing the scales as written, rather than as used, Musson et al. (2010) found no significant differences between MM-56, MSK, MCS and EMS-98 This is in accord with the

table of equivalences by Shebalin in the original MSOP (Willmore 1979) for MM-56, MCS and MSK. It is noted though, that in practice, intensity data points assigned with MCS have a tendency to be higher than those assigned using other scales. The Stover and Coffman (1993) scale breaks the pattern slightly, since 1 MM-93 could be 1 or 2 EMS-98, and more importantly, 4 MM-93 could be 4 or 5 EMS-98.

The higher degrees of scales cannot really be compared because of saturation problems, as discussed above.

A rough equivalence for the JMA (2009) Scale is given in Tab. 12.1 below. This varies significantly from the table given in Musson et al. (2010), due to the changes made in the 2009 revision of the JMA scale, after Musson et al. (2010) was submitted.

**Tab. 12.1** Equivalence between JMA-09 and EMS-98 intensity values

| <i>JMA Scale</i> | <i>EMS-98</i> |
|------------------|---------------|
| 0                | 1             |
| 1                | 2             |
| 2                | 3             |
| 3                | 4             |
| 4                | 5             |
| 5 Lower          | 6             |
| 5 Upper          | 7             |
| 6 Lower          | 8             |
| 6 Upper          | 9             |
| 7                | 10            |

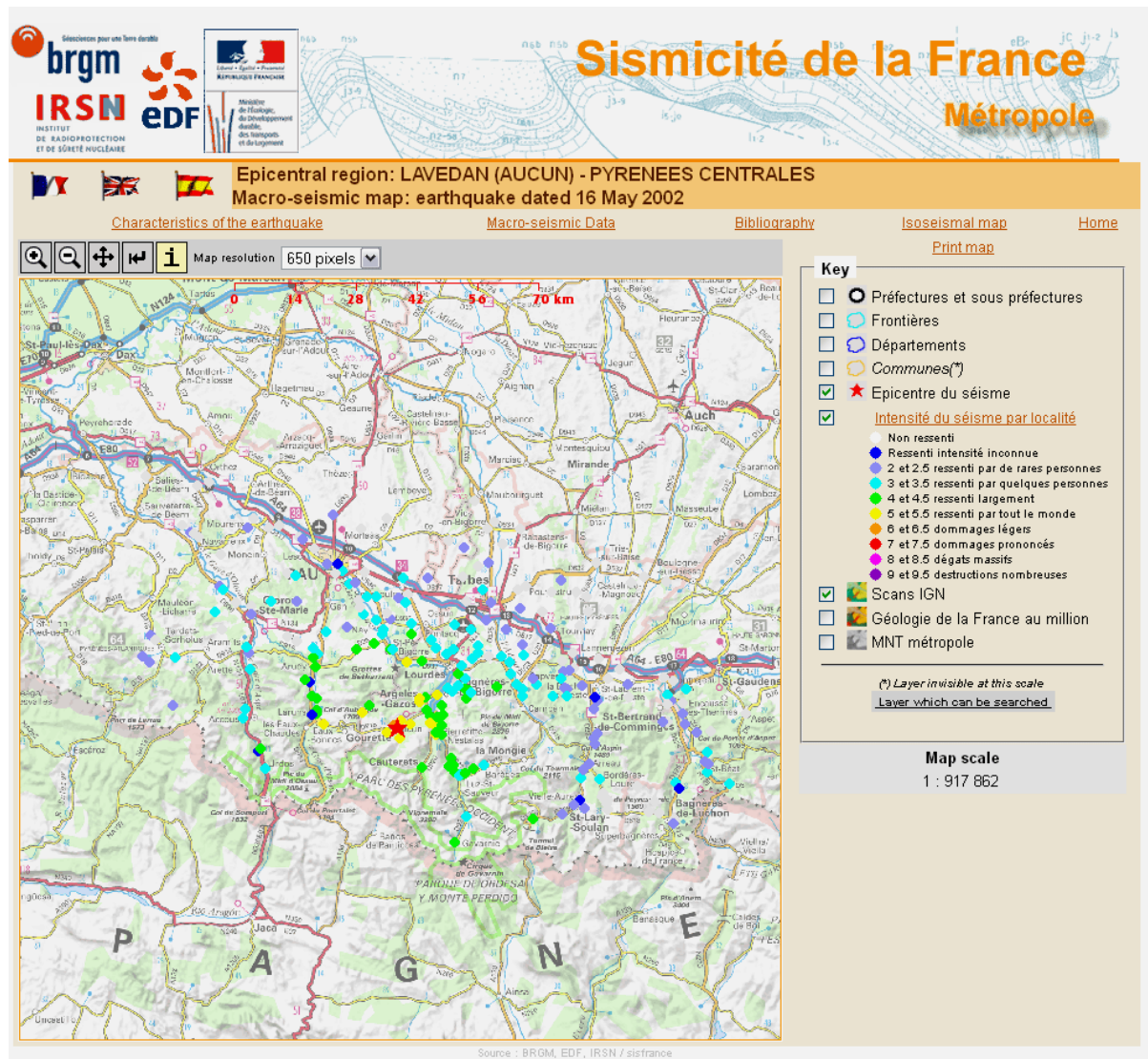
## 12.4.5 Presentation of intensity data

Practice in the earlier part of the 20<sup>th</sup> century was generally to publish intensity data in the form of an isoseismal map. An isoseismal can be defined as a line bounding the area within which the intensity is predominantly equal to, or greater than, a given value. No definitive method of contouring has ever been agreed. Some workers adopt a practice of overlaying a grid on the data and taking the modal value in each grid square prior to contouring, others prefer to work directly on the plotted intensity values. Workers have differing preferences for the amount of smoothing, extrapolation, etc, that is to be employed. Thus, the drawing of isoseismals has always been to some degree subjective.

While an isoseismal map often gives a good visual impression of the effects of an earthquake, and the areas within each isoseismal can be used for some purposes (e.g. estimation of focal depth; see 12.5.5.), such maps are inadequate for modern studies, partly due to the subjectivity problem, and partly because many applications of interest require point data (e.g. Bakun and Wentworth 1997, Gasperini et al. 1999, Musson and Jiménez 2008).

The primary requirement, therefore, is to publish the intensity data for any earthquake in the form of a table of intensity data points (IDPs), where each IDP has, as a minimum, latitude, longitude and intensity (using Arabic numerals, not Roman ones), and may also have locality name, some quality factor (for instance, the number of questionnaires contributing to the

assessment) and other metadata. Such publication is now sometimes done in the form of an online database, much to be encouraged – SISFRANCE (www.sisfrance.net) is an example (Fig. 12.4).



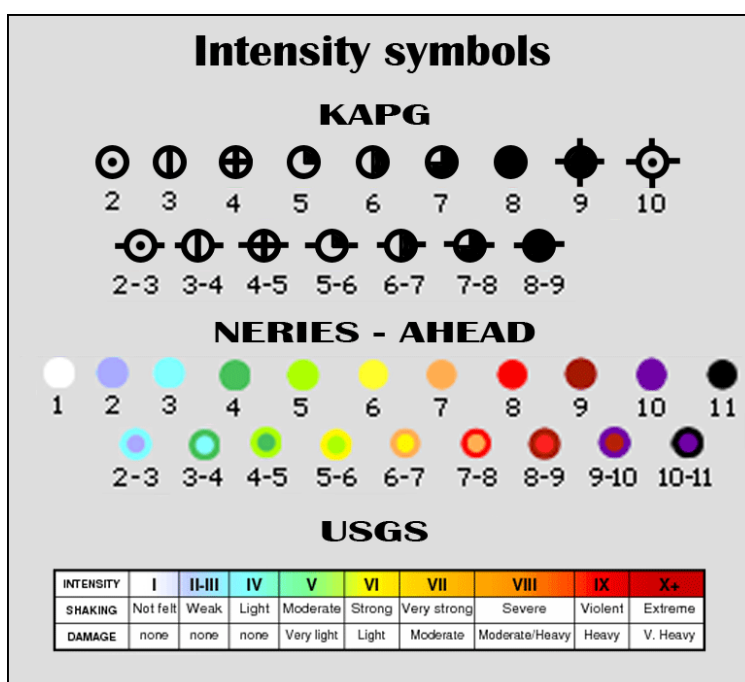
**Fig. 12.4** Screenshot from the SISFRANCE website, showing a macroseismic display of a Central Pyrenees earthquake from 2002.

Mapping of intensity data is now predominantly done by showing maps of the IDPs themselves. There are three principal ways of representing IDPs on a map. The simplest is to plot each intensity as a number. This works adequately if there are only a few well-spaced points to be displayed, but if the point density is high, the map quickly becomes hard to read. Also, such a map does not give a very clear visual impression of the overall intensity distribution. A better alternative is the set of international symbols (first introduced by KAPG, the Commission of the Academies of Sciences of Socialist Countries for Planetary Geophysical Research) based on colouring in different proportions of small circles (Fig. 12.5). This gives more of a visual impression, and can be used in monochrome figures. The KAPG



symbols are only really standard between intensities 2 and 8; outside this range they vary between different studies. Those shown in Fig. 12.5 are as used in Shebalin (1973).

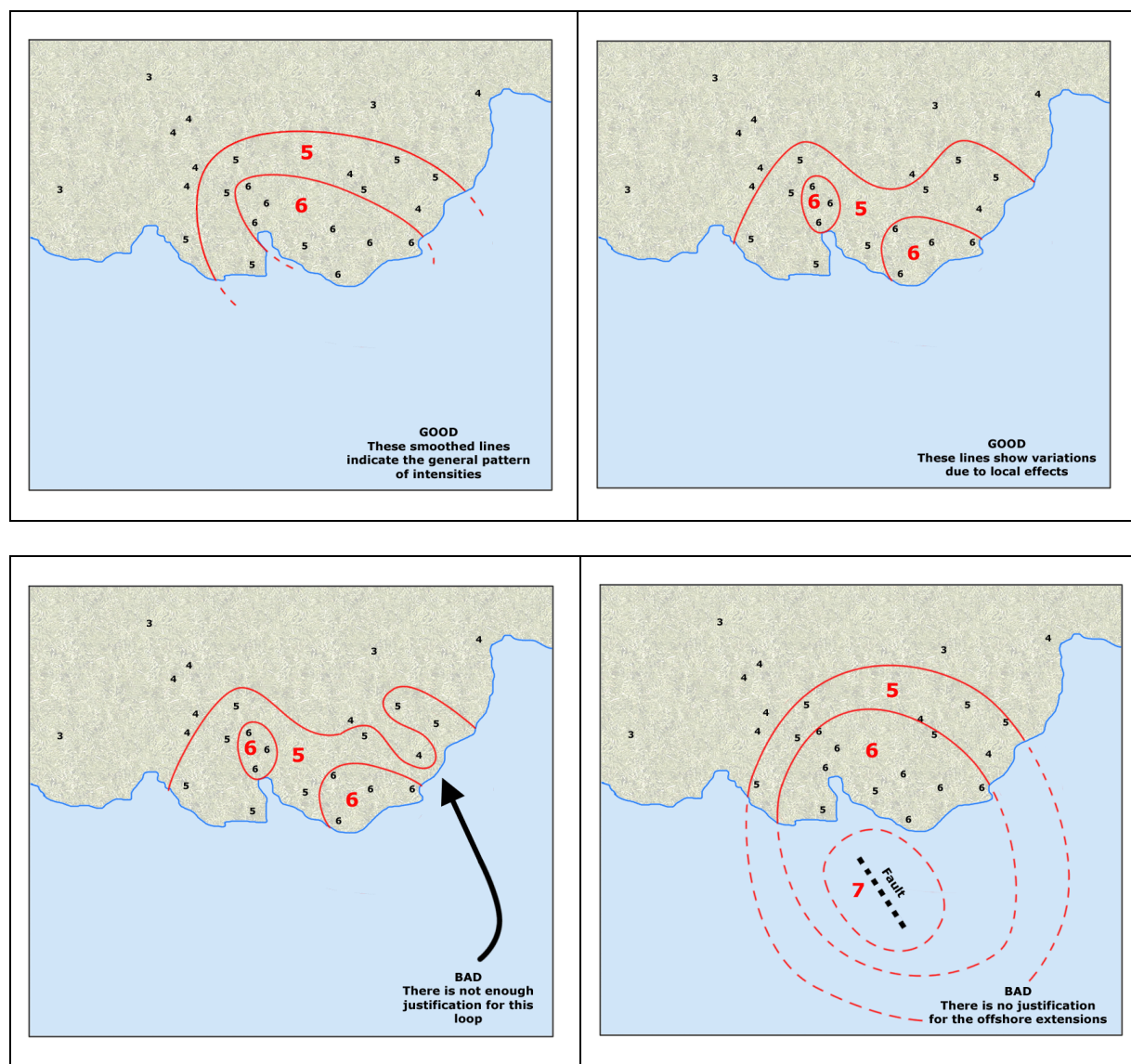
With colour printing becoming more accessible in journals, and straightforward in internet publication, the most common system now is to use coloured dots for IDPs. While there is no single standard colour scale, the norm is to use a colour spectrum such that low intensities use blues and greens, middling ones use yellow and orange, and higher ones use reds, browns, purples and black. Fig. 12.5 shows such a scheme, devised for the Archive of Historical Earthquake Data (AHEAD), produced by the NERIES project in Europe (see <http://www.emidius.eu/AHEAD/>), which includes a convention for uncertain values split over two degrees. In this system, “felt” is plotted as an empty green circle. Also shown is the colour scale for use with the DYFI? System. The SISFRANCE colour scheme is visible in Figure 12.4.



**Fig. 12.5** Three different systems for plotting intensity values; monochrome symbols and two approaches to using colour.

Isoseismal maps still have their uses. The degree of smoothing employed in their construction should reflect the purposes to which the resulting map will be put. If the map is intended for microzonation work, i.e. to point up areas where seismic hazard may be enhanced owing to local soil conditions, then smoothing will be at a minimum, and isoseismals will be as convoluted as the data. If the map is intended for other purposes (calculation of earthquake parameters, attenuation studies, tectonic studies, etc) then the curves will normally be smoothed so that only major re-entrants and outliers are shown. In practice, smoothed isoseismals are much more common. As a general rule, re-entrants and outliers should not be drawn unless suggested by a grouping of at least three data points. If isoseismals have to be interpolated or extrapolated across areas of water, or areas without data points, these sections of the lines should be shown as dashed. In cases where, for example, an epicentre is offshore, and only (say) a 120° arc of each isoseismal would fall onshore, it is not correct to project the whole of the remaining 240° of each isoseismal on a map, even as a dotted line. Only the

onshore section should be drawn, with each line tailing off with a short dotted section offshore if desired. Plotting isoseismals that are completely offshore and merely projections of an intensity attenuation curve should not be done. For onshore earthquakes with few data, it is not good practice to attempt to draw isoseismals conjectured from one or two points only; at least three mutually supporting data points for one intensity value should exist before one attempts to draw even a partial isoseismal for that value. Figure 12.6 shows some examples of acceptable and unacceptable drawing of isoseismals.



**Fig. 12.6** Two examples each of acceptable (upper row) and unacceptable practice (lower row) in drawing isoseismals.

Computer contouring programs often do not give good results with macroseismic data, as too often parts of the macroseismic field are poorly constrained; though Schenková et al (2007) report good results from kriging (a linear least-squares method commonly used for interpolation of geospatial data in the earth sciences).

## 12.4.6 Exchange of macroseismic data

Earthquakes are obviously no respecters of international boundaries, and it is not uncommon for the felt area of a significant earthquake to extend into more than one country. In the case of the 1992 Roermond earthquake, macroseismic data was collected from the Netherlands, Germany, Belgium, Luxembourg, France, the UK, Switzerland and the Czech Republic (Haak et al 1995). Exchange of data is clearly important. However, problems can arise if what is exchanged is simply tables of IDPs, as these may be assessed using different scales or methods, and may not be compatible. Some means are necessary for exchanging data at a lower level – that of the original questionnaire, at least in some abstracted form. A format is needed that can provide a record for each questionnaire that includes, besides location, values representing the effects reported. The greater the degree of international standardisation of questionnaires, the easier such a format will be to construct, but clearly, it needs to represent a superset of the types of information collected by different questionnaires and used by different systems. At the time of writing, there are initiatives to develop such a format in the scope of QuakeML (<https://quake.ethz.ch/quakeml/QuakeML>).

## 12.5 Determination of earthquake parameters from macroseismic data

The determination of earthquake parameters (principally epicentre and magnitude, also depth in some cases) from macroseismic data is of critical importance for historical seismology, where it is the only way in which one can assess earthquakes of the pre-instrumental period. For observatory practice, these parameters will of course be determined from instrumental data, but it can be instructive to compute the macroseismic equivalents as well, partly as a means of checking how a recent earthquake would appear in the catalogue had it occurred before 1900. The highest intensities may in some cases be some distance from the instrumentally-determined epicentre (i.e. the point on the earth's surface above the focus of the earthquake where the rupture initiated).

There is also one purely macroseismic parameter, epicentral intensity (and its close relation, maximum intensity) to be considered.

### 12.5.1 Epicentral intensity

Epicentral intensity, usually abbreviated  $I_0$ , is a parameter commonly used in earthquake catalogues but rarely defined, and it is clear that different usages exist in practice (Cecić et al 1996). The meaning of the term is clearly the intensity at the epicentre of the earthquake, but since it is likely that there will not be observations exactly at the epicentre itself, some way of deriving this value is necessary. The two main techniques that have been used in the past are:

1. Extrapolation from the nearest observed intensity data to the epicentre without changing the value, or use of the value of the highest isoseismal. Thus if there are a few data points of intensity 9 near the epicentre, the  $I_0$  value is also 9. If the epicentre is significantly offshore,  $I_0$  cannot be determined.
2. Calculating a fractional intensity at the epicentre from the attenuation over the macroseismic field, using a formula such as that by Blake (1941) or Kövesligethy (1906).

In this case, because this is not an observed value (and not a "true" intensity) it may be expressed as a decimal fraction without contravening the rule that intensity values are integer.

It is recommended that these two methods be discriminated between by the notation used. Thus an integer number (9 or IX) indicates method (1) and a decimal number (9.0 or 9.3) indicates method (2). It is recommended that one should not add arbitrary values to the maximum observed intensity when deriving an  $I_0$  value; the arbitrary amount is too subjective.

As well as epicentral intensity, a useful parameter is maximum intensity, abbreviated  $I_{\max}$ . This is simply the highest observed intensity value anywhere in the macroseismic field. For onshore earthquakes,  $I_0$  and  $I_{\max}$  may be equal. For offshore earthquakes it is often not possible to estimate  $I_0$  (especially if method 1 is used), but  $I_{\max}$  can be given for any earthquake.

### 12.5.2 IDPs and isoseismals

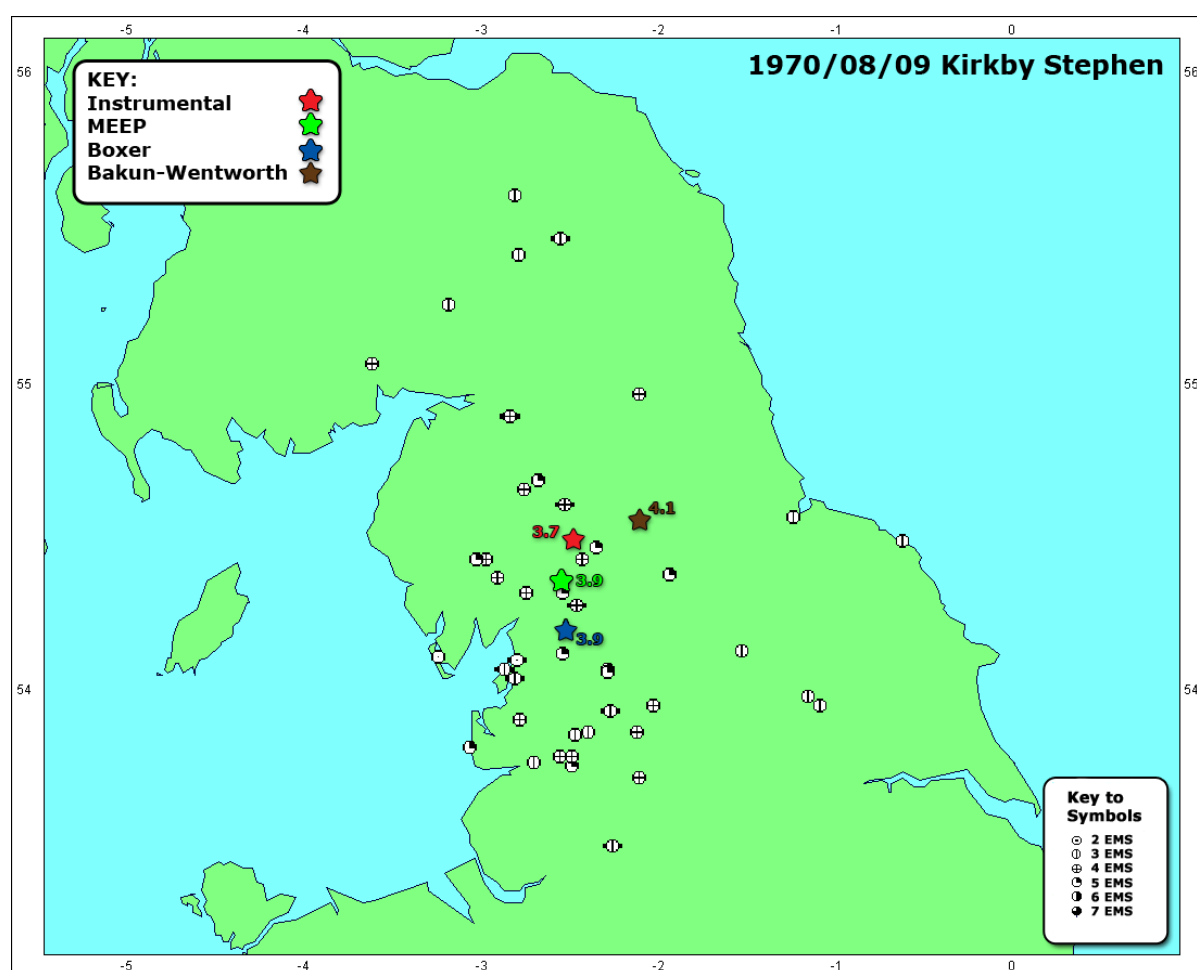
There are two basic strategies in determining parameters from macroseismic data – to work with the raw IDPs, or to draw isoseismals and work with those.

The reason for using isoseismals is this: they can reduce or eliminate effects due to population distribution. Consider Fig. 12.7, in which an earthquake produces its highest intensities at several places on the mainland and in one island location. A seismologist, knowing that the irregular distribution is due to topography, might estimate an isoseismal as shown, which might be a better expression of reality than that obtained from processing of the individual data points, in which the clustering due to population factors has an influence. From an isoseismal one could deduce an epicentre in the position of the solid star; from data points one would be likely to obtain the position of the open star.



**Fig. 12.7** Illustration of different macroseismic epicentres obtained from (a) intensity 6 data points (open star) and (b) a possible interpretation of isoseismal 6 (solid star).

However, because of the subjectivity in isoseismal construction, modern opinion favours the direct use of the IDP distribution, without any isoseismals. Currently three systems are available for the computation of the main earthquake parameters from IDP sets. These are the Bakun-Wentworth (B-W) method (Bakun and Wentworth 1997), the BOXER method (Gasperini et al. 1999) and MEEP (Macroseismic Estimation of Earthquake Parameters, Musson and Jiménez 2008). These will be discussed further in the next sections. Fig. 12.8 shows a comparison of results from the three methods for a sample earthquake. In this case, the parameters obtained can be compared to instrumental values; for historical earthquakes it can be the case that location methods disagree with no clear indication which is more reliable.



**Fig. 12.8** Comparison of macroseismic parameter estimations for a British earthquake, together with the instrumental parameters.

### 12.5.3 Macroseismic epicentre

Macroseismic epicentre is an expression which has been used in the past to convey different concepts, never properly defined. On the one hand, the macroseismic epicentre can be considered to be the best estimate made of the position of the epicentre without using instrumental data. This may be derived from any or all of the following circumstances dictate: position of highest intensities, shape of isoseismals, location of reports of foreshocks



or aftershocks, calculations based on distribution of intensity points, local geological knowledge, analogical comparisons with other earthquakes, and so on. This is a rather judgmental process with some subjectivity, and does not lend itself to simple guidelines that can be applied uniformly in all cases.

More commonly, the term is used for the point on the Earth's surface from which the macroseismic field appears to radiate. In the past this was often the centre of the highest isoseismal or the weighted centre of the two highest isoseismals. The terms “barycentre” “intensity centre”, “macrocentre” and “macroseismic centre” have been proposed as alternative terms.

These two points are often the same, but need not be. For any study of the tectonics of an area, the macroseismic epicentre is perhaps more useful. For studies of seismic hazard, especially those using a technique like extreme value statistics, the barycentre gives a better indication of the hazard potential of an earthquake.

The BOXER method computes epicentres by taking the trimmed mean of the latitudes and longitudes of the highest intensity locations (using also the second-highest values if there are insufficient points otherwise). It also computes an estimate of the rupture dimensions of the fault; this is the only method that handles extended sources from IDP data.

The B-W and MEEP procedures both work on the basis of attempting to minimise residuals from a theoretical intensity distribution that can be derived from some attenuation model, a procedure first proposed by Perruzza (1992). In the case of the B-W method, the model used is of the type expressing intensity as a function of magnitude and distance. In MEEP, the model used is one expressing the reduction of intensity from the  $I_0$  value as a function of distance. In both cases, a grid search is used to locate the position at which the misfit between the best set of predicted values and the actual values is minimised. Both these methods are capable of locating earthquakes offshore, which BOXER generally cannot do. However, especially with poor data sets, they can get trapped in false minima.

#### 12.5.4 Macroseismic magnitude

The use of macroseismic data can give surprisingly robust measures of earthquake magnitude. Early studies attempted to correlate epicentral intensity with magnitude; however, epicentral intensity can be affected by focal depth or soil amplification, so such simple correlations may perform poorly.

The total felt area ( $A$ ) of an earthquake, or the area enclosed by one of the outer isoseismals (usually 3 or 4), tend to be a better indicator of magnitude, being not much affected by depth (except in the case of truly deep earthquakes) or soil conditions, and since the relationship between the area and magnitude is logarithmic, even a substantial mis-estimation of the total felt area will not produce an overly large error in the magnitude. For earthquakes below a threshold magnitude (about 5.5 Mw) magnitude and log felt area scale more or less linearly, and so equations of the form

$$M = a \log A + b \quad (12.2)$$

can be established regionally by examination of data for earthquakes for which macroseismic data and instrumental magnitude are both available. For larger earthquakes, differences in

spectral content may affect the way in which earthquake vibration is perceived, and a different scaling appears to apply. In Frankel (1994) the form

$$M = n \log (A / \pi) + (2 m / (2.3 \sqrt{\pi})) \sqrt{A} + a \quad (12.3)$$

is used to represent the full magnitude range, where  $n$  is the exponent of geometrical spreading,  $a$  is a constant, and

$$m = (\pi f)/(Q \beta) \quad (12.4)$$

where  $f$  is the predominant frequency of earthquake motion at the limit of the felt area (probably 2-4 Hz),  $Q$  is shear wave attenuation quality factor and  $\beta$  is shear wave velocity (3.5 km/sec). Using this functional form and comparing world-wide intraplate earthquakes with interplate earthquakes from one region (California), Frankel found the difference in magnitude for the same felt area to be on average 1.1 units greater for California.

In the above equations,  $M$  has been used for generic magnitude; for any particular magnitude equation it is important to specify what magnitude type the derived values are compatible with ( $M_s$ ,  $M_L$ ,  $M_w$  etc). It is also useful to determine the standard error, which will give a measure of the uncertainty attached to estimated magnitude values.

Magnitudes calculated by BOXER adapt a model proposed by Sibol et al. (1987):

$$M = a + b I_o^2 + c \log^2 (A_I). \quad (12.5)$$

Here  $A_I$  is the area of each intensity  $I$ , calculated from the median distance of IDPs of value  $I$ . The formula is applied for each intensity in the data set, and the magnitude is given by the weighted average.

For the B-W method, since it is based on the optimisation of a model including magnitude, the optimal magnitude is a by-product of the location procedure.

MEEP adapts the Frankel model given above to estimate the limit of intensity 3 for a hypothetical magnitude. The radii of higher “isoseismals” are then interpolated and compared to the data to find the optimal magnitude.

Common to all methods is the need to perform a regional calibration before magnitude can be calculated from intensity data, something not needed for computation of epicentres.

## 12.5.5 Estimation of focal depth

The estimation of focal depth from macroseismic data was first developed by Radó Kövesligethy. His first paper on the subject presented the formula

$$I - I_o = 3 \log \sin e - 3 \alpha (r/K) (1 - \sin e) \quad (12.6)$$

where  $\sin e = h / r$  ( $h$  is depth,  $r$  is isoseismal radius) and  $K$  is the radius of the earth, and  $\alpha$  is a constant representing anelastic attenuation (Kövesligethy 1906). A second paper, (Kövesligethy 1907) contains a different equation:

$$I - I_0 = 3 \log \sin \varphi \quad (12.7)$$

where  $\varphi$  is the angle of emergence. Why the absorption term was dropped in this publication is unclear. Equation 12.6 was subsequently rewritten and modified slightly by János (1907), to reach the now well-known formula

$$I_0 - I_i = 3 \log (r / h) + 3 \alpha m (r - h) \quad (12.8)$$

where  $r$  is the radius of the isoseismal of intensity  $I_i$  and  $m = \log e$  (Euler's constant). This work was developed further by Blake (1941) whose contribution was essentially a reduction and simplification of equation (12.8); Blake's version is still used by some workers today, but Kövesligethy's original equation (in János's version) is more commonly encountered. Kövesligethy's equation became more widely known, in the form of equation (12.8), through the work of Sponheuer (1960). However, although Sponheuer references Kövesligethy (1906) in his text, he cited Kövesligethy (1907) in the reference list, and the relative inaccessibility of these papers, and this misreference, has caused some confusion which it has only recently been possible to unravel. A further puzzle is that János (1907) attributes equation to (12.8) to Cancani, transmitted by Kövesligethy; the involvement of Cancani in this is unclear (Zsiros 1999, pers. comm.).

The constant value of 3 used in equations 12.6 - 12.8 represents an equivalence value between the degrees of the intensity scale and ground motion amplitudes. This may vary regionally, and can be optimised (Levret et al 1996). The attenuation parameter  $\alpha$  should be determined regionally by group optimisation on an appropriate data set - not for individual earthquakes; some proposed values for different parts of Europe are given in Karník (1969).

$I_0$  here is an interpolated value, not the highest observed integer value, and has to be solved for as well as solving for  $h$ . This can be done graphically - one can fit the isoseismal data to all possible values of  $h$  and  $I_0$  and find a minimum error value consistent with the observed maximum intensity (e.g. Burton et al 1985, Musson 1996).

Macroseismic depth determination has attracted less interest than epicentre and magnitude, partly because for large earthquakes, focal depth is not of great practical importance. While the application of equation (12.8) can give quite good results for small crustal earthquakes, irregular data sets have a tendency to depress the depth estimate to the maximum allowed, and this is probably a worse problem when using IDPs rather than isoseismals. Of the three recent analysis methods from IDPs, MEEP is the only one that supports depth determination, using equation (12.8).

## 12.5.6 Intensity attenuation

Intensity attenuation, the rate of decay of shaking with distance from the epicentre, can be expressed in two ways. Firstly, there is the drop in intensity with respect to the epicentral intensity. This is shown by the Kövesligethy (1906) formula in equation (12.8).

It is generally more useful to express intensity attenuation as a function of magnitude and distance. Such formulae often have the functional form

$$I = a + b M + c \log R + d R \quad (12.9)$$



where  $R$  is hypocentral distance, and  $a$ ,  $b$ ,  $c$  and  $d$  are constants. (The last term is sometimes dropped, especially in intraplate areas). Since most earthquake catalogues include magnitude as a parameter, this form of intensity attenuation is extremely useful in seismic hazard studies. This type of model is also that required by the Bakun-Wentworth method described above.

A particular use of this type of model is in estimating the expected impact of an earthquake that has just occurred. Since a magnitude will be available at once, an equation of the form of (12.9) can give an immediate idea of how far the earthquake will be felt.

## Acknowledgments

The first author would like to thank the numerous people who made comments on the original text and suggested improvements; particularly Gottfried Grünthal and Tibor Zsiros. I am also appreciative of the support from members and associates of the NERIES-NA4 project, especially Bill Bakun, Antonio Gomez Capera, Paolo Gasperini, Maria-José Jiménez, Carlo Meletti and Max Stucchi. Thanks also to Jim Dewey of USGS and Maria-José Jiménez of CSIC for reviewing the final text; and also to Susanne Sargeant of BGS. This contribution is published with the permission of the Director of the British Geological Survey (NERC).

## References

- Ad-hoc Panel (1981). Report on the Ad-hoc Panel meeting of experts on up-dating of the MSK-64 intensity scale, 10-14 March 1980, *Gerl. Beitr. Geophys.*, 90, 261-268.
- Atkinson, G. M., and Wald, D. J. (2007). “Did You Feel It?” Intensity Data: A Surprisingly Good Measure of Earthquake Ground Motion. *Seismol. Res. Lett.*, **78**, 362-368.
- Bakun, W. H., and Wentworth, C. M. (1997). Estimating earthquake location and magnitude from seismic intensity data. *Bull. Seism. Soc. Am.*, **87**, 1502-1521.
- Blake, A. (1941). On the estimation of focal depth from macroseismic data. *Bull. Seism. Soc. Am.*, **31**, 225-231.
- Boatwright, J., Bundock, H. and Seekins, L. C. (2006). Using Modified Mercalli intensities to estimate acceleration response spectra for the 1906 San Francisco earthquake. *Earthquake Spectra*, **22**(S2), 279-296.
- Boatwright, J. and Bundock, H. (2005). Modified Mercalli intensity maps for the 1906 San Francisco earthquake plotted in ShakeMap format. *Open-File Report 2005-1135*, USGS.
- Brazee, R. J. (1978). Reevaluation of Modified Mercalli intensity scale for earthquakes using distance as determinant, *NOAA Tech. Mem EDS NGSDC-4*.
- Burton, P. W., McGonigle, R., Neilson, G., and Musson, R. M. W. (1985). Macroseismic focal depth and intensity attenuation for British earthquakes. In: *Earthquake engineering in Britain*, Telford, London, 91-110.
- Cecić, I., Musson, R. M. W., and Stucchi, M. (1996). Do seismologists agree upon epicentre determination from macroseismic data? A survey of ESC Working Group “Macroseismology”, *Annali di Geofisica*, **39**, 5, 1013-1027.
- Davison, C. (1900). Scales of seismic intensity, *Philosophical Magazine*, 5<sup>th</sup> Series, **50**, 44-53.

- Davison, C. (1921). On scales of seismic intensity and on the construction of isoseismal lines, *Bull. Seism. Soc. Am.*, **11**, 2, 95-129.
- Davison, C. (1933). Scales of seismic intensity: Supplementary paper. *Bull. Seism. Soc. Am.*, **23**, 158-166.
- Ebel, J., and Wald, D. J. (2003). Bayesian estimations of peak ground acceleration and 5% damped spectral acceleration from Modified Mercalli intensity data. *Earthquake Spectra*, **19**, 511-530.
- Ad-hoc Panel (1981). Report on the Ad-hoc Panel meeting of experts on up-dating of the MSK-64 intensity scale, 10-14 March 1980, *Gerl. Beitr. Geophys.*, **90**, 261-268.
- Frankel, A. (1994). Implications of felt area-magnitude relations for earthquake scaling and the average frequency of perceptible ground motion. *Bull. Seism. Soc. Am.*, **84**, 2, 462-465.
- Gasperini, P., Bernardini, F., Valensise, G., and Boschi, E. (1999). Defining seismogenic sources from historical earthquakes felt reports. *Bull. Seism. Soc. Am.*, **89**, 94-110.
- Grünthal, G. (Ed.) (1998). European Macroseismic Scale 1998, *Cahiers du Centre Européen de Géodynamique et de Seismologie*, **15**, Conseil de l'Europe, Luxembourg, 99 pp.
- Haak, H. W., van Bodegraven, J. A., Sleeman, R., Verbeiren, R., Ahorner, L., Meidow, H., Grünthal, G., Hoang-Trong, P., Musson, R. M. W., Henni, P., Schenková, Z. and Zimová, R. (1995). The macroseismic map of the 1992 Roermond earthquake, the Netherlands. *Geologie en Mijnbouw*, **73**, 265-270.
- Jánosi, I. (1907). Makroszeizmikus rengések feldolgozása a Cancani-féle egyenlet alapján. In: Réthly, A., (Ed.) Az 1906 évi Magyarországi Földrengések, *A. M. Kir. Orsz. Met. Föld. Int.*, Budapest, 77-82.
- JMA (Japan Meteorological Agency) (1996). Explanation Table of the JMA Seismic Intensity Scale (February 1996), 4pp.
- JMA (Japan Meteorological Agency) (2009). Explanation Table of JMA Seismic Intensity Scale, <http://www.jma.go.jp/jma/kishou/known/shindo/kaisetsu.html> (in Japanese).
- Kaka, S.-L., and Atkinson, G. M. (2004). Relationships between instrumental ground-motion parameters and Modified Mercalli intensity in Eastern North America. *Bull. Seism. Soc. Am.*, **94**, 1728-1736.
- Karnik, V. (1969). Seismicity of the European area. Part I. *Reidel Publishing Company*, Dordrecht, 364 pp.
- Kövesligethy, R. (1906). A makroszeizmikus rengések feldolgozása, *Math. És Természettudományi Értesítő*, **24**, 349-368.
- Kövesligethy, R. (1907). Seismischer Stärkegrad und Intensität der Beben. *Gerlands Beitr. zur Geoph.*, **8**, 363-366.
- Levret, A., Cushing, M., and Peyridieu, G. (1996). Etude des caractéristiques de séismes historique en France: Atlas de 140 cartes macroseismiques, IPSN, Fontenay-Aux-Roses.
- Medvedev, S. V. (1962). Engineering seismology (in Russian). *Academy of Sciences, Inst. of Physics of the Earth*, Publ. house for literature on Civil Engineering, Architecture and Building Materials, Moscow.
- Musson, R. M. W. (1996). Determination of parameters for historical British earthquakes, *Annali di Geofisica*, **39**(5), 1041-1048.

- Musson, R. M. W. (2002). Historical earthquakes of the British Isles. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam*, 803-806.
- Musson, R. M. W. (2006). Automatic assessment of EMS-98 intensities. *British Geological Survey, Technical Report IR/06/048*.
- Musson, R. M. W., and Jiménez, M.-J. (2008). Macroseismic estimation of earthquake parameters. *NERIES project report, Module NA4, Deliverable D3, Edinburgh*.
- Musson, R. M. W., Grünthal, G. and Stucchi, M. (2010). The comparison of macroseismic intensity scales. *J. Seismol.*, **14**, 413-428.
- Levret, A., Cushing, M., and Peyridieu, G. (1996). Etude des caractéristiques de séismes historique en France: Atlas de 140 cartes macroseismiques, IPSN, Fontenay-Aux-Roses.
- Medvedev, S. V. (1962). Engineering seismology (in Russian). *Academy of Sciences, Inst. of Physics of the Earth*, Publ. house for literature on Civil Engineering, Architecture and Building Materials, Moscow.
- Musson, R. M. W. (1996). Determination of parameters for historical British earthquakes, *Annali di Geofisica*, **39**(5), 1041-1048.
- Musson, R. M. W. (2002). Historical earthquakes of the British Isles. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam*, 803-806.
- Musson, R. M. W. (2006). Automatic assessment of EMS-98 intensities. *British Geological Survey, Technical Report IR/06/048*.
- Musson, R. M. W., and Jiménez, M.-J. (2008). Macroseismic estimation of earthquake parameters. *NERIES project report, Module NA4, Deliverable D3, Edinburgh*.
- Musson, R. M. W., Grünthal, G. and Stucchi, M. (2010). The comparison of macroseismic intensity scales. *J. Seismol.*, **14**, 413-428.
- Omori, F. (1900). Intensity scale. *Earthquake Investigation Committee publication in foreign languages*, vol 4, 137-141.
- Peruzza, L. (1992). Procedure of macroseismic epicentre evaluation for seismic hazard purposes. In: *Proceedings of XXIII General Assembly of the ESC*, Prague, 434-437
- Richter, C. F. (1958). Elementary seismology. *W. H. Freeman and Company*, San Francisco and London, viii + 768 pp.
- Sarconi, M. (1784). Istoria dei fenomeni del tremoto avvenuto nelle Calabrie, e nel Valdemone nell'anno 1783. *Luce dalla Reale Accademia della Scienze, e delle Belle Lettere di Napoli*, Naples.
- Schenkova, Z., Schenk, V., Kalogeras, I., Pichl, R., Kottbauer, P., Papatsimba, C., Panopoulou, G. (2007). Iseismal maps drawing by the kriging method. *J. Seismol.*, **11**, 345-353.
- Shebalin, N.V. (1973). Macroseismic data as information on source parameters of large earthquakes. *Phys. Earth Planet. Int.*, **6**, 316-323.
- Sieberg, A. (1912). Über die makroseismische Bestimmung der Erdbebenstärke. *Gerl. Beitr. Geophys.*, **11**, 227-239.

- Sieberg, A. (1923). Geologische, physikalische und angewandte Erdbebenkunde. *Verlag Gustav Fischer*, Jena, 572 S.
- Sieberg, A. (1932). Geologie der Erdbeben. *Handbuch der Geophysik*, Gebrüder Bornträger, Berlin, vol. 2, pt 4, 550-555.
- Sponheuer, W. (1960). Methoden zur Herdtiefenbestimmung in der Makroseismik, *Freiberger Forschungshefte C88*, Akademie Verlag, Berlin.
- Sponheuer, W., and Karnik, V. (1964). Neue seismische Skala, In: Sponheuer, W., (Ed.), Proc. 7th Symposium of the ESC, Jena, 24-30 Sept. 1962, *Veröff. Inst. f. Bodendyn. u. Erdbebenforsch. Jena d. Deutschen Akad. d. Wiss.*, **No 77**, 69-76.
- Stover, C. W., and Coffman, J. L. (1993). Seismicity of the United States, 1568-1989 (Revised). *United States Government Printing Office*, Washington.
- Wald, D. J., Quitoriano, V., Dengler, L. A., and Dewey, J. W. (1999). Utilisation of the Internet for rapid community intensity maps. *Seism. Res. Let.*, **70**, 680-697.
- Willmore, P. L. (Ed.) (1979). Manual of Seismological Observatory Practice. *World Data Center A for Solid Earth Geophysics*, Report **SE-20**, September 1979, Boulder, Colorado, 165 pp.
- Wood, H. O., and Neumann, F. (1931). Modified Mercalli intensity scale of 1931. *Bull. Seism. Soc. Am.*, **21**, 277-283.
- Wu, Y.-M., Teng, T.-L., Shin, T.-C., and Hsaio, N.-C. (2003). Relationship between peak ground acceleration, peak ground velocity, and intensity in Taiwan. *Bull. Seism. Soc. Am.*, **93**, 386-396.

# Chapter 13

## Volcano Seismology

(Version August 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch13)

Joachim Wassermann

Geophysikalisches Observatorium der Ludwig-Maximilians Universität München, Ludwigshöhe 8, D-82256  
Fürstenfeldbruck, Germany,  
Tel.: +49 (0)89 2180 73962, E-mail: [j.wassermann@lmu.de](mailto:j.wassermann@lmu.de)

|                                                                                  | page |
|----------------------------------------------------------------------------------|------|
| <b>13.1 Introduction</b>                                                         | 2    |
| 13.1.1 Why a different chapter?                                                  | 3    |
| 13.1.2 Why use seismology when forecasting volcanic eruptions?                   | 3    |
| 13.1.3 Why is a revised version needed?                                          | 4    |
| <b>13.2 Classification and source models of volcano-seismic signals</b>          | 4    |
| 13.2.1 Transient volcano-seismic signals                                         | 5    |
| 13.2.1.1 Volcanic-Tectonic events (deep and shallow)                             | 5    |
| 13.2.1.2 Low-Frequency events                                                    | 9    |
| 13.2.1.3 Hybrid events, Multi-Phases events                                      | 12   |
| 13.2.1.4 Explosion quakes, very-low-frequency events, ultra-low-frequency events | 13   |
| 13.2.2 Continuous volcanic-seismic signals                                       | 18   |
| 13.2.2.1 Volcanic tremor (low-viscous two-phase flow and eruption tremor)        | 19   |
| 13.2.2.2 Volcanic tremor (high-viscous - resonating gas phase)                   | 21   |
| 13.2.2.3 Surface processes                                                       | 23   |
| 13.2.3 Special note on noise                                                     | 24   |
| <b>13.3 Design of a monitoring network</b>                                       | 25   |
| 13.3.1 Station site selection                                                    | 25   |
| 13.3.2 Station distribution                                                      | 26   |
| 13.3.3 Seismic arrays in volcano monitoring                                      | 26   |
| 13.3.4 Arrays of seismic arrays                                                  | 27   |
| <b>13.4 Analysis and interpretation</b>                                          | 28   |
| 13.4.1 One-component single station                                              | 29   |
| 13.4.1.1 Spectral analysis                                                       | 29   |
| 13.4.1.2 Envelope, RSAM and cumulative amplitude measurements                    | 30   |
| 13.4.1.3 Failure Forecast Method (FFM)                                           | 33   |
| 13.4.2 Three-component single station                                            | 34   |
| 13.4.2.1 Polarization                                                            | 35   |
| 13.4.2.2 Polarization filters                                                    | 35   |
| 13.4.3 Network                                                                   | 37   |
| 13.4.3.1 Hypocenter determination by travel-time differences                     | 37   |
| 13.4.3.2 Cross-correlation of diffuse wavefields                                 | 38   |
| 13.4.3.3 Amplitude - distance based hypocenter determination algorithm           | 39   |
| 13.4.4 Seismic arrays                                                            | 42   |
| 13.4.4.1 f-k beamforming                                                         | 43   |
| 13.4.4.2 Array polarization                                                      | 45   |

|                                                          |    |
|----------------------------------------------------------|----|
| 13.4.4.3 Hypocenter determination using seismic arrays   | 45 |
| 13.4.5 Automatic classification/detection                | 46 |
| 13.4.6 Moment tensor inversion                           | 47 |
| 13.4.6.1 Computation of Green's function in 3D           | 48 |
| 13.4.6.2 Source modelling of complex processes           | 49 |
| 13.4.6.3 Limitations of Moment Tensor Inversion          | 50 |
| 13.4.7 Automatic analysis and databases                  | 51 |
| <b>13.5 Other monitoring techniques</b>                  | 52 |
| 13.5.1 Infra sound                                       | 52 |
| 13.5.2 Ground deformation (D-GPS, P-InSAR, strain, tilt) | 54 |
| 13.5.3 Micro-Gravimetry                                  | 57 |
| 13.5.4 Gas monitoring                                    | 57 |
| 13.5.5 Thermal and video recording                       | 59 |
| 13.5.6 Meteorological parameters                         | 59 |
| <b>13.6 Technical considerations</b>                     | 60 |
| 13.6.1 Site and vault                                    | 60 |
| 13.6.2 Sensors and digitizers                            | 61 |
| 13.6.3 Digital telemetry and satellite communication     | 62 |
| 13.6.4 Power considerations                              | 64 |
| 13.6.5 Data center                                       | 65 |
| <b>Acknowledgments</b>                                   | 67 |
| <b>Recommended overview readings</b>                     | 67 |
| <b>References</b>                                        | 68 |

## 13.1 Introduction

Volcanic eruptions and their impact on human society, following earthquakes and meteorological disasters, are the most severe natural hazards. Since the pioneering works of Omori (1911), Sassa (1936) and Imbo (1954), much attention was focused on the seismic signals preceding or accompanying a volcanic eruption. Soon after the start of more extensive seismic monitoring it became clear that volcanoes show a variety of different seismic signals, which often differ from those produced by common tectonic earthquake sources, i.e., double couple type sources.

Starting with the availability of small portable seismographs in the late 1960's to early 1970's, a tremendous number of observations were made at different volcanoes and during different stages of activity. At the same time, first attempts were made to explain some of the seismic signals recorded and to classify the different signals by their proposed (but still mostly unknown) source mechanisms. Following this very enthusiastic period the progress in the study of accelerated magma transport to the surface was stagnating. Too many open questions remained unsolved, such as the mostly unknown source mechanisms of volcanic signals, the influence of the topography of volcanoes, the problem of proper hypocenter determination, the relationship between the occurrence of seismic signals of different type and the associated surface activity of a volcano. Since the late 1980's to early 1990's the use of portable and robust broadband seismometers and newly developed low power consuming 24bit A/D converters as well as the extensive use of seismic array techniques opened new horizons and different views onto the source mechanisms and the importance of volcano-seismic signals in the framework of early warning.

This chapter should be seen as a guideline for establishing a seismic monitoring network or at least a temporary experiment at an active volcano. Because of the large number of different volcanoes and many different kinds of source mechanisms, which may produce seismic signals, a description of all aspects is not possible. Also, a comprehensive review of case studies, including the variety of volcanic earthquake sequences, is beyond the scope of this paper. Relevant references include the excellent text books *Encyclopedia of Volcanoes* (Sigurdsson, 2000), *Monitoring and Mitigation of Volcano Hazards* (Scarpa and Tilling, 1996) and the comprehensive article of McNutt (2005). Most of the relevant topics dealt with in these text books are summarized below.

### **13.1.1 Why a different chapter?**

Volcano seismology uses many terms and methods known in earthquake seismology. This is no surprise as the same instruments and the same mechanism of elastic wave propagation through the Earth are used to investigate the subsurface structure and the activity state of a volcano. However, there are some deviations from conventional earthquake seismology, both in the physics of the signals and the methods of analyzing them. As outlined below, the signals vary from “earthquake-like” transients to long-lasting and continuous “tremor” signals. The most striking differences between earthquake and volcano seismology are the proposed source mechanisms and the related analysis techniques. In Section 13.2 and Section 13.4 we will discuss some of these aspects.

When setting up an earthquake monitoring network an optimal station coverage is needed in order to locate the events precisely. Depending on the tasks of the network, at least some stations should be located as close as possible to the active volcanic area in order to model the related seismic source with sufficient accuracy and determine the source depth. Hence, we are looking for a site-distribution which optimizes the station coverage and minimizes the influence of shallow structure and topography of the Earth. In contrast, in volcano-seismology we are left with sometimes very rough topography and nearly unknown propagation and site properties of the medium. Some of these aspects will be discussed in Section 13.3 and Section 13.4.

### **13.1.2 Why use seismology when forecasting volcanic eruptions?**

The use of seismological observations in the monitoring and forecasting of volcanic eruptions is justified because nearly all seismically monitored volcanic eruptions have been accompanied by some sort of seismic anomaly. The Pinatubo 1991 (Pinatubo Volcano Observatory Team, 1991), the Hekla 2000 (<http://hraun.vedur.is/ja/englishweb/heklanews.html#strain>) and the 2010 Mt. Merapi eruption ([http://www.merapi.bgl.esdm.go.id/aktivitas\\_merapi.php?page=aktivitas-merapi&subpage=seismisitas](http://www.merapi.bgl.esdm.go.id/aktivitas_merapi.php?page=aktivitas-merapi&subpage=seismisitas)) are examples of successful long- and short-term eruption forecasts made by mainly seismic observations. For further case studies on volcanic “early warning” see the comprehensive articles by McNutt (1996, 2000a, 2000b).

While most of these “early warnings” were simply deduced by counting the number and type of volcanic events per hour or day or even better by monitoring their hypocenter distributions, the physical meaning of the different seismic events and their relationship to the fast ascending magma are not yet well understood. To give an example: increasing volcanic

tremor is always a sign of high volcanic activity, but although the occurrence of tremor will increase the alert level, its role for short-term prediction is still not known with sufficient precision because we do not know the related physical process of this signal (fluid flow; movement of magma, water and/or gas; crack extension etc.). Furthermore, how can we distinguish between an intrusion and a developing eruption, both of which generate a large number of seismic signals and often share similar characteristics?

The extensive use of seismic methods during the last decades also showed that using them alone will not help the improvement of our knowledge about the internal processes of rapid magma ascent. This will be discussed in more detail in subchapter Section 13.5. Planning a new monitoring network or a short-term seismic experiment, we must also keep in mind that every volcano has its own characteristics, both with respect to seismic signal generation and wave propagation effects.

### **13.1.3 Why is a revised version needed?**

Since the writing of the first version of this chapter in 2002, much has changed both in the understanding of volcano-induced seismic signals as well as regarding technical aspects in recording the associated signals. The most dramatic changes appeared in the interpretation of the source mechanisms and the propagation of seismic waves at active volcanoes. This increased knowledge is mainly attributed to the availability of efficient numerical algorithms for simulating the signal generation as well as seismic wave propagation in complex media and by the increasing amount of high performance, parallel computing facilities. Another rising field in interpretation of volcanic unrest can be seen in the increasing number of applications in the field of remote sensing.

Both topics are far too large to be covered in detail in this manual. Especially the high performance computing part requires a lot of investment, which will not easily be covered especially in countries with multiple natural hazards and often small national budget. Therefore we will only briefly discuss the most important outcome of this expensive modeling.

While leaving the classification part nearly untouched, most of the changes in the revised version will be made in the source model part of Section 13.2 and some newly introduced analysis techniques in Section 13.4. Another major change is made in Section 13.5 where other monitoring techniques than seismology are briefly discussed. Here much more emphasis is given to new but already quite well established remote and ground based techniques (e.g., P-InSAR; SO<sub>2</sub> remote sensing; infra-sound measurement etc.). We finalize the revision by describing new technical developments in communication technology and software, i.e., database management systems, which are now thought to be “state of the art”.

## **13.2 Classification and source models of volcano-seismic signals**

Most of the confusion in volcano seismology is caused by the huge number of different terms for classifying volcano-seismic events. While this is mainly caused by the imperfect knowledge about the source mechanisms, we will focus on the basic nomenclature widely used in the literature. Most of these terms simply describe the appearance and frequency content of the signal, while others imply a certain source mechanism. However, one should be



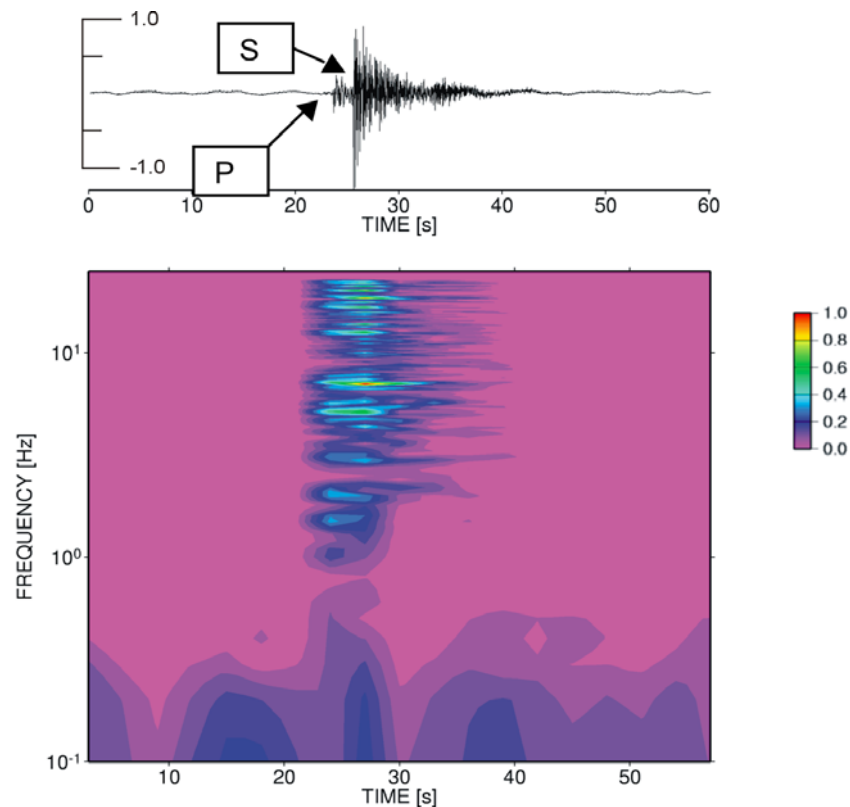
aware in both cases that the sources are often unknown and the propagation medium alters the shape and the spectral content of the signals significantly.

While pioneering work in classifying volcano-seismic signals was made by Shimozuru (1972) and Minakami (1974), most of the following discussion follows the work of McNutt (1996, 2000a) and Chouet (1996a). We will divide the known signals mainly into transient and continuous signals, even if some of these continuous signals reflect simply the superposition of discrete signals. We will also discuss, where appropriate and physically plausible, differences in the signal generation related to different types of magma (i.e., low/high viscous, gas rich/poor). In addition to the usual classification, “up-to-date” source models and interpretation of the occurrence of different signals will be given.

## 13.2.1 Transient volcano-seismic signals

### 13.2.1.1 Volcanic-Tectonic events (deep and shallow)

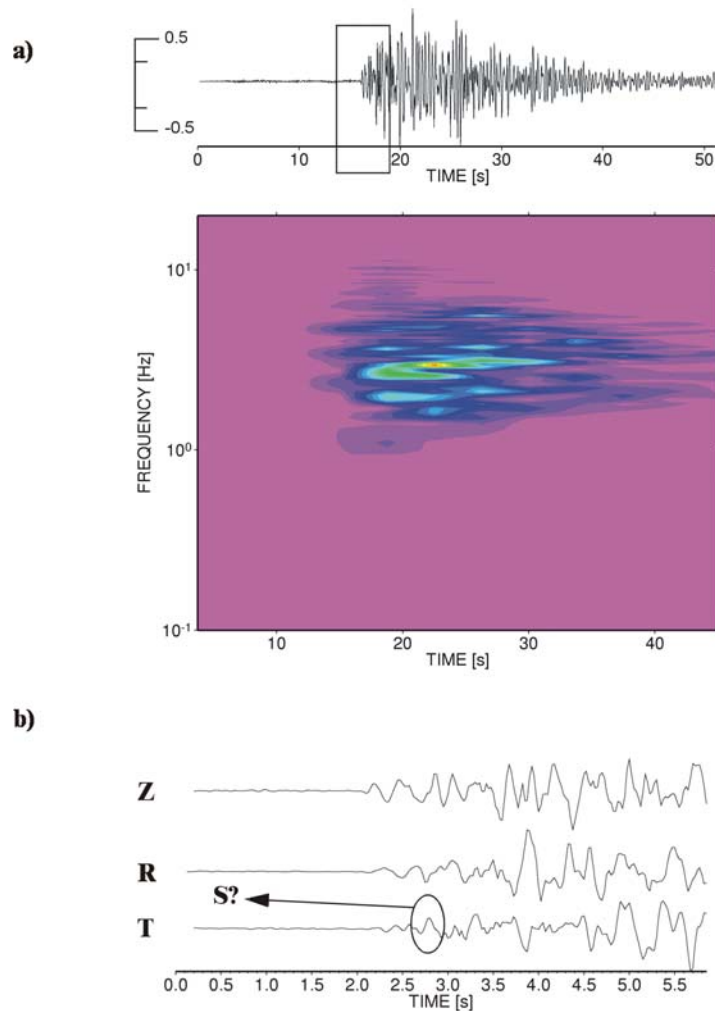
*Deep (below about 2 km) Volcanic-Tectonic* events (VT-A) manifest themselves by the clear onsets of P- and S-wave arrivals and their high frequency content ( $> 5\text{Hz}$ ). This leads also to the class name *high-frequency* event (HF) (Fig. 13.1).



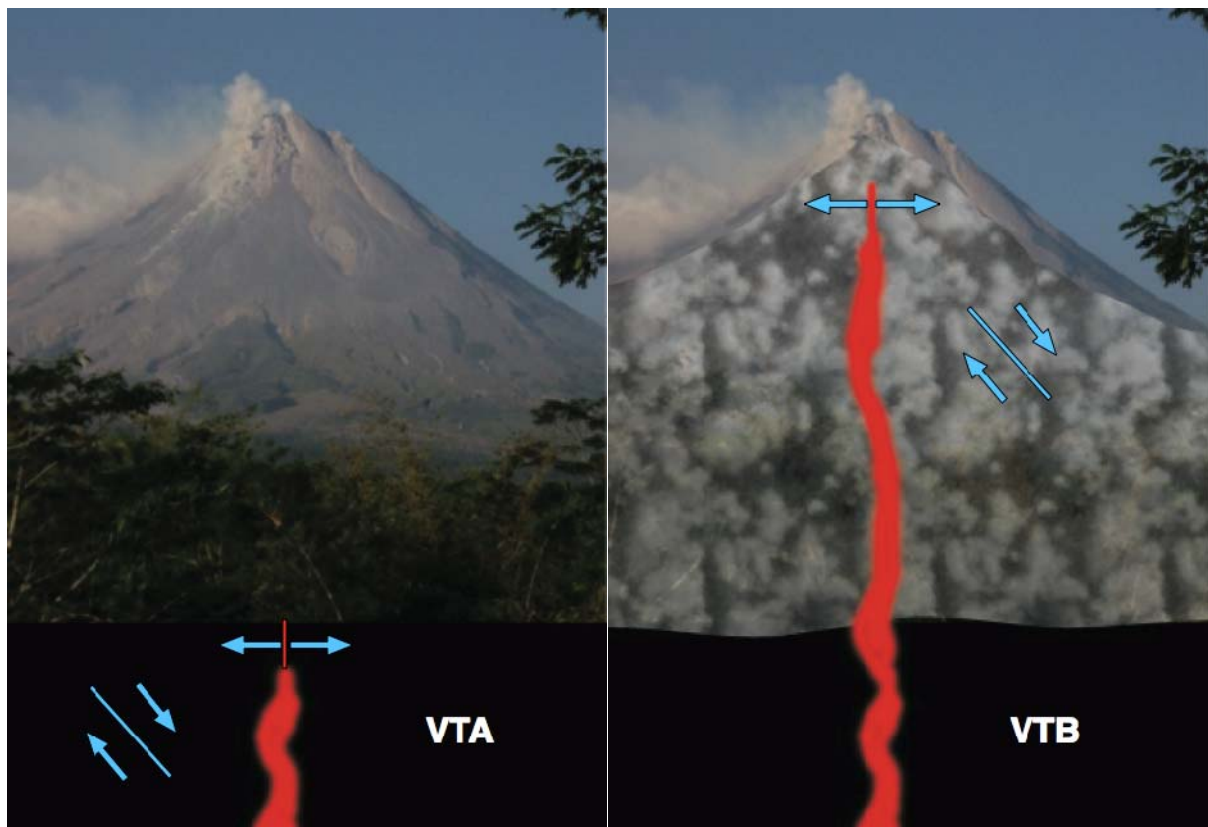
**Fig. 13.1** VT-A type event recorded at Mt. Merapi, Indonesia. The impulsive P- and S-wave arrivals are clearly visible in this signal, as well as their high-frequency content and short signal duration. The given color coding, representing normalized amplitude spectral density, is valid for all following figures. Also the amplitudes of the corresponding amplitude-time representation are in arbitrary units and when given are normalized to the amplitude shown in this figure.

The name of this event type implies a well known source mechanism, namely a common shear failure caused by stress buildup and resulting in slip on a fault plane similar to a tectonic earthquake source. The only difference from the latter is the frequent occurrence of swarms of VT events, which do not follow the usual main-after-shock distribution (McNutt, 2000a) and are located within a volcanic edifice. An earthquake swarm is a sequence where the largest events are similar in size and not necessarily at the beginning of the sequence. The high frequencies and the impulsiveness of the P- and S-wave arrivals seems to be caused by low scattering due to the short travel path through high scattering regions and low attenuation.

In contrast, *shallow (above about 1-2 km) Volcanic-Tectonic* events (VT-B) show much more emergent P-wave onsets and sometimes it is even impossible to detect any clear S-wave arrival (Fig. 13.2). The spectral bands are shifted to lower frequencies (1-5 Hz). Both observations are thought to be caused by a shallow hypocenter location and therefore a larger amount of scattering during wave propagation, especially of higher frequencies.



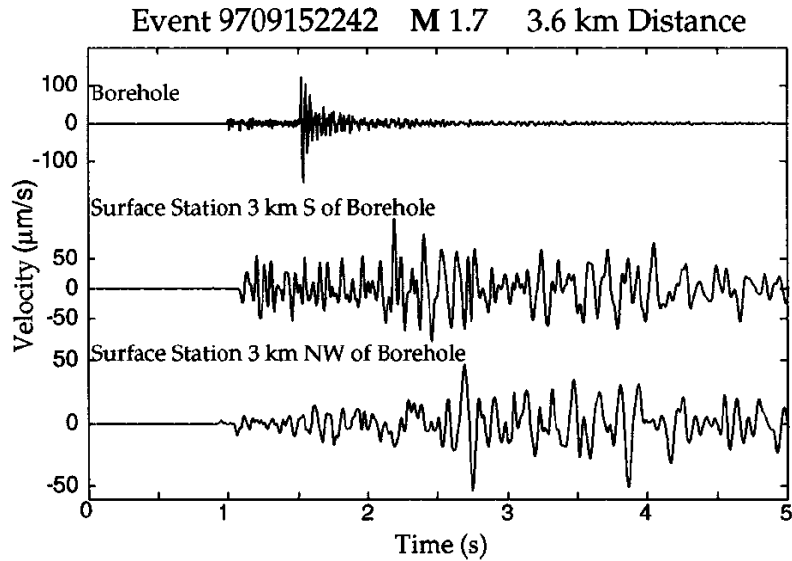
**Fig. 13.2** **a)** Typical example of a VT-B type event recorded during a high activity phase at Mt. Merapi. Note that the overall frequency content is mainly between 1 – 10 Hz with a dominant frequency at roughly 3 Hz. **b)** zoomed-in version of all three components of the same event. Whereas the P-wave arrival is clearly visible, no clear S-wave arrival can be seen. The circle marks the wavelet that has the approximate S-wave travel time for the estimated source location.



**Fig. 13.3** Two different models describing the occurrence of VT –A/B events when a volcanic crisis is eminent. While some VT events are simply caused by reactivation of existing faults by the increasing stress due to magma ascent, the other possible mechanism is directly linked to the emplacement of the magma in the feeder system and the associated tensile faulting. In this picture the difference between VT-A and VT-B is simply their location with respect to a high scattering region.

#### *Source models and interpretation*

Recently, detailed studies showed that the sources of some VT events deviate significantly from that of a pure shear failure, but show some similarities with the later described *Low-Frequency* events. An indication for location-dependent waveform changes is shown in Fig. 13.4. While the borehole station shows a seismogram which is typical for a common double couple source, the station at the surface within a volcanic caldera shows no visible S-wave onset, a lower dominant frequency but a very long coda. However, even using very sophisticated Moment-Tensor inversion techniques, which includes 3D Greens function computation (see Section 13.4.5) it is still not possible to resolve the ambiguity in estimating the “true” source mechanism of this class of signals. While there is a general agreement that the dominant part of the source process can be modeled by a double couple (DC), the role and amount of non-DC contributions remains debatable. Several papers on the inversion of the seismic moment tensor showed a significant contribution of non-double couple parts (Dahm and Brandsdottir, 1997; Saraò et al., 2001). The reason for the imperfect inversion will be discussed in more detail in Section 13.4.5.



**Fig. 13.4** Comparison of vertical component seismograms recorded in a borehole with records from the two surface stations nearest the borehole at the volcanic area of Long Valley Caldera, California. (from: Prejean and Ellsworth, 2001)

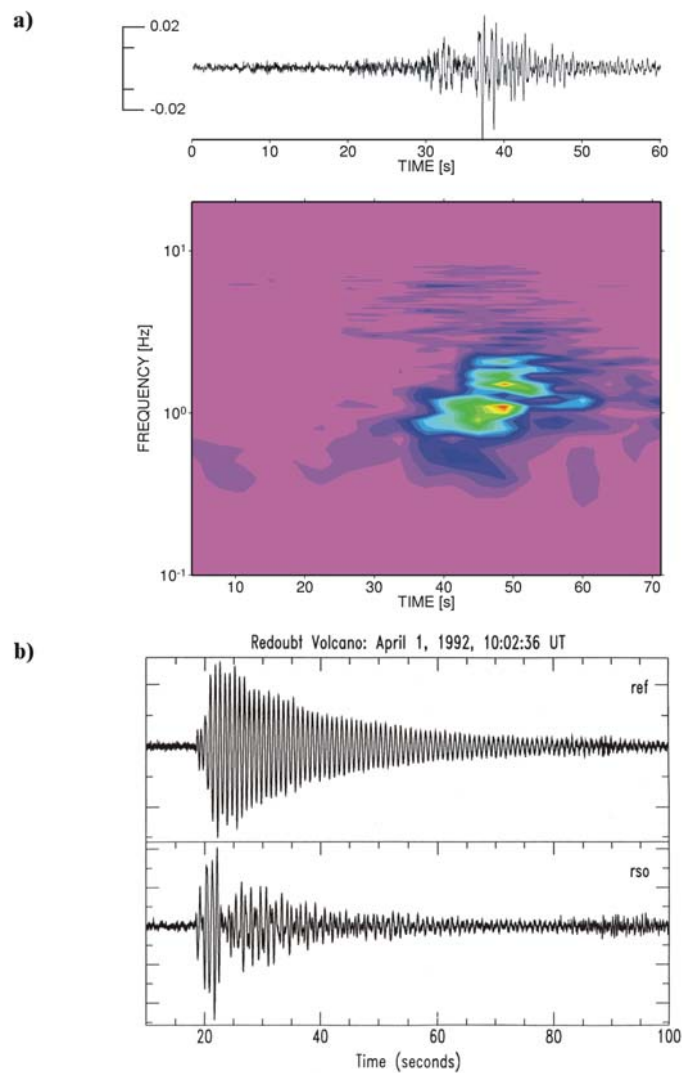
The main benefit in recording and analyzing VT-type earthquakes originates from the assumption that rising magma is acting as an additional stress source. This additional stress possibly superimposed by a regional stress field either leads to tensile faulting when magma is breaking the host rock or to earthquakes on preexisting faults in the country rock as a reaction to additional stress (see Fig. 13.3). While the first is directly linked to rising magma, the second possibility is a more passive and blurred image of magma inflation. Roman et al. (2006) showed that estimating fault plane solutions (see Section 3.4 in Chapter 3) gives important information of the current state of an active volcano and might, under certain conditions, serve as an important early warning tool. In their work Roman et al. (2006) showed that deep VT-A type events at Montserrat indicate a changing pattern of the P- and T-axis before a new eruption cycle starts. In their study the authors used a simple fault plane solution program (FPFIT, Reasenber and Oppenheimer, 1985) and thus neglect a more realistic velocity and topography. The estimated pattern, however, is striking and while planning a surveillance procedure, fault plane solutions should be considered as an important indicator of changing volcanic activity.

Recently found evidence in laboratory experiments (Lavalee et al., 2008) and in field observations (Tuffen et al., 2003) point towards a source region, which lies in the rising magma itself. Depending on the applied strain rate viscous magma might change its rheologic behavior from visco-plastic to brittle. Within this model, tracking the sources of VT events makes a direct estimate of the fluid path and rate possible.

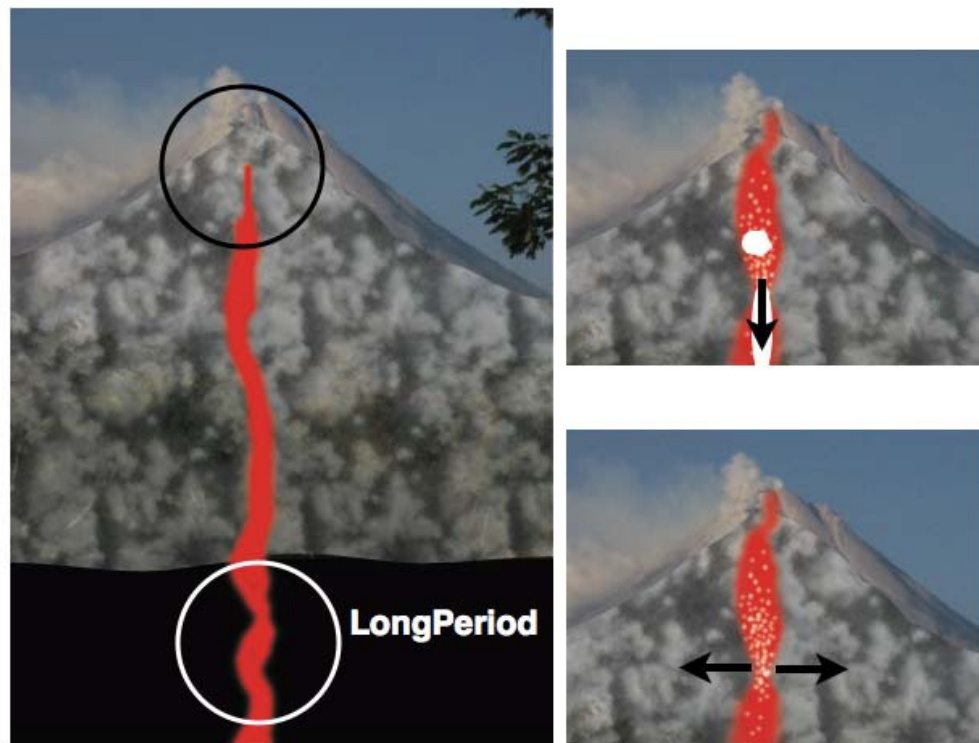
A very distinct feature of VT events is their swarm like occurrence. McNutt (2002) pointed out that the main duration of VT-swarms is about 5 days. At several volcanoes the VT-A and VT-B type events happen several days to weeks prior a volcanic eruption. Whatsoever, it must be emphasized that the majority of magma ascent events, which on one hand are accompanied also by VT events, results in intrusions and will, on the other hand, not be associated with any surface, i.e., eruptive activity. This ambiguity is one major source of misclassification of the current state of a volcano.

### 13.2.1.2 Low-Frequency events

*Low-Frequency* events (LF or Long Period - LP) usually show no S-wave arrivals but a very emergent signal onset (Fig. 13.5). The frequency content is mostly restricted to a narrow band between 0.2-10 Hz. In the still rare cases in which the location of LF sources are determined, they are often situated in the shallow part of the volcano ( $< 2$  km). Locations are deduced mainly by amplitude distance curves, from the rare hypocentral determinations using clear first onset recordings, and by semblance location techniques from particle motions recorded on a broad-band seismometer network (Kawakatsu et al., 2000). Some volcanoes (e.g., Kilauea) are known to produce deep (30-40 km) LF events (Aki and Koyanagi, 1981; Shaw and Chouet, 1991). Fig. 13.6 depicts different models to described two of the proposed source mechanisms of LP –events.



**Fig. 13.5** **a)** example of a LF-wave group recorded at Mt. Merapi. Clearly the dominant frequency is around 1 Hz. **b)** shows an example of a LF event recorded at two different sites located at Redoubt volcano, Alaska (courtesy of S. McNutt, Alaska Volcano Observatory; AVO). The spindle shaped signal is also known as Tornillo.



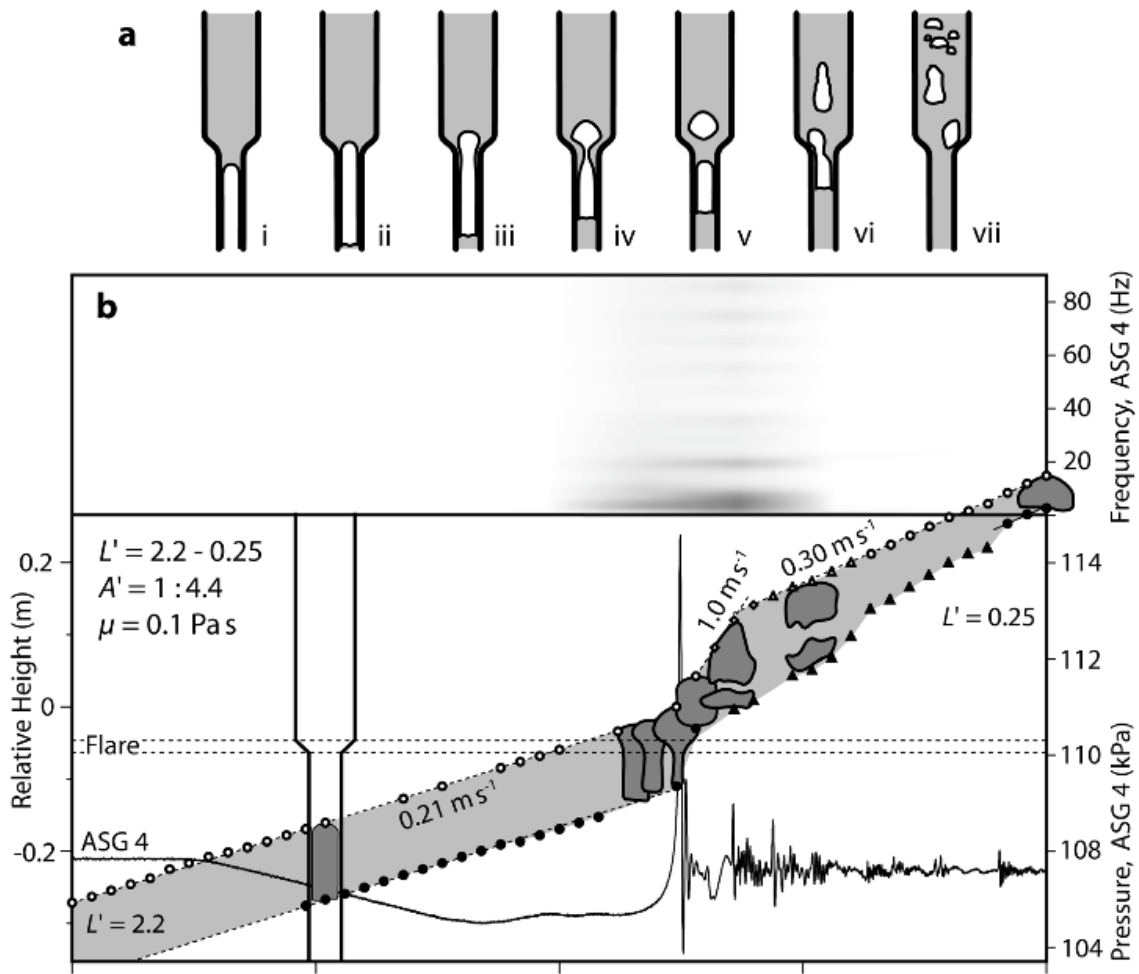
**Fig. 13.6** Two different models to describe two of the proposed source mechanisms of LP – events. While in deeper parts of the volcanic edifice an oscillating crack model is the most appropriate source model, the signature of LP events in shallower parts can be explained best by pressure disturbances, caused by an ascending gas bubble within a (moving) fluid.

#### *Source models and interpretation*

The associated source models of LP events range from an opening and resonating crack when the magma is ascending towards the surface (Chouet, 1996a) to the existence of pressure transients within the low viscous fluid-gas mixture causing resonance phenomena within the magma itself (Seidl et al., 1981 see Fig 13.6). Both models are able to explain large parts of the observed features in the time as well in the spectral domain but leave the physics and the trigger mechanism unresolved. Papers by James and others (2006, 2009) show that the source process of low frequency events within low viscous magma can be modeled by pressure transients within the magma caused by changing geometry of the dike system (Fig. 13.7). This pressure transient in turn can cause an oscillation in the fluid.

Neuberg et al. (2006) and Collier and Neuberg (2006) propose a very detailed source and trigger mechanism for low frequency events in a more viscous magma environment. These authors argue that most of the magma ascent happens aseismically up to a certain depth level is reached where large portions of gas bubbles are formed. Within the magma column a strong gradient of viscosity is assumed, caused by the cooling of the outer layer of magma at the conduit wall. At a certain depth the transition from no slip to friction controlled slip occurs when reaching the brittle failure condition (Dingwell and Webb, 1990). After cracks and veins are created, gas can now escape and the ascent is then dominated by friction controlled slip which is aseismic (Collier and Neuberg, 2006).





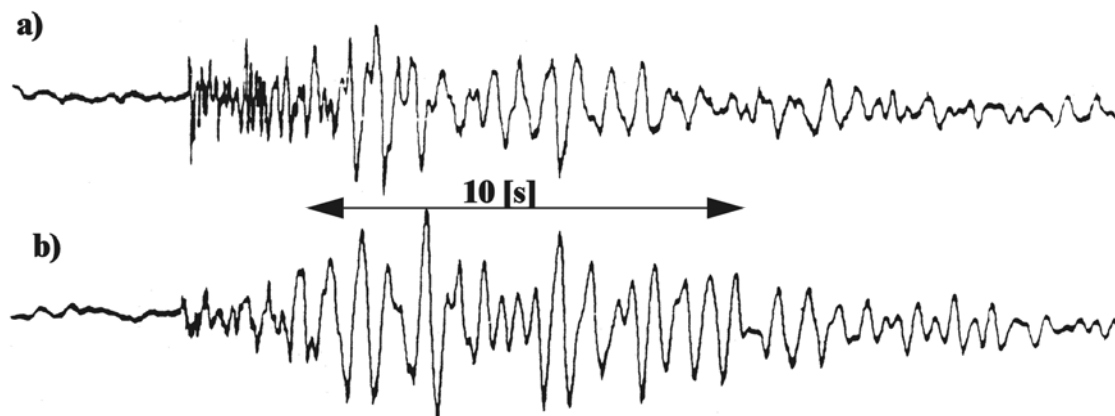
**Fig. 13.7** Experimental setup and results of an ascending slug passing through a widening conduit. A clear pressure perturbation can be detected in the fluid and the accelerating slug may be also an effective mechanism for generating elastic waves in the surrounding subsurface structure (from: James et al., 2006).

An alternative mechanism for triggering LP or LF events is given by Denlinger and Hoblitt (1999). These authors favor a supply rate controlled stick-slip mechanism, which is deduced from observations of technical polymer melts. Both proposed trigger mechanisms are able to explain the frequently observed timely very regular repetition of LF events, which may lead even to a continuous signal (see Section 13.2.2.2). The model of Neuberg et al. (2006) and Collier and Neuberg (2006) is also able to explain the frequently observed similarity of waveforms in the recorded data. This is a surprising feature by its own as a volcano is a very heterogeneous geological structure making scattering distortion a very dominant feature. Numerical simulations by Sturton and Neuberg (2006) and Neuberg et al. (2000) show that the overall shape of the seismograms and their frequency content can simply be explained by the excitation of multiple reflected borehole (i.e., Stonely) waves at the magma-conduit interface.

In the context of using seismology as an early warning tool in volcanology the detection and analysis of LF/LP events is a very important tool, because of their direct connection to the ascending i.e. flowing magma within the subsurface. Monitoring fluid movement or pressure disturbance directly makes early warning much more successful and effective in both timing and size estimates of an eruption. However, there is still a large ambiguity caused by intrusion rather than extrusions, i.e. eruptions.

### 13.2.1.3 Hybrid events, Multi-Phases events

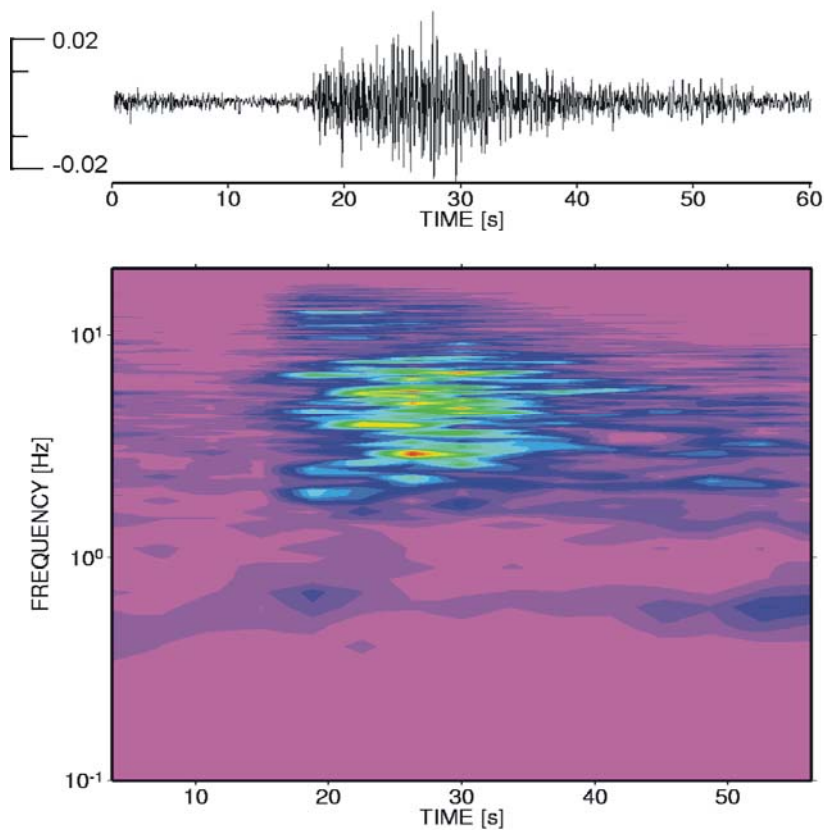
Some volcano-seismic signals share the signal and frequency characteristics of both LF and VT-(A,B) events. Signals of this class are usually labeled as *Hybrid* events, which may reflect a possible mixture of source mechanisms from both event types (see Fig. 13.8) and/or additionally reflect possible path effects (Harrington and Brodski, 2007). For example, a VT microearthquake may trigger a nearby LP event. Lahr et al. (1994) and Miller et al. (1998) detected swarms of *Hybrid* events during the high activity phase of Redoubt (Alaska) and Soufriere Hills volcano (Montserrat, West Indies), respectively. Miller et al. (1998) concluded that such events reflect very shallow activity associated with a growing dome.



**Fig. 13.8** **a)** shows a Hybrid event and **b)** a VT-B event for comparison. The higher frequencies at the beginning of the Hybrid event are an obvious feature, while the later part shows the similarity with the VT-B event (courtesy S. McNutt, AVO).

*Multi-Phase* events (MP also *Many-Phases* event; see Fig. 13.9; Shimozuru, 1972) are somewhat higher in their frequency content (3 to 8 Hz) than *Hybrid* events but are related as well to energetic dome growth at a very shallow level. The name reflect their complicated waveforms with seems to consist of multiple phases just after each other.





**Fig. 13.9** MP-event recorded at Mt. Merapi during strong dome formation. The frequency is restricted between 3 - 10 Hz and resembles that of a VT-B type event at this volcano. Note the long duration of this event whilst its amplitude is much smaller than for the VT-B event shown in Fig. 13.2.

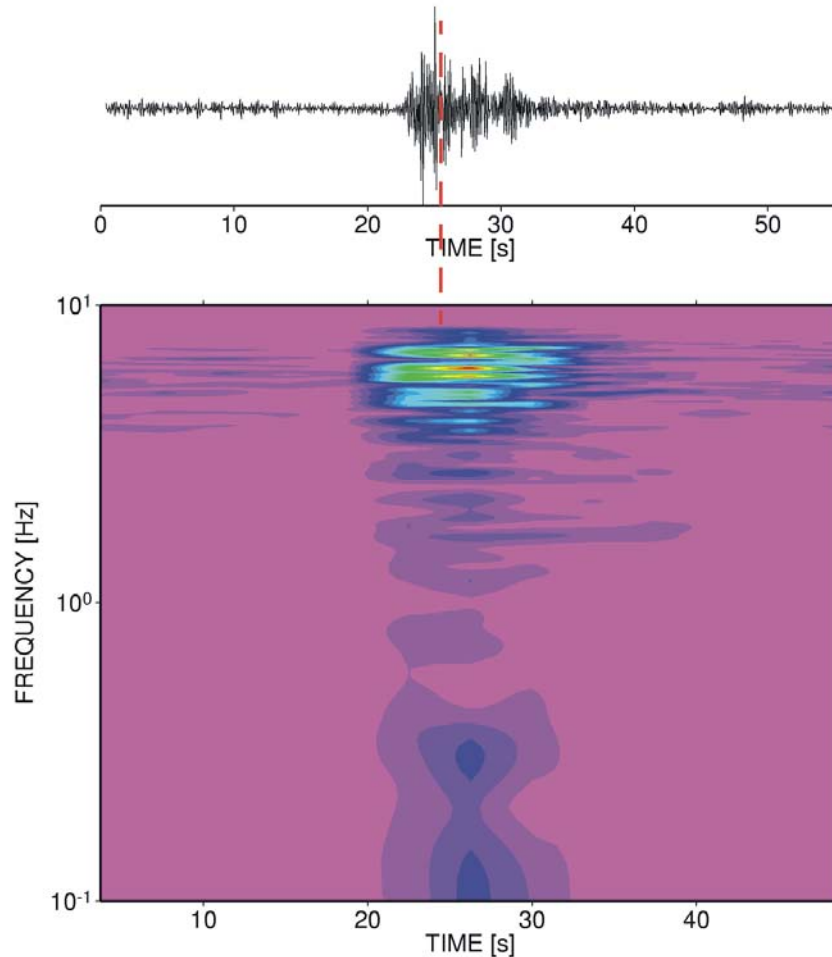
### **Source models and Interpretation**

Both types of signals and their associated mechanisms are still a topic of research as their occurrence might be a good indicator for the instability and/or the growth rate of highly viscous lava domes. The trigger or source mechanism is likely the same for LP/LF events (see Section 13.2.1.2), while the source depth is significantly shallower at least in case of MP events. Recent advances in laboratory experiments, which investigate the plastic to brittle transition, point towards a source mechanism, which is located in the moving fluid itself (Lavallée et al., 2008).

#### **13.2.1.4 Explosion quakes, very-low-frequency events, ultra-low-frequency events**

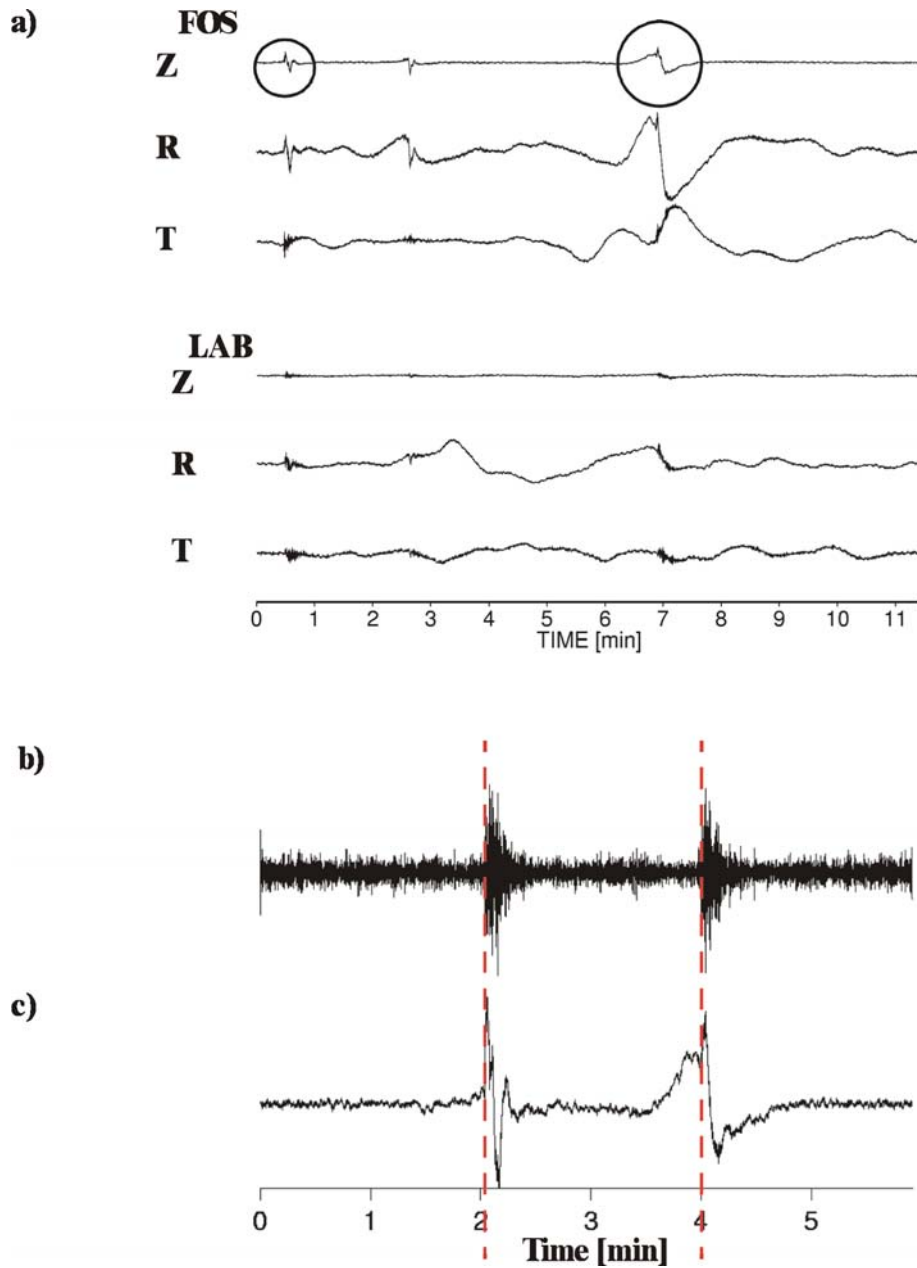
Very pronounced *ultra long period* (ULP;  $f < 0.01\text{Hz}$ ) and *very low frequency* (VLF;  $f \sim 0.01 - 0.1\text{ Hz}$ ) signals are observed and are well documented at many volcanoes in Japan and on Hawaii (e.g., Aso: Kawakatsu et al., 2000; Iwate: Nishimura et al., 2000; Kilauea: Ohminato et al., 1998) using several broadband seismometers located in the near-field to intermediate-field distance from the source. To the same of class belong the *explosion quakes*, a signal that accompanies Strombolian or other (larger) explosive eruptions, showing clear waveform characteristics. Most of these signals can be identified by the occurrence of an air-wave which is caused by the sonic boom during an explosion, when the expanding gas is accelerated at the

vent exit (see Fig. 13.10, see also Section 13.5.1). This wave mainly travels through the air with the typical speed of sound (330 m/s at 20°C). While we do not discuss the explosive mechanism at the magma air interface in detail, the source, which leads to this explosion is not yet completely deciphered and several concurrent models will be discussed later. Some LF events show the same frequency-time behavior as the explosion quakes but lack an air phase (McNutt, 1986). This might reflect a common driving source of deeper LF-events and shallower explosion quakes.

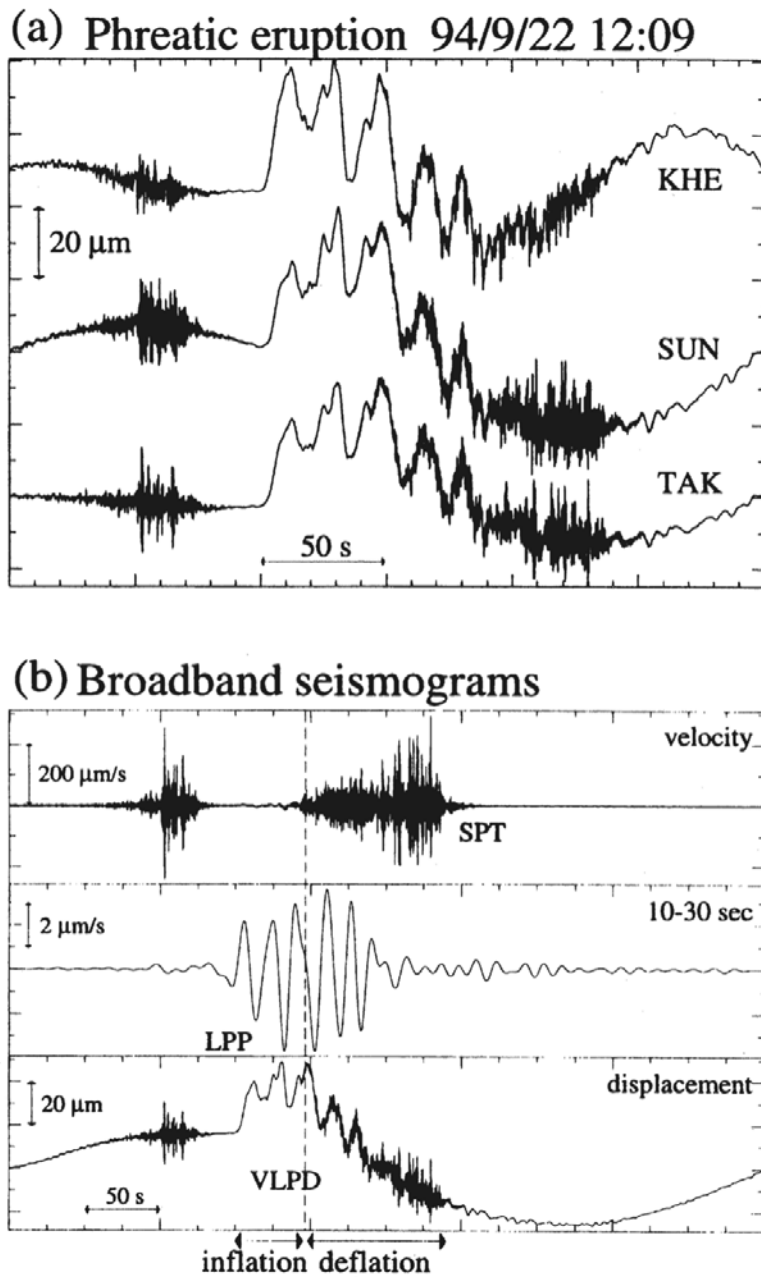


**Fig. 13.10** An explosion signal recorded at Stromboli volcano, Italy. The seismic station was located just 400 m from the active vent. The dashed line gives a rough estimate of the onset of a sonic wave also visible as high (red) amplitudes in the time-frequency plot around 5 Hz.

Since the early 1990ies portable broadband seismometers with corner frequencies as low as 0.00833 Hz shed new light on this open question (see Fig. 13.11). It could be verified that at Stromboli volcano (Italy) ULF pressure buildup takes place several minutes before the onset of a Strombolian eruption (Dreier et al., 1994; Neuberg et al., 1994; Wassermann, 1997; Kirchdörfer, 1999). As this is only visible in the near-field of the seismic sources with a geometrical spreading factor proportional to  $r^{-2}$  (implying a simple volumetric source), the seismic stations must be located close to the active vent of the volcano (see Fig. 13.11). Since the first near-field broadband observations of strombolian activity many examples at volcanoes around the globe exist nowadays, making ULF signals a common signal accompanying low to medium viscous volcanism.



**Fig. 13.11** **a)** ULP signal recorded with a Streckeisen STS2 broadband seismometer at Stromboli volcano. We removed the instrument response down to 300 s and the resulting traces are integrated to reflect ground displacement. The three uppermost traces show the three-component seismograms of a station located 400 m from the vent, whereas the lower three traces show the same but at a site located 1800 m from the active vent indicating a large signal only visible in the near-field. The two circles mark two different types of seismic signals caused by strombolian activity which were not linked to different active craters. **b)** shows the seismogram of a 1 Hz seismometer during two different explosion quakes, the dashed lines mark the visible onset of strombolian eruptions. **c)** shows the displacement signal of two different explosion quakes. Note, not all explosion signals produce the same amount of long-period displacement signals.

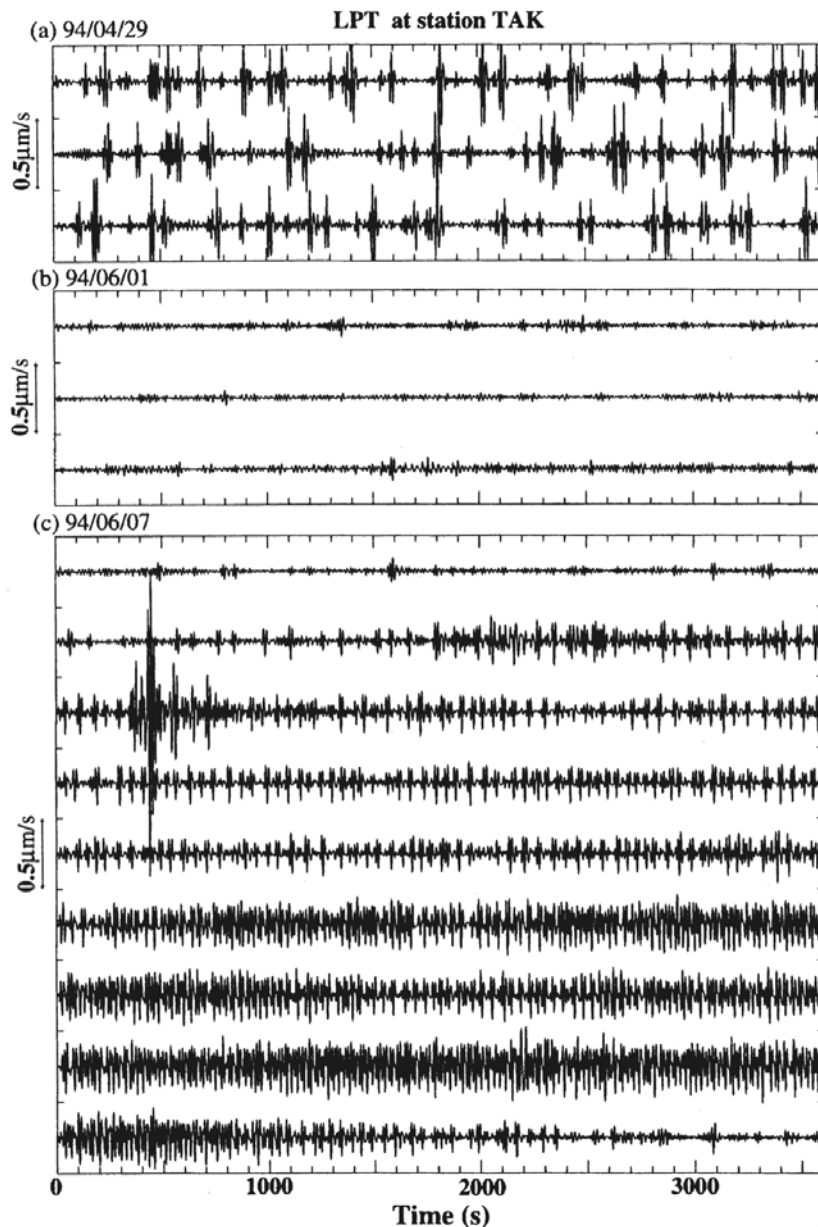


**Fig. 13.12** a) ULP (or very long-period displacement) signal observed at three broadband stations during a phreatic eruption of Aso volcano. b) original velocity, band-pass filtered velocity and displacement seismogram of the same event observed at station TAK. The vertical line in b) indicates the onset of the eruption, all stations are located within 2 km from the volcanic centre (from: Kawakatsu et al., 2000).

ULF and VLF events are still unknown at most andesitic and rhyolitic volcanoes, which possibly implies that the described slug flow (Vergnolle and Jaupart, 1990, James et al. 2006) may be non-operative at this regime of viscosities. In contrast to this statement, the work of Hidayat et al. (2000) showed that there exists a moderate (0.25 Hz) VLF signal in the near-field of some MP events recorded at Mt. Merapi (Indonesia).

### Source models and Interpretation

Since the late 1980's to early 1990's many of the ULP-VLP observations were interpreted as shallow ( $z < 1.5$  km) phreatic eruptions with a strong low frequency pressure pulse ( $f \sim 0.01$  Hz; see Fig. 13.12). At Aso volcano Kawakatsu et al. (2000) also detected a second signal with dominant frequencies roughly at 0.06 Hz at the same depth range as the modeled phreatic source. The authors classified this signal as *long period tremor* (LPT), which is seen to reflect the merging of isolated pulses into a nearly continuous signal (see Fig. 13.13 and Section 13.2.2.2: Fig. 13.17). Kawakatsu et al. (2000) interpret these signals as caused by the interaction of hot magma/fluid with an aquifer situated in 1 - 1.5 km depth below the craters of Aso volcano.



**Fig. 13.13** Vertical component broadband seismograms band-pass filtered at 0.033 to 0.1 Hz at Aso volcano during three different days 1994. The isolated ULP pulses visible in a) and b) merge together in c) forming the continuous signal of *long period tremor* (from: Kawakatsu et al., 2000).

A different model of ULP/VLP generation, which fits the visual and seismological observations very well consists of a shallow magma chamber and a tiny feeder system to the surface, which may additionally vary in size and width. The accumulation of a gas pocket and the ascent of this pocket as a gas slug explain the observed pressure buildup up to a certain precision (Vergnolle and Jaupart, 1990). However, most of the strombolian eruptions at Stromboli and other volcanoes show only small over-pressure without any visible difference in the associated surface activity. Some more recent results of Gerst (2010) will be discussed in Section 13.5.1.

Results from still rare moment tensor inversion imply a rather complicated mixture of double couple (DC), compensated linear vector dipole (CLVD) and isotropic components (IC). In order to explain the recorded low frequency seismograms some authors combine a number of cracks, cylinders and additional single forces (Chouet et al., 2003; Cesca and Dahm, 2008; Aster et al., 2008; Chouet et al., 2010) for eruption at Kilauea (Hawaii) and Stromboli volcano or more violent explosions at Mt. Erebus (Antarctica). In Section 13.4.5 we will focus on the method of moment tensor inversion and its associated problems in much more detail.

A very comprehensive physical model for mild strombolian eruptions is given by a series of papers by James et al. (2006) and James et al. (2009). In these papers the physics of an ascending large gas pocket (slug) is experimentally and mathematically investigated.

The conceptional model of James et al. (2009) combines the viscous flow of the fluid film along the conduit wall and the pressure above in the fluid head (the nose) of a rising gas slug which is ascending in a volcano's conduit. This model is able to explain the change from gas puffing to strombolian eruption as well as the observed overpressures. The controlling factor between puffing and bursting is seen in the starting pressure, the length of the slug and therefore in the involved gas mass which is transported towards the surface (James et al., 2009). While this model can explain the observed infra-sound signals it will not explain the generation of LP/VLP/ULP signals during slug ascent.

In a different paper James et al. (2006) showed that changes in the conduit diameter can act as a very prominent source of pressure transients which then are able to couple into the surroundings and produce elastic deformation signals (Fig. 13.7). While models by Schick et al. (1982) and Seidl et al. (1981) in the early 80ies already describe similar mechanisms, the paper of James et al. (2004, 2006, 2009) put these ideas in the context of a physically and mathematically more quantitative model.

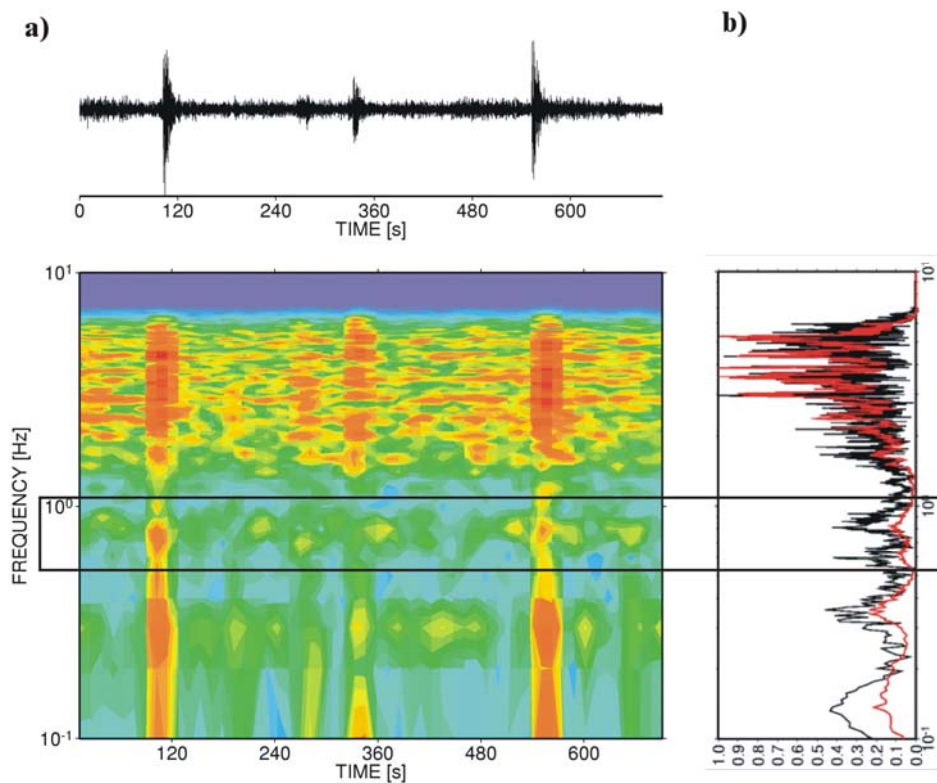
### **13.2.2 Continuous volcanic-seismic signals**

The appearance of continuous seismic signals at active volcanoes demonstrates the most profound difference between tectonic earthquake and volcano seismology. The suspected mechanisms range from obvious surface effects such as rockfall, landslide or pyroclastic density flow to internal ones such as volcanic tremor. Nearly every volcano world-wide shows the signal of volcanic tremor during different activity stages. Volcanic tremor is the most favored parameter in volcano early eruption warnings. Because of possibly differing source mechanisms we discuss tremor separately for the two flow regimes: high and low viscosity.

Recently the observation of so called non-volcanic tremor is attracting a lot attention in the seismologic community. This type of earth tremor is associated by slow, episodic slip in subduction zones, which makes it an important tool to understand the dynamical behavior of subduction related earthquakes. Sharing a lot of the signal characteristics with its volcanological counterpart, many of the methods developed in volcano-seismology are now further investigated in order to shed more light on this signal type.

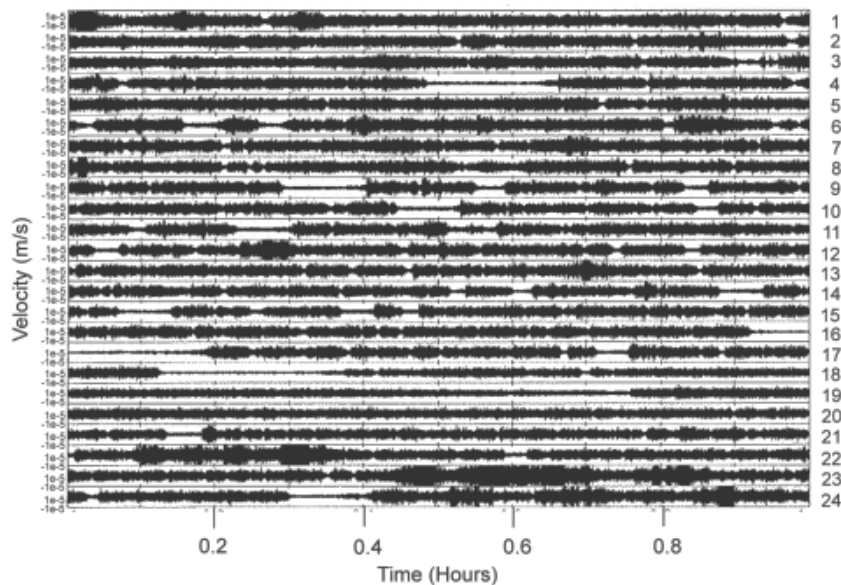
### 13.2.2.1 Volcanic tremor (low-viscous two-phase flow and eruption tremor)

Most of the monitored basaltic volcanoes show some kind of cyclic appearance of *volcanic tremor*. The tremor signals can last between minutes and months in duration and, in most of the cases, their spectra are very narrow-band (1-5 Hz; Fig. 13.14).



**Fig. 13.14** **a)** explosion signals superimposed on the continuous signal of volcanic tremor at Stromboli volcano. The box marks the frequency band of a weak but typical volcanic tremor band at Stromboli volcano. Note that the explosion quake also excites the same frequency band whereas below this frequency band the spectral amplitude of the explosion quake type signals are somewhat smaller. The tremor band with frequencies above 2.0 Hz is partially distorted by the ejected volcanic debris falling back to the surface and tumbling down the slope of the volcanic edifice (see Section 13.2.2.3). **b)** the normalized Fourier transform of a explosion quake type signal (black) and of a normalized power spectrum of six hours of continuous recording (red). While the first reflects the typical spectrum of all explosion quakes, the overall behavior of the second spectrum is mainly due to volcanic tremor. The overall similarity between the explosion quake and tremor signal types is obvious.

Some tremor signals show strong and short-pulsed amplitude variations (termed *beating tremor*), while others are nearly stationary over several days or even months. The common similarities in the spectra of volcanic tremor and LF and even explosion quake events are another important observation which has to be explained when looking for the source mechanisms. At Mt. Etna volcano (Italy), strong fluctuations and changing signal characteristics of volcanic tremor amplitude are associated with lava fountaining at one of its summit craters or after the opening of a flank fissure (Cosentino et al., 1989, Langer et al., 2011). Gottschämmer (1999) described a tremor cycle at Bromo volcano (Indonesia) where the tremor amplitude fluctuation could be visually correlated with heavy ash plumes (large amplitude - *eruption tremor*) or white steam (small tremor amplitude) episodes (see Fig. 13.15).



**Fig. 13.15** Volcanic tremor at Bromo volcano (Indonesia) during a high activity phase at the end of 1995 (courtesy of E. Gottschämmer, University of Karlsruhe). Large tremor amplitudes correlate with the eruption of heavy ash plumes while small tremor amplitudes appear during quiet steam emissions (Gottschämmer, 1999).

These observations made at different volcanoes with either low viscosity magma or a huge amount of volatiles (free or after the fragmentation of high viscosity magma; steam) suggest the involvement of gas/fluid interaction in generation of volcanic tremor. In general it must be stated that the appearance of volcanic tremor is always a sign of high volcanic unrest, which might lead to an eruption. Whether this eruption may follow or the activity terminates by an intrusion cannot be decided without other geophysical/geodetical/geological evidences. Whatsoever, even in the cases where a well established and “complete” monitoring system exist signs of unrest or precursors may still carry significant ambiguities caused by the complexity of magma flow and the physics of volcanic eruptions.

### **Source models and Interpretation**

The similarities in the overall spectral content of LF/LP events and volcanic tremor are reflected in similarities of the proposed source mechanism or of the source region (resonating fluid). Flow instability is thought to play an important role in the excitation of volcanic tremor

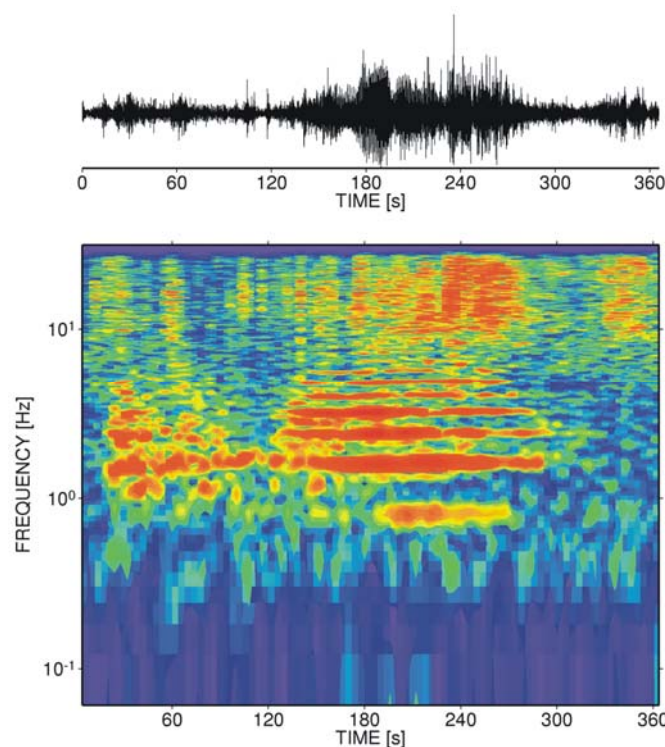


in multiple phase flow pattern (Seidl et al., 1981; Schick, 1988; James et al., 2004, 2006) and the associated LF events are seen as a transient within the same physical system. On the other hand, Chouet (1987) and Chouet (1986) state that a repeated excitation of a connected crack system could cause a harmonic and long-lasting signal, where the fluid is only passively reacting to the crack oscillations. James et al. (2006) could show, that under certain circumstances the pressure transients produced by changing geometry of the feeder system can couple into the elastic solid surrounding of the feeder system (Fig. 13.7).

The spectral content observations support both the low viscosity magma and volatile interpretations. Explosions at Stromboli volcano excite the same frequency band as volcanic tremor does, which supports the idea of a common resonating system (Fig. 13.14). However, care must be taken when interpreting the frequency spectra of volcanic tremor. Detailed studies on the spatial frequency distributions at Stromboli showed that single frequency peaks are possibly influenced, to still an unknown amount, by the propagation medium (Mohnen and Schick, 1996).

### 13.2.2.2 Volcanic tremor (high-viscous - resonating gas phase)

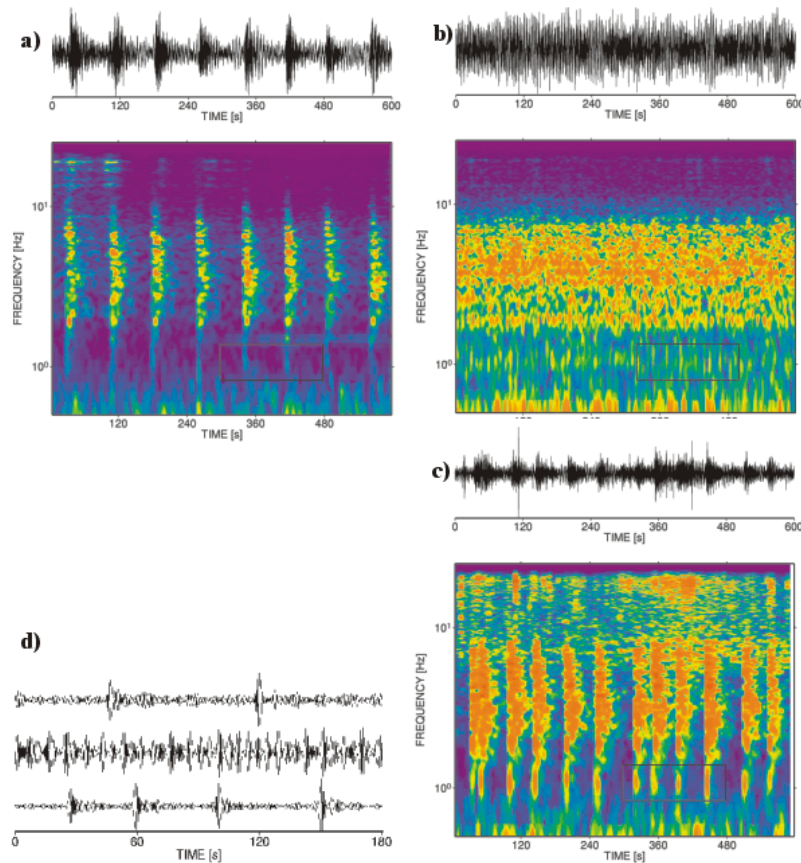
During the last decade, many observations were made of the occurrence and characteristics of volcanic tremor at volcanoes with high-viscosity lava. At Semeru volcano (Indonesia) the spectra of volcanic tremor contained up to 12 overtones. This supports the assumption of a resonating medium with a high quality factor ( $Q$ ) as well as a precisely working feedback mechanism (Hellweg et al., 1994; Schlindwein et al., 1995) (Fig. 13.16). Similar observations were also made at Lascar volcano (Chile), where up to 30 overtones could be identified in the seismic signals (Hellweg, 1999).



**Fig. 13.16** Harmonic tremor signal recorded at Mt. Semeru, Indonesia. Up to six overtones can be recognized starting with a fundamental mode located at roughly 0.8 Hz.

### Source models and Interpretation

Schlindwein et al. (1995) proposed a feedback mechanism similar to that of sound generation in a recorder, and also discussed a repeating source with precise repetition time as a possible mechanism. This model was refined by Johnson and Lees (2000) and Neuberg et al. (2000). In the feedback mechanism case, the resonating body must consist of a pure gas phase as the lava at Mt. Semeru is too viscous for resonance at the observed frequencies. The second mechanism requires a very precise timing mechanism for producing the highly stable overtones. Observations at Montserrat volcano (Neuberg et al., 2000) and Mt. Merapi volcano (Indonesia) support the hypothesis of a repeating source (Fig. 13.17). During several cycles of increased volcano-seismic activity we recognized the transition from closely timed MP/Hybrid events into the continuous signal of volcanic tremor and vice versa. As the source mechanisms of both types of signals are still unknown, the driving force behind these mechanisms is also not known. Also the type of feedback mechanism, which needs to be involved in this model could not yet be identified.

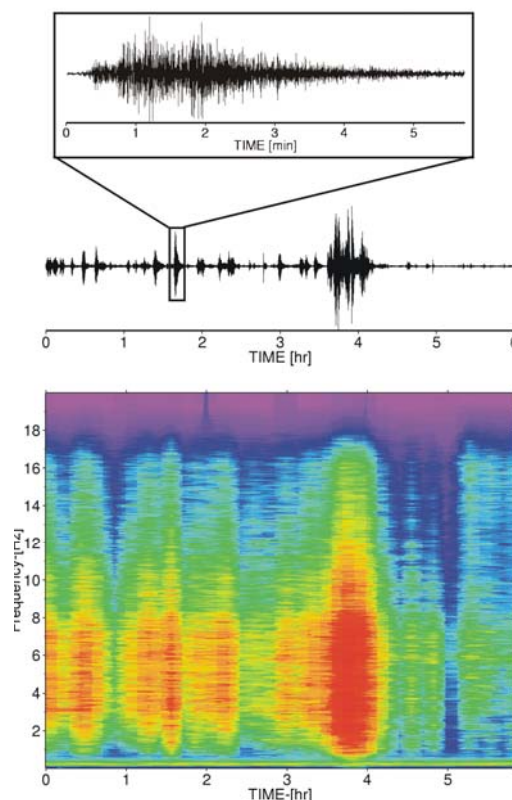


**Fig. 13.17** Sequence of repeating seismic signals at Mt. Merapi volcano in 1996. a) very regularly timed MP-events before they merge together to form volcanic tremor (see b). c) after some hours the tremor is replaced by a sequence of discrete events with slightly higher amplitudes than before. Note: in contrast to the classification given in Fig. 13.5, the frequency content of these signals is lower (0.7 - 10 Hz) and might not resemble “pure” MP-events. In d) the marked time-frequency region of plots a)-c) are plotted in time domain. A band-pass between 0.8 - 1.3 Hz was applied before zooming. The individual wavegroups seen in the filtered continuous signal also supports the idea of the merged events causing the volcanic tremor.

Volcanic tremor is, as previously noted, always a sign of high activity. However, since the exact mechanisms are still unknown, the importance of and timing between the first appearance of tremor and possible eruptive activity is still a matter of discussion (McNutt, 2000a). The review papers by Konstantinou and Schlindwein (2002) and Kawakatsu and Yamamoto (2007) may give further information about the types and proposed source mechanisms of volcanic tremor.

### 13.2.2.3 Surface processes

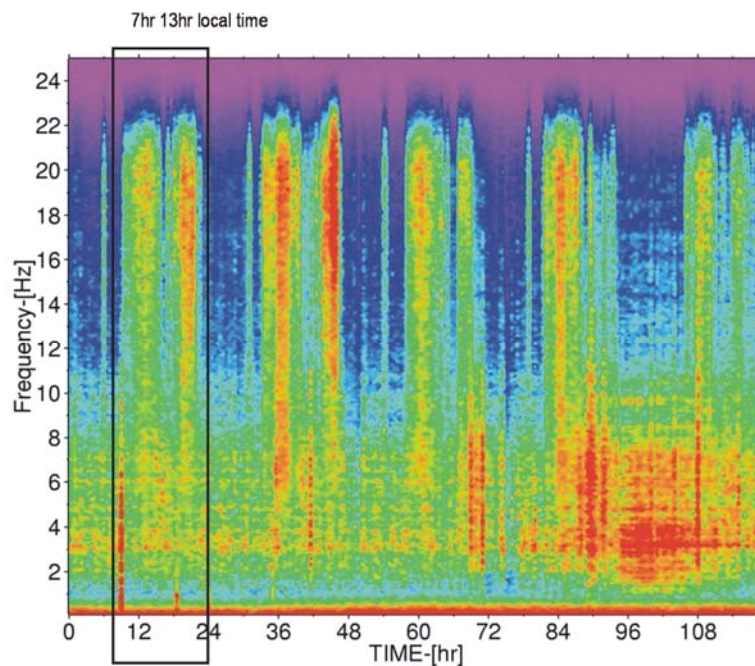
Substantial release of seismic energy at active volcanoes is related to surface processes acting directly on the volcanoes edifice. For example, pyroclastic flows, lahars (volcanic debris flows) and rockfalls from unstable domes or crater walls can generate seismic signals with amplitudes exceeding several times those of the typical volcano-seismic signals. The most important signals for monitoring purposes are those associated with pyroclastic flows and lahars. The monitoring of lahars, which includes also acoustic and visual monitoring, is especially important when monitoring a volcano which is capped by a glacier or which is located in a tropical area. Melting the snow during an eruption or heavy rainfall during the rainy season will occasionally mobilize a huge amount of volcanic debris. The signals of all this activity are mostly high-frequency ( $>5$  Hz), show spindle (cigar) shaped seismogram envelopes that can last several minutes (Fig. 13.18).



**Fig. 13.18** Sequence of medium to larger pyroclastic flows recorded at Mt. Merapi volcano during the 1998 dome collapse. Note the 6-hour time scale and that individual events last many minutes, longer than the seismograms of typical earthquakes. Just before 4 hours the largest pyroclastic flow in the whole eruption sequence takes place and lasts for about 30 minutes.

### 13.2.3 Special note on noise

Most of the extensively monitored volcanoes lie in densely populated areas with strong human activity (that is why they are monitored). Hence, care must be taken when interpreting the signals usually classified as volcanic tremor. In some cases, human activity excites signals occupying the narrow spectral band between 1-4 Hz (big machines etc.). Also a distinct 24 h rhythm is very likely caused by increasing human activity during daylight time and should therefore be analyzed with special care (Fig. 13.19). Even when using three-component seismometers it is not easy to discriminate for sure between volcano-seismic and man-made noise. The topography at active volcanoes is very often radially shaped and the propagation paths to the seismic stations are shared by ambient seismic noise and volcanic signals.



**Fig. 13.19** Spectrogram of background noise recorded at a seismic station at Mt. Merapi. As the station is located in farming area, the human daylight activity can be clearly recognized by its distinct 24 hour periodicity. Furthermore, it is possible to see that there are two main working periods during daytime (marked by a box). Large spectral amplitudes are visible around 7 hrs local time and a second peak is located around 15 hrs after a time of quiescence during noon.

In conclusion, we note that most of the above classifications and proposed source mechanisms are deduced from simple observations of spectral content and overall shape of the associated seismograms rather than by physically verified constraints. Care must be taken when interpreting the occurrence of one of these signals during increasing volcanic activity. There are many examples of increasing numbers of VT events and increasing volcanic tremor amplitude without any surface activity at volcanoes. Thus, to be truly effective and diagnostic, seismic monitoring should be complemented, wherever possible, by other instrumental monitoring techniques (e.g., geodetic, geochemical) and regularly made visual observations of the volcanoes being monitored remotely (see Section 13.5).



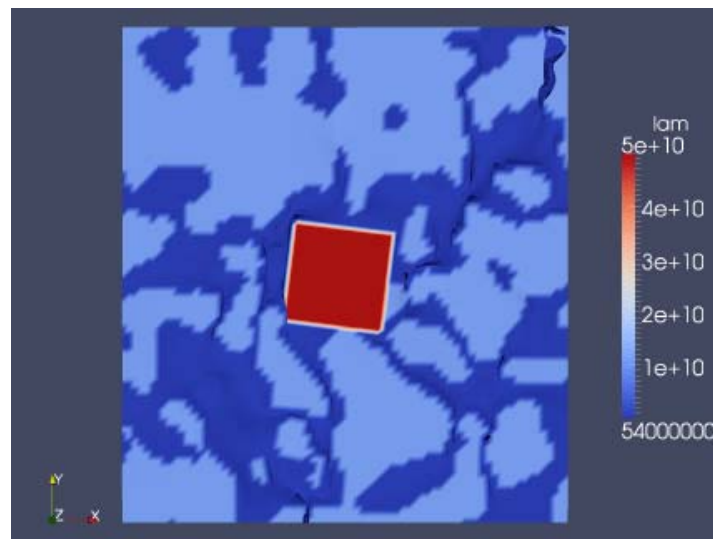
## 13.3 Design of a monitoring network

One of the most important decisions to be made, when establishing a seismic monitoring network, is the design of the station distribution. In most cases, volcanoes are monitored with at least four to six seismic stations, which are distributed around the volcanic center. Newer deployments try to setup arrays of sensors or, even better, a network of different arrays (also named “Array of Arrays”). However, some of the design criteria deviate from the usual earthquake monitoring networks and are discussed in the following sections.

### 13.3.1 Station site selection

Considering a location as a possible site for a seismic station is always a compromise between noise considerations and accessibility. Of course, it would be best to place the seismic station far away from any human activity (see Fig. 13.19), away from big trees or sharp cliffs and ridges. However, the accessibility is very important, especially at the beginning of a surveillance campaign at a volcano. Also, the rough and harsh environment typical of many volcanoes usually requires frequent station visits for maintenance. Valleys, which generally are accessible places for seismic stations rather than ridges or cliffs, are often flooded during winter or the rainy season (not to mention the higher exposure to possible pyroclastic flows).

A second important decision must be made when choosing on “what” the station should be placed. Usually, seismologists prefer hard rock to unconsolidated sediments. At many active volcanoes, hard-rock sites are rare and, even if they exist, they are not necessarily good choices. Hard-rock sites often are small lava tongues or big blocks of lava buried in ash or soil, causing waveguide effects or even block rotation to an unknown degree. This is especially important when installing broadband seismic stations, which are very sensitive to tilt (see Chapter 5 and Fig. 13.20).



**Fig. 13.20** Simulated effect of strain-tilt coupling. The uniform input strain acts on the y-direction. As a consequence of medium inhomogeneity an assumed rigid plate (which is a placeholder for the seismometer monument) is rotated and tilted considerably. The inhomogeneity is caused by the modelled random (gaussian) medium which is represented by alternating values of Lamé's parameter  $\lambda$ . (courtesy M. v. Driel).

A network-wide homogeneous installation with good temperature isolation is preferable to apparent “hard-rock” installations (see Section 13.6 for a more detailed description). Sites near singular obstacles should be avoided such as high trees, cliffs, big towers etc., as they are likely sources of wind-generated noise. While wind noise is usually high in a frequency range  $> 5$  Hz, wind pressure is a very strong source of tilt-noise in the low frequency part ( $< 0.1$  Hz). Hence, special care must be taken when installing a broadband seismic instrument.

### 13.3.2 Station distribution

Good station coverage is crucial for nearly all monitoring efforts as well as for successful scientific research. A good choice is to install a network at two scales - one large scale network extending into non volcanic regions ( $\Delta < 20$  km) and one network with stations concentrated on the flanks and on the top of the volcano ( $\Delta \sim 0 - 2$  km). The large-scale seismic networks are very useful to distinguish between volcano-seismic signals and regional or local earthquake activity. Also the larger dimension improves the localization accuracy for deep-seated sources of magmatic activity. On the other hand, most of the seismic signals at active volcanoes are very shallow, usually small in amplitude and additionally influenced by strong scattering and high attenuation path effects. For detailed studies of the volcano-induced seismic signals, most of the stations need to be placed close to the activity center(s). One or two stations should be placed as near as possible (without danger to researchers and instruments) to the active volcanic region. Other stations should be placed such as to ensure a good overall azimuthal coverage. If possible, the station spacing should be comparable to the source depth to insure good depth control. It is good to have all parts of the focal sphere surrounding a source to be sampled by seismic stations. Best results are expected when the source is located within the station network, both lateral and vertical.

If a broadband seismometer is available, best results are achieved if it is installed as close as possible to the active area, provided that the safety of operating personnel is assured. Most of the recorded ULF and LF signals at volcanoes are detectable only in the near-field distance range (see Fig. 13.10) as well as the effect of scattering is lower as closer the station is located to the active source region (Section 13.4.5). For the best characterization of shallow low and ultra low frequency event source mechanisms, at least a few seismic stations should be placed as close to the crater or source region as possible. If an on-line radio link is desired, the station site must additionally be chosen such as to guarantee an undisturbed direct line-of-sight to repeaters or receivers (see Section 13.6 and Chapter 7, Section 7.3).

### 13.3.3 Seismic arrays in volcano monitoring

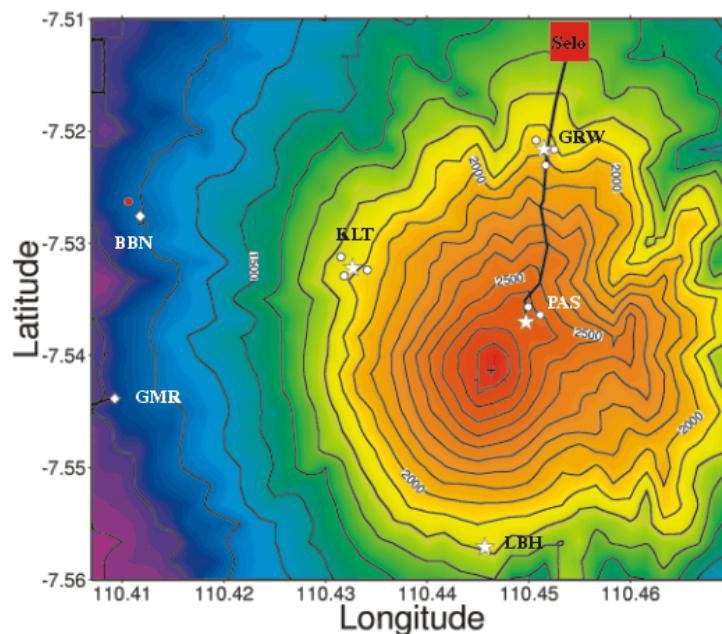
Modern approaches in volcano seismology are based on deploying seismic antennas (arrays) at active volcanic areas. Stations in an array should be spaced close enough to sample a wavefield several times in a wavelength, often requiring a spacing of about 100 m. The main advantage of such antennas and the application of array techniques is the improvement in evaluating the radiated wave-field properties, velocity structure and the source location (see Chapter 9). A comprehensive review paper dealing with standard seismic array techniques at volcanoes has been published by Chouet (1996b).

Most of the problems in operating a seismic array at an active volcano are of a technical

nature. The requirements on array site conditions are demanding, the cost of array components are rather high, and the installation and maintenance of an array during different activity stages and weather conditions require significant economic and human resources. Such requirements generally preclude the long-term use of arrays in volcano monitoring. Therefore, most of the work done so far in using array techniques at active volcanoes were short-term deployments of occasionally large arrays. Despite the mostly short duration of deployment, however, much information was gathered during these experiments. The results range from a more comprehensive description of the wave-field properties (Saccorotti et al., 1998; Chouet et al., 1997) to tracking the source volume of volcanic tremor signals (Almendros et al., 1997; Furumoto et al., 1990).

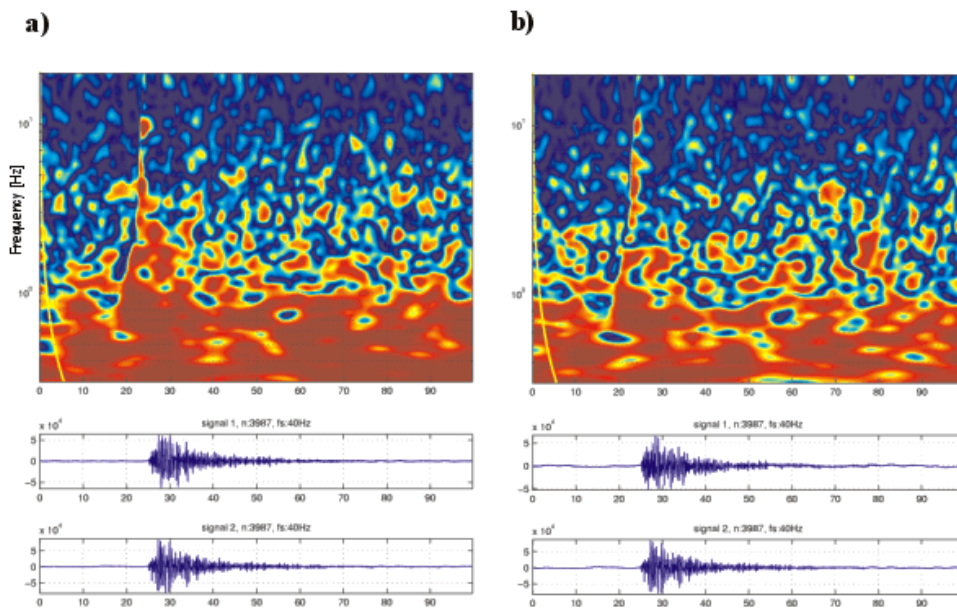
### 13.3.4 Array of seismic arrays

In attempting to achieve both monitoring and research objectives, a good compromise is to establish a network of small-aperture seismic arrays. The advantage compared to single (dense) array applications is the better spatial evaluation of the wave-field properties as well as the better azimuthal coverage when focusing on the location of the different seismic signals. In any event one has to compromise between aperture, number of instruments, spatial sampling and station accessibility. In 1997, a network of small-aperture arrays was established at the Merapi volcano, Indonesia (see Fig. 13.21). This network consists of three different array sites distributed around the volcano. The main objective of this array configuration is to attempt the automatic classification of the volcano-seismic events on the basis of the wave-field properties and an automatic hypocenter determination of the classified volcano-seismic events (Wassermann and Ohrnberger, 2001).



**Fig. 13.21** Example of a combined seismic array/network approach at Mt. Merapi volcano. The stars show the location of broadband seismometers, whereas the circles mark the position of three-component short-period seismometers, in total forming three small-aperture arrays. The diamond symbols show the location of seismic acoustic stations (short-period sensors with a microphone array).

Before installing a network of arrays, a detailed plan of features to be investigated and criteria to be met should be made, e.g., required spatial coverage and resolution, accuracy of hypocenter determination, shallow and/or deep seismicity, broadband signals etc. A good choice is a network with at least four different array sites. Each array should consist of one three-component broadband seismometer as central station surrounded by three to six short-period, vertical-component seismometers deployed in a configuration which fits best the number of seismic stations (see Chapter 9). The most suitable distance between related seismometers must be carefully evaluated during the initial stage of the setup. Decisions must be made between the peak values in the spectral domain and the desired coherence band of the signals recorded. Ideally, the stations should be roughly 100 to 200 m apart from each other (Fig. 13.21). Reducing the inter-station distances with the same number of seismometers will cause an undesired loss of resolution in slowness due to the smaller aperture and also increase the noise coherence (see Fig. 13.22).



**Fig. 13.22** Time-frequency coherence plot of a) the station combinations GRW0- GRW1 and b) GRW1-GRW2 (see Fig. 13.21). The seismometers in a) are deployed 170 m from each other, whereas in b) GRW1 and GRW2 are separated by roughly 300 m. Note: the signal coherence is computed in a sliding window with the time axis centered at the middle of the sliding window. High coherence above 2 Hz is only visible in the very beginning of a seismic event, indicating an array wide coherent phase arrival. It is also obvious that the overall coherence is somewhat lower in b) than in a) indicating the reduced signal coherence at more separated stations. On the other hand the noise coherence is also reduced in b) which improves the signal to noise ratio of the semblance estimation significantly.

## 13.4 Analysis and interpretation

In this chapter, we will briefly review the basic techniques of analyzing volcano-seismic signals. Most of the described concepts are based simply on visual pattern recognition abilities of the responsible interpreter. The last two sections will discuss more recent and objective approaches that attempt to automate these tasks.



### 13.4.1 One-component single station

Most of the observations made in the 60's and 70's were obtained by using only a few instruments located at most active volcanoes. Since then, nearly all well-monitored volcanoes are equipped with at least four to six instruments and, for a number of volcanoes, dozens of instruments. However, the basics of the classification scheme discussed in Section 13.2 is deduced by the single station approach and even today the statement “better one than nothing” holds regarding the number of deployed instruments. This is especially true when initiating short-term projects or monitoring very remote volcanoes.

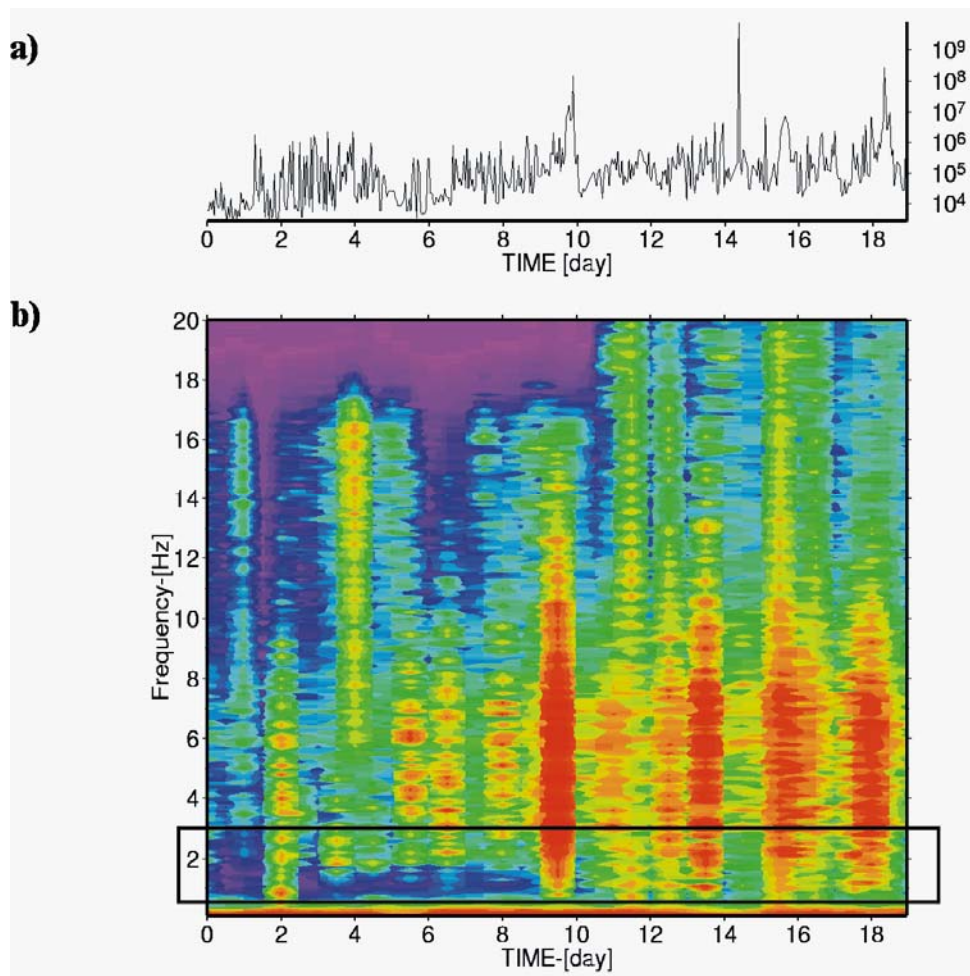
#### 13.4.1.1 Spectral analysis

With the advent of inexpensive, portable and efficient computers, spectral analysis became an increasingly important tool for monitoring the activity of an active volcano. As mentioned already in Section 13.2, most of the classification is based on the time-frequency characteristics of seismic signals. Volcanic tremor episodes were distinguished by their spectral shape and appearance. There are many different techniques for computing the spectral seismic amplitude such as *Seismic Spectral Amplitude Measurement* (SSAM; Rogers and Stephens, 1991), short-term Fourier transform or power spectral density estimates, which provide the observer with signal information in the spectral domain (e.g., Qian and Chen, 1996). An important feature of volcano-seismic signals is their narrow-band spectra. In particular, volcanic tremor sometimes shows just one dominant spectral band with a bandwidth as small as 0.2 Hz. This is the reason why it is often called “harmonic tremor”. Monitoring the changes of spectral properties is a useful tool not only for signal discrimination but also for characterizing the state of volcanic activity. An example is given in Fig. 13.23. In Fig. 13.23a the total power in the frequency range between 0.6 and 3.0 Hz is plotted as a function of time. This frequency range was chosen because of its importance in discriminating between rockfall and pyroclastic flow signals (see Section 13.2.2.3). Three pronounced peaks are obvious with amplitudes well above the average value. The peaks at day 9 and day 18 are associated also with a significant increase of the power density between 2 to 10 Hz (Fig. 13.23b). On the other hand, the sharp peak in day 14 in Fig. 13.23a seems to be of a different nature and might be caused by a regional or teleseismic earthquake. The remaining times with high power density amplitudes in b) might be due to small pyroclastic flows or rockfall.

At Mt. Etna (Italy), Cosentino et al. (1989) reported a significant frequency shift in the volcanic tremor spectra prior to a flank fissure eruption. The authors detected a significant shift to lower frequency values of the dominant spectral peaks of volcanic tremor just several hours before the opening of flank fissures.

Because the efficiency of today's computers is rapidly increasing, a good choice would be to calculate complete spectrograms (or periodograms) first and decimate the amount of data only in a later step (e.g., to SSAM). This would allow the extraction of any hidden information in a later “off-line” step of the analysis without any redundant work load. Crucial in this context is a good knowledge of the possible features of different signals and their relationship to the state of volcanic activity at a specific site. It must be emphasized that stations of monitoring networks at volcanoes should be maintained for years (even decades) without any changes in the system (gain, position etc.). When upgrading an old station to new “up-to-date” technology, a sufficient overlap of both systems should be guaranteed. This precaution cannot

be overemphasized.



**Fig. 13.23** a) shows the the total power (per 60 minutes) calculated in the frequency band between 0.6 - 3.0 Hz from 01 - 19th July 1998 at Mt. Merapi, displayed on a logarithmic scale. Two of the visible peaks (i.e., day 9 and day 18) are associated with pyroclastic flows, while the sharp peak visible at day 14 is caused by a regional earthquake. b) the power spectral density vs. time in the same time range, where the box shows the frequencies used for total power plotted in a).

#### 13.4.1.2 Envelope, RSAM and cumulative amplitude measurements

An important source of information which can be deduced by small networks is the overall appearance of the signal shape in the time domain. This is important both for event classification (e.g., volcanic tremor, rockfall etc.) as well as for monitoring changes in the seismic activity of a volcano. A very efficient tool for visualizing increasing seismic activity is the *Real-Time Seismic Amplitude Measurement* (RSAM) technique proposed by Endo and Murray (1991). In its original form, RSAM was designed for analog telemetry and consisted of an A/D converter, averaging of the seismic signal in 1 min or 10 min intervals and storing of the reduced data on the computer:

$$RSAM(iT) = \frac{1}{T} \sum_{t=iT-\frac{T}{2}}^{iT+\frac{T}{2}} |s(t)|$$

Here  $T$  is the averaging interval (originally 1 or 10 min) and  $s(t)$  the sampled seismic trace. Various examples for successful applications of this technique are given by McNutt (2000b).

Some applications try to normalize the records from several seismic sensors located at different distances from the volcanic center by correcting the measured seismic amplitude for the assumed source distance (McNutt, 2000b):

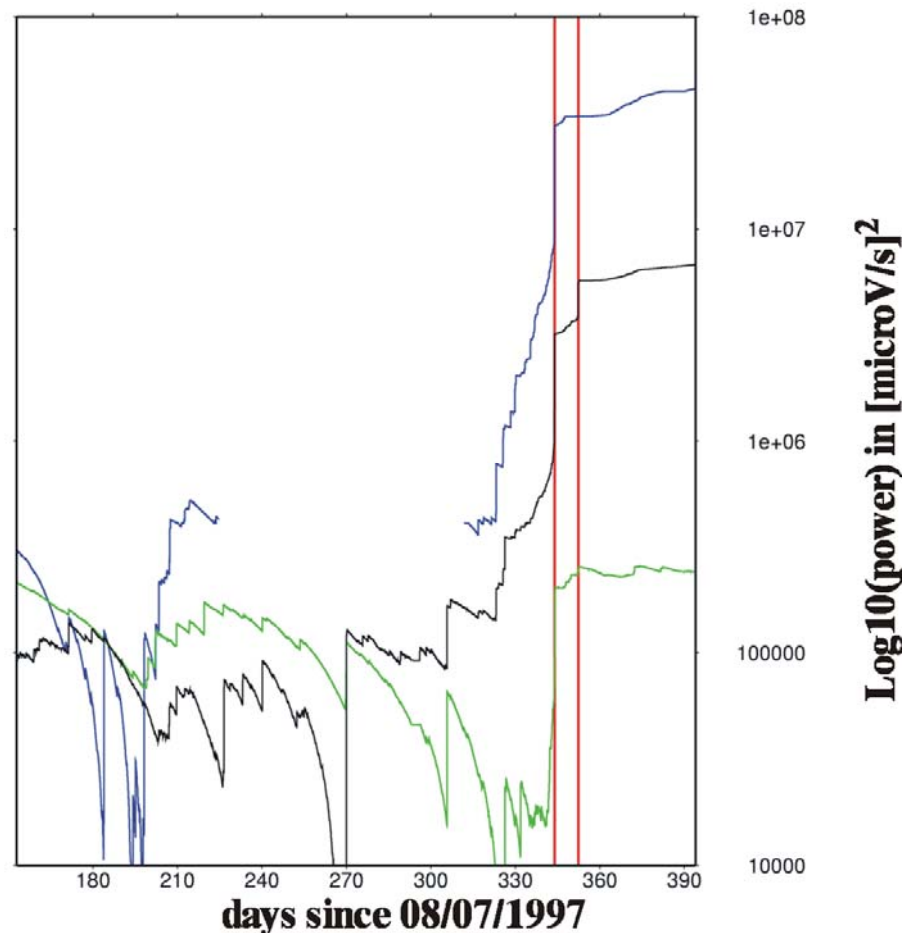
$$D_R^b = \frac{Ar}{2\sqrt{2}G}, \text{ and } D_R^s = \frac{A\sqrt{\lambda r}}{2\sqrt{2}G}$$

where  $D^b$  and  $D^s$  are the reduced amplitude for body and surface waves, respectively.  $A$  is the peak to peak amplitude in centimeters,  $r$  the distance to the source,  $\lambda$  the seismic wavelength in cm and  $G$  the gain factor (magnification) of the seismic sensor. The main difference between these two equations are the different correction terms for the geometrical spreading and the effective wavelength. The reduced amplitude measurements should be considered as a pure observation parameter without any physical meaning. It should definitely not be used for the physical interpretation of an ongoing eruption. The reduction of the seismic amplitude assumes specific modes of wave propagation, i.e., body wave and surface waves, respectively. As there is no reliable estimate of the wave-field properties, it is possible with just one or a few seismometers that the assumption of the degree of geometrical spreading is highly speculative. Also the effect of site amplification and the strong scattering observed frequently at volcanoes (Wegler and Lühr, 2001), which depend in general on the source location, structure and topography of the volcano, may alter the amplitude-distance relationship significantly. They are neglected in this approach.

Another way of displaying changes in the radiated seismic wave-field is based on the computation of the de-trended cumulative radiated power of the seismograms at a single station (see Fig. 13.20):

$$P_{cum}(f_{1-2}, t) = \sum_t \left( \sum_{f=f_1}^{f_2} P_t(f) - trend \right)$$

with  $P_t(t)$  being the power spectral density during the time interval  $t$  and  $f_1, f_2$  the lower and upper frequency for computing the cumulative power. *trend* is the slope of the cumulative power, calculated during a quite time period, i.e., baseline activity of the volcano (Fig. 13.24).

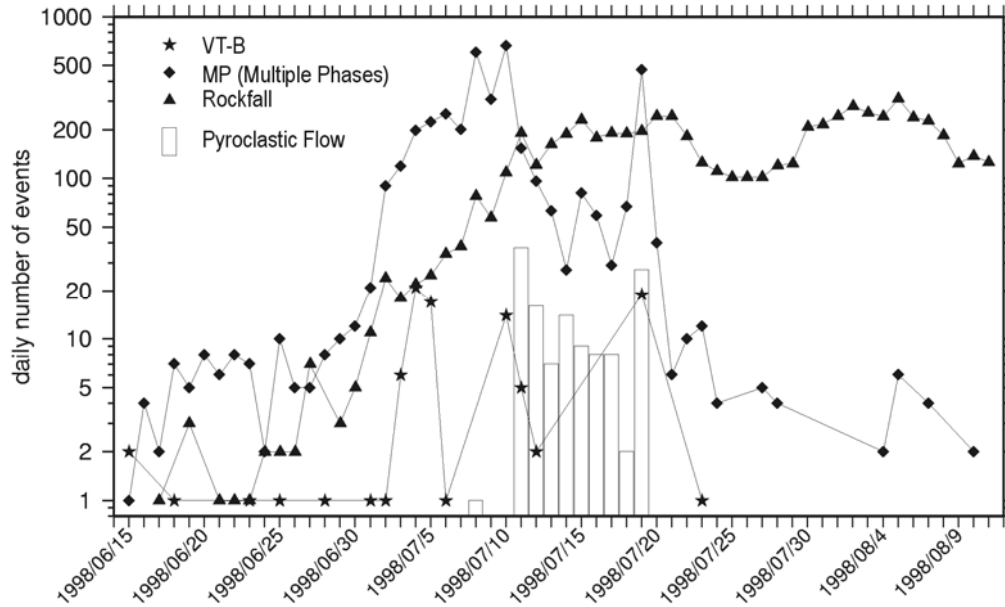


**Fig. 13.24** De-trended cumulative power of the vertical components of all broadband seismometers at Mt. Merapi during 1998 activity. The red lines mark the occurrence of two pyroclastic flows. A steep increase of the total cumulative power 10 days before the onset of the first pyroclastic flow is visible, following a period with very low seismicity. Also the second eruption is preceded by an increase of cumulative power at two stations, while one station (blue) was out of operation. The background trend was estimated during a low activity phase in 1997.

To avoid a fast saturation of the cumulative power values, a good way is to estimate the slope of the cumulative power when the activity of the volcano is on its baseline. This estimated trend can be removed for each time step resulting in a de-trended cumulative power plot, which shows strong deviation from the “normal” background seismicity. Furthermore, cumulative power can be used to analyze certain frequency bands (see Fig. 13.23a), unlike the power change in time, which resembles the method of RSAM. In Fig. 13.24, the de-trended cumulative power of three broadband stations at Mt. Merapi is shown. Note the steep increase of seismic power roughly ten days prior to the first eruption. A further increase in cumulative power is obvious for two stations preceding the second large pyroclastic flow. The third station was out of operation caused by ash fall on the solar panels.

A common way to display information on the current status of a volcano is to count the different seismic event types in a hourly or daily manner (Fig. 13.25). While the physical interpretation of the type of event is sometimes impossible when using only one station, such event/time plot is an excellent tool for displaying all information (objective and subjective)

within one single plot. There are many papers, which rely strongly on this kind of activity measurements. Most of the observations are summarized by McNutt (1996, 2000b). Also in this case, we must emphasize that, without a complementary detailed seismological study, this is just a visualization of observed patterns with, strictly speaking, unknown physical meaning.



**Fig. 13.25** Event-type per day plot during the high activity of Mt. Merapi during July 1998. Note the increase of the three event classes before the onset of the first pyroclastic flow. Also note the similarity of the VT-B type event curve to the occurrence of pyroclastic flows (courtesy of VSI- BPPTK, Yogyakarta).

Whatsoever, if knowledge about the hypocenters (see Section 13.4.3.1) and even source mechanism is available these event-time plots are very valuable in order to evaluate the activity state of a volcano. At Soufriere Hills volcano (Montserrat island) it was possible to distinguish different activity phases with the help of these seismicity plots (Miller et al., 1998). Just before the surface activity starts to develop a swarm of VT-A earthquakes appeared. During cycles of inflation in the upper part of the growing dome a large number of Hybrid and LP events were detected. Finally, a large number of surface events, mainly rockfall signals, were recorded as the dome became more and more unstable. While this pattern of seismic signals is very important during a high activity phase of a volcano, it must be emphasized that every volcano and every eruption has its own unique pattern.

#### 13.4.1.3 Failure Forecast Method (FFM)

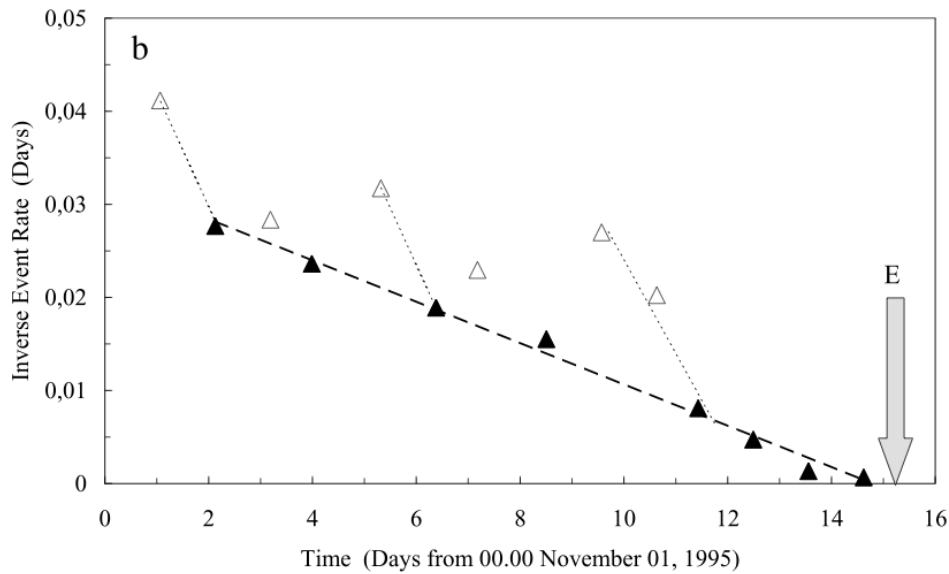
While the FFM originates from material sciences and relates the rate of precursory signs of failure to the likeliness of failure, Voight (1989), Cornelius and Voight (1995) and Kilburn (2003) introduced an early warning method, which is based on the rate of precursory signs (e.g., number of a certain type of seismic signal, RSAM, cumulative tilt etc.) and correlates their accelerations to the onset time of an eruption. The governing equation is given by:

$$\partial^2 \Omega / \partial t^2 = A(\partial \Omega / \partial t)^\alpha$$

where  $\partial^2\Omega/\partial t^2$  and  $\partial\Omega/\partial t$  are the acceleration and rate of the monitored parameter.  $A$  and  $\alpha$  are empirically determined parameters. Cornelius and Voight (1995) pointed out that  $\alpha$  evolve from 1 to 2 before an eruption. Kilburn (2003) pointed out that the peak of event rates are better suited to forecast the timing of a failure (i.e. eruption) and that  $\alpha = 2$  should be used. With this assumption the equation can be simplified to:

$$1/(\partial\Omega/\partial t) = A(t_f - t)$$

where  $t_f$  is the time of failure. Fig. 13.26 gives an example of a successful application of FFM.



**Fig.13.26** Change in recorded seismicity before the 1995 eruption of Soufriere Hills on Montserrat. Arrow E marks onset of magmatic activity. On an inverse-rate diagram, the filled triangles, which correspond to inverse-rate minima, lie along a linear trend (large dashed line). Such a trend possibly could be recognized by Day 8, which results in a maximum 6-day warning of eruption. Note that peak values are more readily identified from the inverse-rate minima than the actual event-rate maxima, and that steeper linear trends occur among individual sequences (small dashed lines) (from: Kilburn, 2003).

### 13.4.2 Three-component single station

Most modern seismometers are three-component sensors, which record the vector of ground motion produced by seismic waves. Observation of the particle motion will not help to precisely determine source locations and their variations without a detailed knowledge of the wave-field properties (e.g., Rayleigh waves, Love-waves, P- or SH and SV-waves). In some cases it might be possible to “steer” or point towards the active source region by estimating the particle motion of the very first wavelets of LF or ULF events. As this is possible mainly in the near field there might be an unknown degree of distortion. Changing patterns of particle motion may help estimate the activity changes of a volcanic system in a qualitative way, however.

### 13.4.2.1 Polarization

Seidl and Hellweg (1991) showed results from analyzing the 3D-trajectories of a single seismic broadband station at Mt. Etna volcano (Italy) using very narrow bandpass filters. They argued that the occasional strong variations in the azimuth and incidence angles of the trajectories may reflect sudden changes of the active source location. Recent experiments using array techniques showed, however, that the wave field radiated from a volcanic source is a combination of complex source mechanisms and strong path influences (e.g., Chouet et al., 1997). Hence, the wave-field consists of a mixture of many wave types and care must be taken when only polarization information is available. On the other hand, carefully extracted information and the associated changes of polarization pattern during different cycles of volcanic activity may help to identify changes in the state of the volcanic system (see Fig. 13.27 below).

The use of a broadband seismic station located close to an active vent, i.e., in the near-field, improves the quality of source estimations based on simple polarization analysis. This is because of the small influence of the propagation path in the near-field. Unfortunately, complicated source mechanisms, i.e., when the usual assumption of a point source is no longer valid, will complicate the interpretation of the observed polarization pattern to an unknown degree (Neuberg and Pointer, 2000). Also, the nearly unknown influence of the topography of the volcano on signals with a wavelength comparable to the topographic obstacle will make interpretation difficult. Near-field measurements at Stromboli volcano (Italy) showed, that in some cases, a fairly good estimation of the source region could be made using just a single three-component broadband station (Kirchdörfer, 1999; Hidayat et al., 2000) under the assumption of a simple source mechanism.

### 13.4.2.2 Polarization filters

When evaluating the polarization properties of volcano-seismic signals as part of a monitoring system, an automatic estimation of parameters is needed. Best results will be obtained when focusing on the basic parameters, i.e., the azimuth, incidence angle and a measure of the rectilinearity of the signals. Various approaches will extract this information from a continuous data stream. Most of them are based on a least-square fit of the 3D-trajectory of the seismic vector to a 3D-ellipsoid. Typical algorithms consist of solving the eigenequation and simultaneously searching for the orientation of the eigenvector corresponding to the largest eigenvalue (e.g., Flinn, 1965, Montalbetti and Kanasevich, 1970):

$$\begin{bmatrix} x_i \\ y_i \\ z_i \end{bmatrix} \left| C - \lambda_i I \right| = 0$$

where  $x_i$ ,  $y_i$ ,  $z_i$  represent the components of the  $i$ -th eigenvector,  $I$  is the identity matrix and  $\lambda_i$  is the eigenvalue according to the  $i$ -th eigenvector.  $C$  represents the covariance matrix of the 3D signal recorded:

$$C_{ij} = \sum (u_i - \bar{u}_i)(u_j - \bar{u}_j)$$

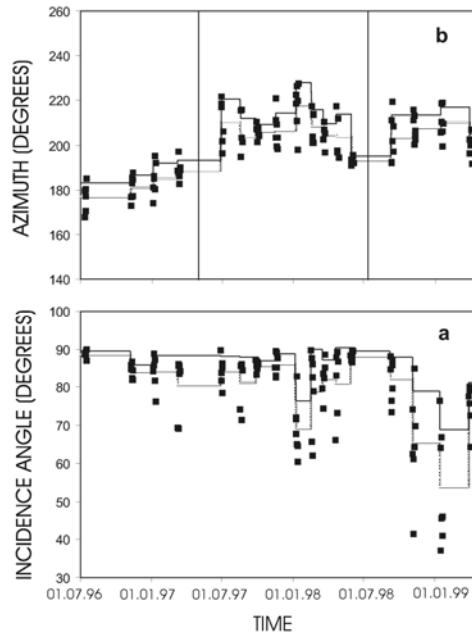
where  $u_i, u_j$  are the  $i$ -th and  $j$ -th component of the seismic sensor and  $\bar{u}_i, \bar{u}_j$  represents the mean values of the data traces within the analyzed time window. A possible way to display the polarization properties vs. time is to plot the orientation of the eigenvector associated with the largest eigenvalue (corresponding to the major axis of the ellipsoid) in the coordinate system of the sensor, i.e., its azimuth  $\Phi_{az}$  and incidence angle  $\Theta_{inc}$ :

$$\Phi_{az}(t) = \text{atan}\left(\frac{y_1(t)}{x_1(t)}\right), \text{ and } \Theta_{inc}(t) = \text{atan}\left(\frac{z_1(t)}{\sqrt{x_1^2(t) + y_1^2(t)}}\right)$$

with  $x_1, y_1, z_1$  representing the eigenvector components of the largest eigenvalue  $\lambda_1(>\lambda_2>\lambda_3)$ . Note: without any further assumption about the analyzed wave-type, i.e., P-, SH- or SV-wave etc., the computed azimuth has an ambiguity of 180 degrees, whereas the incidence angle varies between 0 - 90 degrees. Typically, a measure of the rectilinearity of the signal's polarization (i.e., the relative elongation of the ellipsoid in one direction) is computed (e.g., Vidale, 1986):

$$L(t) = 1 - \left(\frac{\lambda_2(t) + \lambda_3(t)}{\lambda_1(t)}\right)$$

$L(t)$  is only larger than 0 if  $\lambda_1$  is bigger than the combination of the other two. Fig. 13.27 gives an example of the variation of the parameters  $\Phi_{az}$  and  $\Theta_{inc}$  over a long time period at Stromboli volcano.



**Fig. 13.27** Long-term variations of incidence angle (a) and azimuth (b) in the 115 time windows selected from July 1996 to April 1999 at Stromboli volcano. In both panels, solid and dotted line depict mean value and standard deviation, respectively, computed over 5 to 7 consecutive days in the 17 time windows selected from July 1996 to April 1999. The polarization parameters were estimated using the technique of Montalbetti and Kanasevich (1970). The variation of this waveform information seems to match changes in the activity states of the volcano (courtesy of S. Falsaperla, Istituto Nazionale di Geofisica e Vulcanologia).



Because we have no knowledge of the wave type represented by the computed polarization parameters, they must be seen as varying activity parameters rather than interpreting them as part of a technique for hypocenter determination.

### **13.4.3 Network**

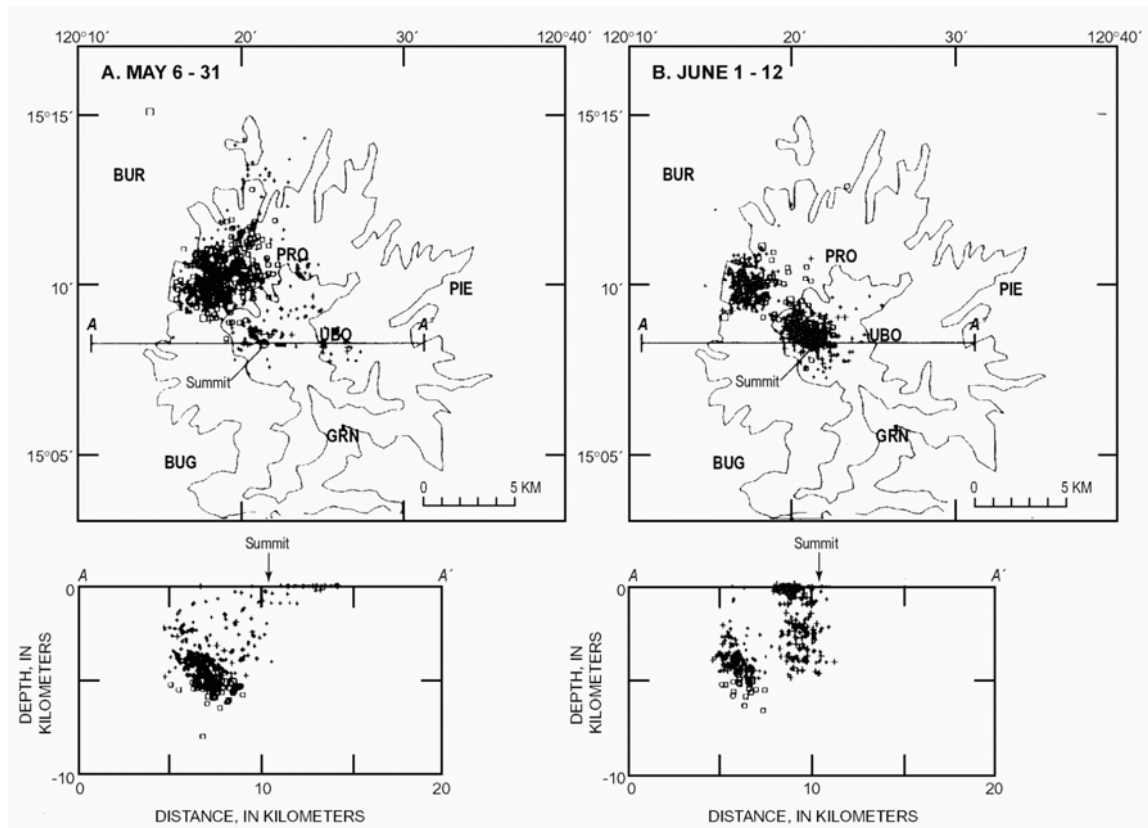
#### **13.4.3.1 Hypocenter determination by travel-time differences**

Modern seismic monitoring networks at active volcanoes usually consist of at least four to six seismic sensors distributed in various azimuths and distances from the volcanic center. While continuous signals such as volcanic tremor or transients like LF-events often lack any clear phase arrival, some signals (VT, explosion quake) with clear onsets can be located using standard seismological techniques.

Usually, events with clear P- and/or S-wave onsets are selected visually and the first breaks are picked interactively. The inversion for the source location is frequently done using algorithms such as HYPO71 (Lee and Lahr, 1975) or HYPOELLIPSE (Lahr, 1989) or HYP2000 (Klein, 2002). Note, however, that most of the standard hypocenter determination programs are based on the assumption of a horizontally layered half-space and/or models with linear gradients with no topography. Also new approaches exist, which are not restricted to 1D or 2D velocity models and which try to locate the sources in a non-linear, probability based manner (e.g., Lomax et al., 2000). However, in most cases no good velocity models for the monitored volcanoes exist and the computed source coordinates, especially when focusing on shallow events, must be seen just as an approximation to the true hypocenter. Relative earthquake locations of multiplets with similar waveforms can greatly improve the resolution of volcanic structures (Rubin et al, 1998; Waldhauser and Ellsworth, 2000; Ratdomopurbo and Poupinet, 1995). De Barros et al (2009) showed that when the network is dense enough such that events correlate at different station very good and reliable locations are possible using a grid search method.

There are many papers on the topic of imaging the hypocenter distribution during or before a volcanic eruption (e.g., Newhall and Punongbayan, 1996; Power et al., 1994; Chouet et al., 1994). Very useful information about the geometry of the plumbing system as well as the physical properties of the host rocks can be deduced by analyzing the time-space pattern of frequently occurring swarms of deeper earthquakes (Power et al., 1994).

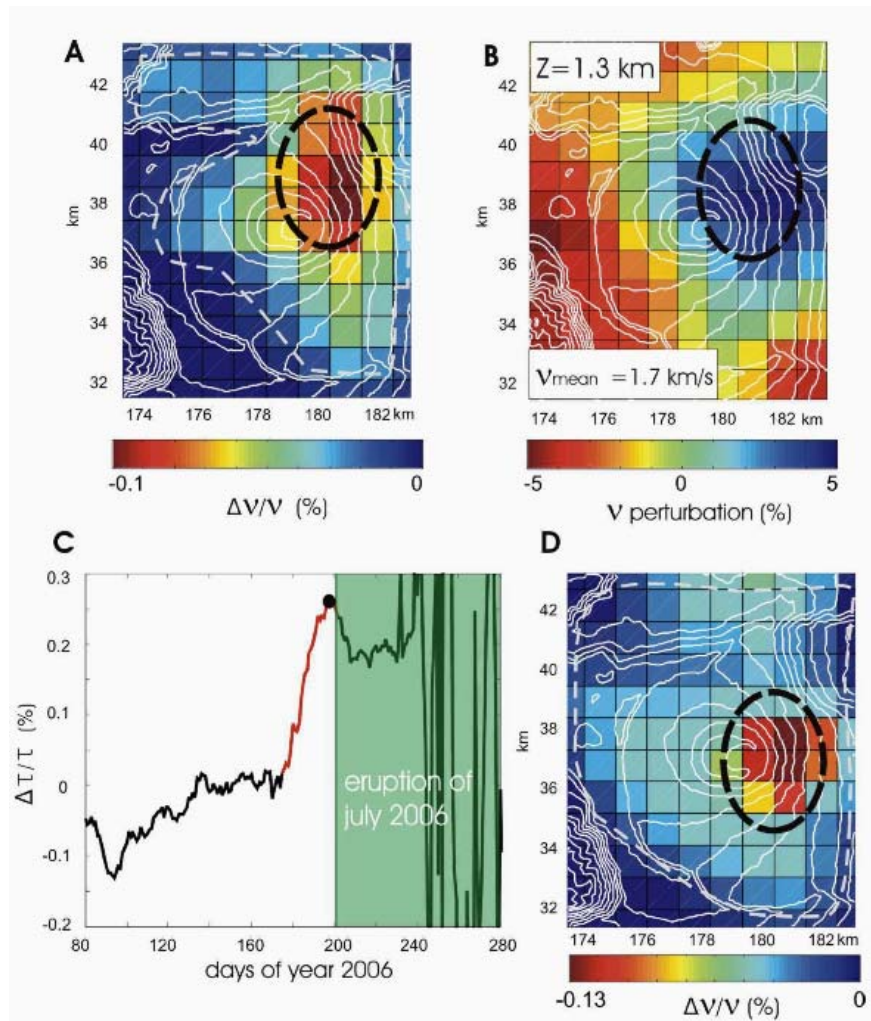
Also the migration of hypocenters during a high activity phase of a volcano is important in forecasting the subsequent volcanic eruption. In Fig. 13.28 an example of the 1991 Mt. Pinatubo eruption is shown. The migration of the seismic events from a cluster at 5 km depth north-west of the volcano in A) to a very shallow location directly underneath the erupting vent in B) is very obvious and possibly marks the ascending magma.



**Fig. 13.28** A) Mt. Pinatubo seismicity during May 6 to May 31. The seismic events are clearly clustering northwest of the volcanic center. B) shows the seismicity between June 1 to June 12 indicating a shift of the hypocenters to shallow depths and closer to the summit of Mt. Pinatubo (courtesy of Pinatubo Observatory Team (1991), EOS Trans. Am. Geophys Union, 72, 545, 552-553, 555)

#### 13.4.3.2 Cross correlation of diffusive wavefields

Weaver and Lobkis (2001a), Campillo and Paul, (2003) and Wapenaar et al. (2006) showed that correlating the coda of earthquakes which are azimuthally well distributed around a seismic station pair will result in an estimate of the Greens function of the subsurface structure between these two stations. Shapiro and Campillo (2004) also showed that the Greens function, or at least parts of it can also be retrieved when using the diffusive wavefield of noise. Sens-Schönfelder and Wegler (2006) used this property to monitor the seasonal changes at shallow depth on a volcanoes flank which is thought to represent different ground water saturation. Brenguier et al. (2008) showed, that monitoring the changes of the Green's function can be interpreted as a change in the velocity field at a volcano. This change is caused by the deformation of the volcanoes edifice caused by inflating or deflating magma. Fig 13.29 gives an example.



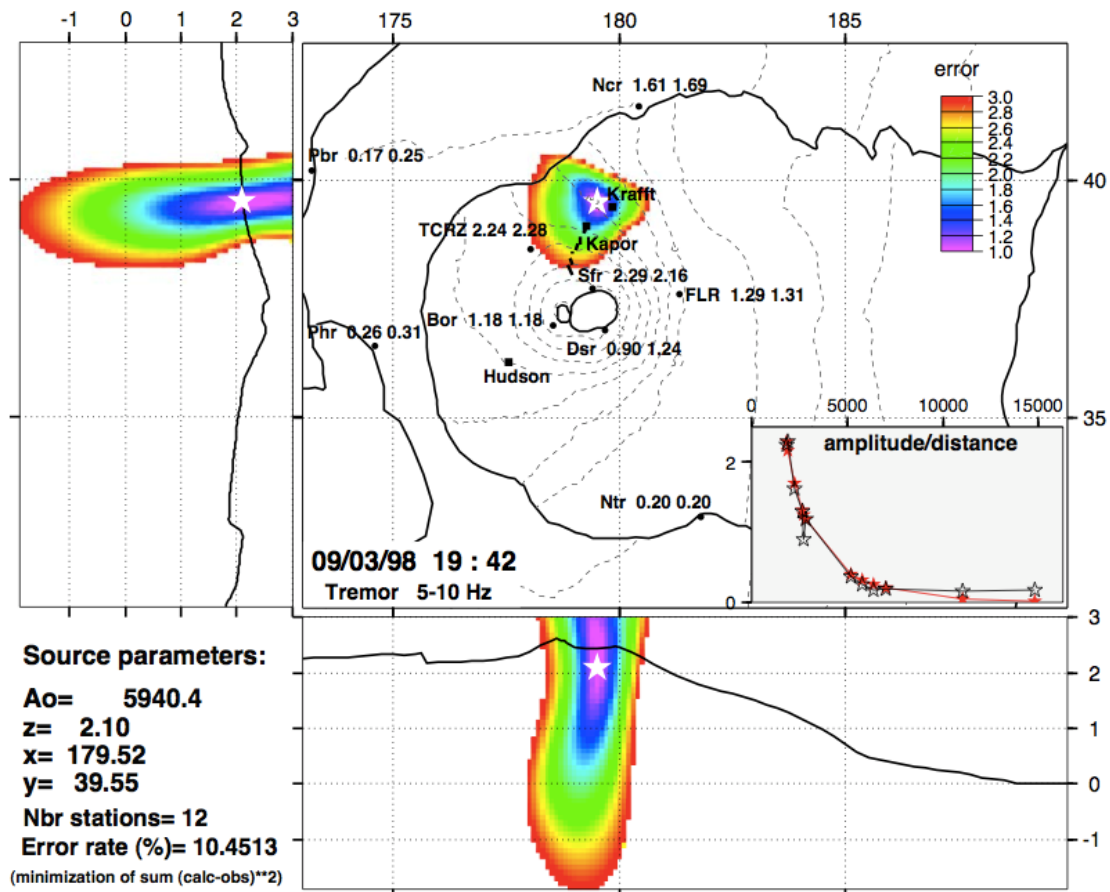
**Fig. 13.29** (A) Relative velocity perturbations associated with the second eruption precursor (day 85, Sept. 1999) at Piton de la Fournaise volcano (Reunion island). The gray dashed line represents the limits of ray coverage. (B) Fast shear velocity anomaly imaged by passive surface wave tomography (and interpreted as an effect of solidified dykes associated with the zone of magma injection). (C) Relative time perturbations before the eruption of July 2006. (D) Associated regionalized relative velocity perturbations one day before the beginning of the eruption (from: Brenguier et al., 2008)

While in praxis the technique is rather complex and sophisticated, parts of it are already included in daily monitoring routines (e.g., EARTHWORM, see Section 13.4.6).

### 13.4.3.3 Amplitude - distance based hypocenter determination algorithms

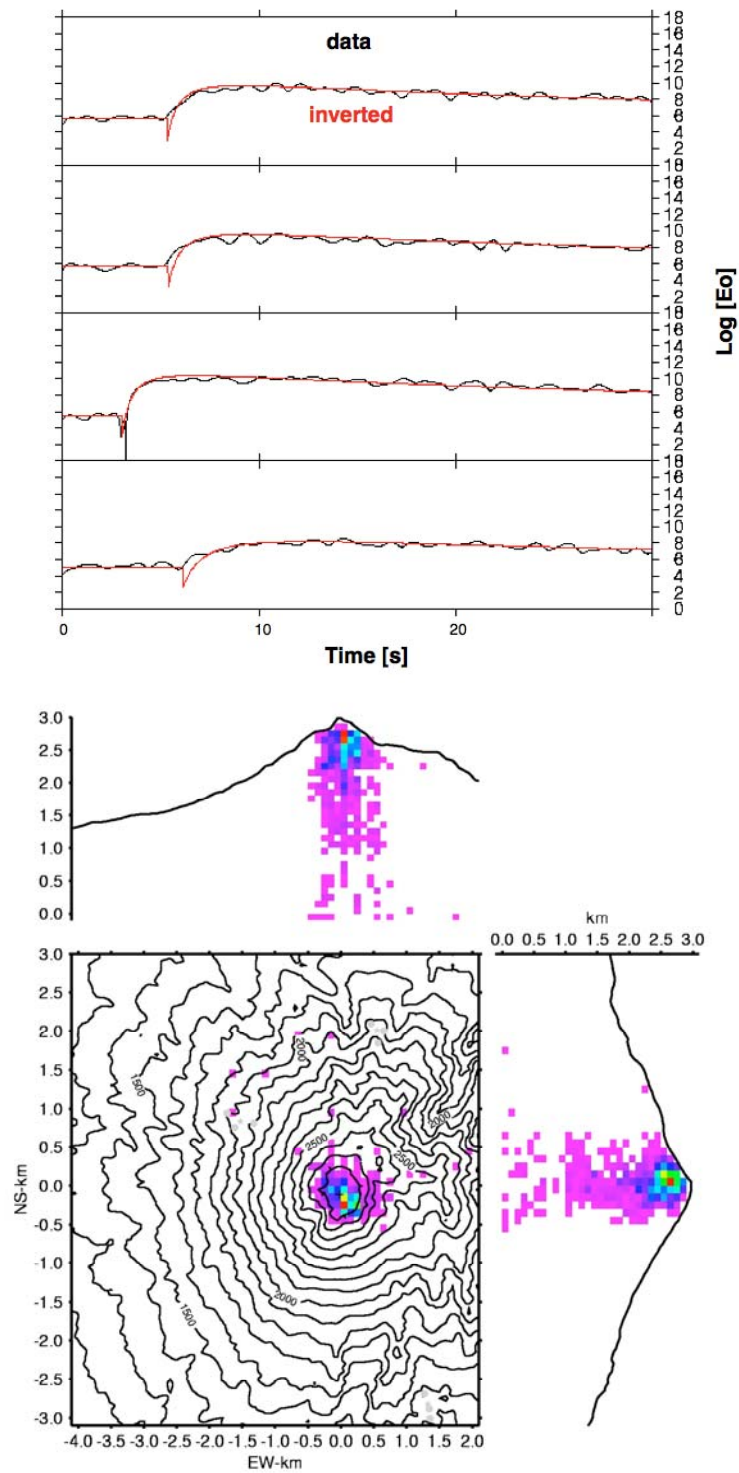
Even in the case of no clear P- or S-wave arrival, it is sometimes possible to estimate an approximate source area. Assuming a certain wave type (body or surface wave), neglecting an uneven radiation pattern of the source, and assuming a simplified propagation path, it is possible to compute amplitude-distance curves and model the source region. This can be seen as an iterative approach of fitting or contouring the amplitudes or radiated energy measured in the whole network. Successful applications of this technique were reported for locating volcanic tremor at Bromo volcano, Indonesia (Gottschämmer and Surono, 2000) and Mt.

Etna, Italy (Cosentino et al., 1984). However, care must be taken in the a priori assumption of the wave-type, i.e., body or surface waves. When locating the paths of pyroclastic density at Soufriere Hills volcano (Mountserrat), Jolly et al. (2002) were using a surface wave relationship. Battaglia et al. (2005) showed that volcanic tremor can be successfully located (see Fig. 13.30) by applying a rather simple amplitude-distance relationship. However, Battaglia et al. (2005) clearly pointed out that before applying this technique to the data a so called coda normalization has to be done in order to remove station-site dependent amplifications (Battaglia et al., 2005).



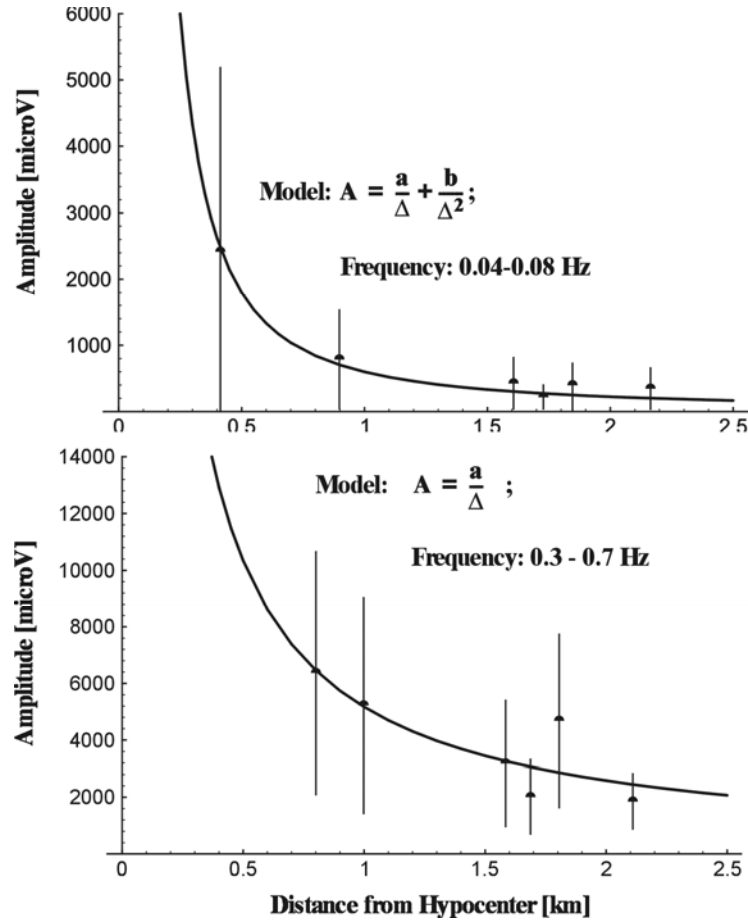
**Fig. 13.30** Example of an individual location in the 5–10 Hz frequency band for a tremor sample collected during the 1998 eruption of Piton de la Fournaise volcano (Reunion island). Map view and NS and EW cross-sections are shown. The result of the location is plotted as a white star, which is superposed on the normalized error contours. (from: Battaglia et al., 2005)

Wegler and Lühr (2001) and Friedrich and Wegler (2005) showed that the largest amplitudes visible in the seismograms recorded at the Mt. Merapi volcano are fitted best by assuming a strong scattering regime (see Fig. 13.31), which also alters the amplitude-distance relationship.



**Fig. 13.31** **a)** Envelope of MP event measured at different stations around Mt. Merapi and the result of fitting the scattering relationship of Wegler and Lühr (2001) to the data. **b)** shows the stack of locations of several hundreds of MP events using this amplitude distance relationship.

Also the influence of near-field effects may influence the amplitude-distance curve significantly (see Fig. 13.32).



**Fig. 13.32 Top:** Amplitude-distance relationship at Stromboli volcano. The amplitudes were measured in the frequency range 0.04 - 0.08 Hz. The best fit of the amplitudes at different distances from the active vent was obtained when an additional near-field term was added ( $A \sim 1/\Delta^2$ ). **Bottom:** Same as top diagram but in the frequency range 0.3 - 0.7 Hz. In this case the best fitting curve follows the usual factor of geometrical spreading  $1/\Delta$  for body waves. It is also obvious that in b) site effects are more pronounced than in a).

Amongst others the following relationships have been proposed:

- body waves (Battaglia et al., 2005):  $A^2 \sim A_0^2/r^2$ ;
- surface waves:  $A^2 \sim A_0^2/r$  (Jolly et al., 2002);
- near field:  $A^2 \sim (A_0/r + B_0/r^2)^2$  (Wassermann, 1997);
- scattering:  $A^2 \sim A_0^2 (1/(2\pi dt))^{3/2} \exp\{-bt - r^2/4dt\}$  (Wegler and Lühr, 2001)

#### 13.4.4 Seismic arrays

The analysis of seismic signals using array techniques is seen as the most prominent and emerging modern tool for locating volcano-seismic signals and evaluating the seismic wave-

field properties (e.g., Chouet, 1996b). While most of the array techniques are discussed in Chapter 9, we will focus on some results obtained when applying them in volcano monitoring and signal analysis.

The main deviation from typical array techniques in earthquake seismology is that the height differences between the array stations can not be neglected when the array is deployed on the flanks of a volcano. The effect of a 3D-distribution of stations can be minimized by fitting a plane to the station locations which dips according to the topography. One then transforms all auxiliary information (i.e., station coordinates) and refers all estimated parameters (incidence, azimuth and horizontal slowness) to this “best fitting plane”.

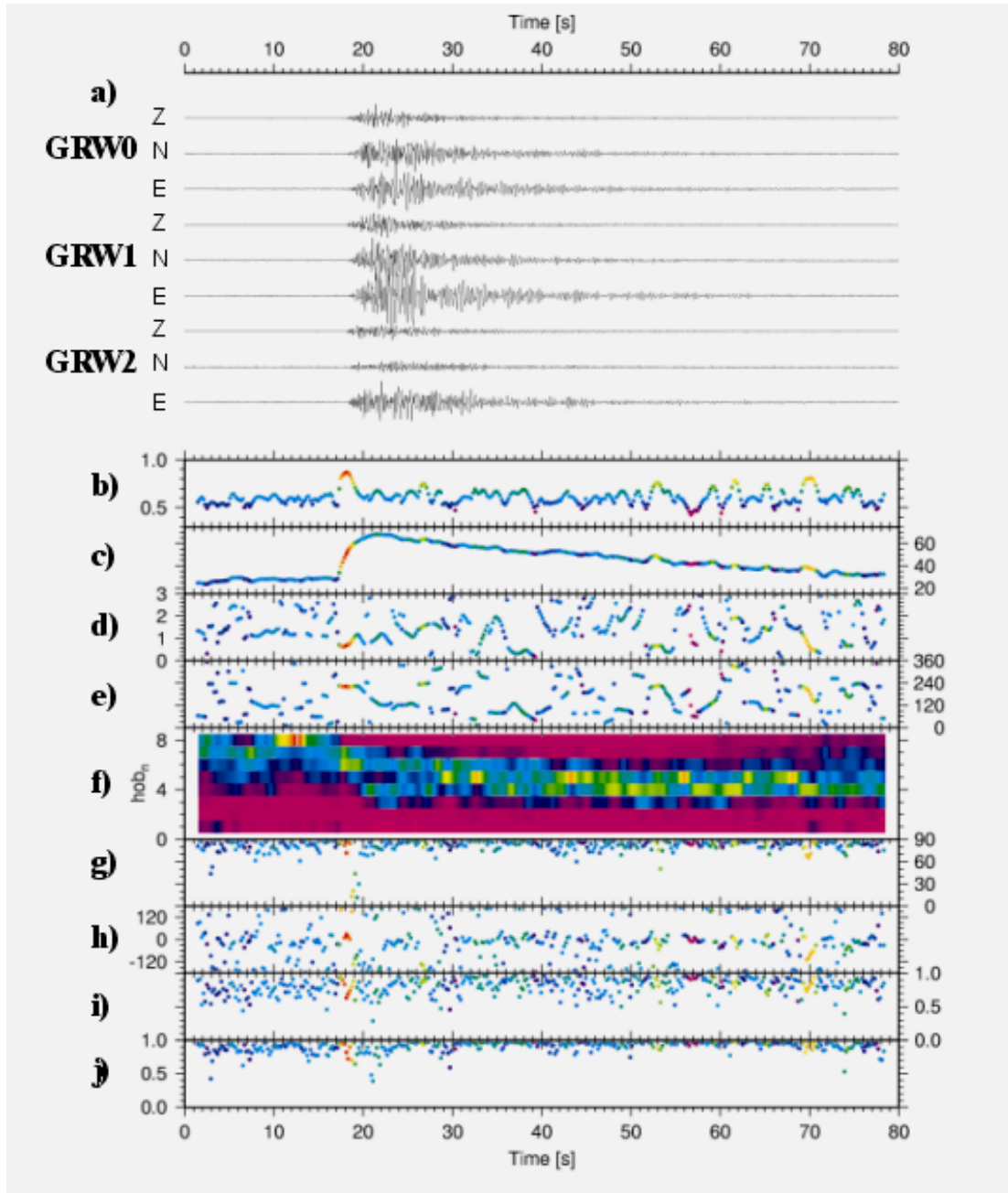
#### **13.4.4.1 f-k beamforming**

One of the most useful properties of a seismic array is its capability for suppressing undesired signals by filtering the incoming wave-field in the spatial as well as in the frequency domain. Thus, we can estimate the coherence, the signal power, the azimuth and the apparent velocity of an incoming wave. Because most seismic signals map into different regions of the frequency-wave number plane (see, e.g., Fig. 9.28, 9.38 and 9.40), f-k beamforming is an excellent tool to distinguish between the different wave-types. Beamforming can be thought of as delaying each seismic trace in time such that waves will add constructively when summed. The delay times necessary to obtain maximum “beam-power” is used to determine the direction of wave propagation through the array. Beams “aimed” in a direction, which is different than the true source direction will add destructively and produce a low signal. If the seismic array is located several wavelength from the assumed source area and the spatial extend of the array is small compared to the distance towards the source, we can also assume plane-wave propagation. Under these assumptions it is possible to estimate the backazimuth towards the source and, if the velocity model directly below the array is known, we can also estimate the incidence of the incoming plane wave. We can then invert for the source area of the signal (see Chapter 9, Sections 9.4.2 - 9.4.4).

Fig. 13.33a-e) gives an example of a broadband f-k analysis with data recorded at Mt. Merapi. Obviously only the very first part of the signal shows a phase with high coherence b), which additionally shows a small slowness c) (high apparent velocity). In contrast, later arrivals have randomly fluctuating backazimuth and slowness values. A possible interpretation of this pattern is that the recorded event consists of an array-wide coherent body phase (indicated by the high coherence and red color coding), which could be used for locating the event combining the backazimuth information and, if the velocity model just beneath the array is known, the incidence angle is estimated from the slowness. This coherent phase is followed by randomly incident waves.

Thus, the potential of a seismic array to discriminate between various types of incoming seismic waves and to quantify their properties makes the f-k beamforming perhaps the most powerful tool for investigation of continuous signals (i.e., volcanic tremor). Furumoto et al. (1990) and Almendros et al. (1997) showed the results of tracking a volcanic tremor source in space and time using seismic array beamforming. Other applications of large seismic arrays have been reported by Saccorotti et al. (1998), Chouet et al. (1997) and La Rocca et al. (2000), to name a few.







However, the application of beamforming techniques in volcano monitoring requires high computer power, rarely available during a volcanic crisis. In order to reduce it, one can use the spatial filter properties of seismic arrays and apply the f-k beamforming to steer in one or several directions of special interest (similar to beamsteering used to detect underground nuclear explosions; see Chapter 9, Section 9.6 and Section 9.7.7). Another way to reduce computational processing includes optimization of estimating the maximum of the beam in the f-k plane (i.e., highest coherence value) by using simulated annealing and/or simplex techniques (Ohrnberger, 2001).

#### **13.4.4.2 Array polarization**

As three-component seismometers are becoming the standard instrument nowadays and seismic arrays consist frequently of large numbers of three axial sensors, it is also possible to evaluate the polarization properties of the whole array. While there is no straight-forward method to include the polarization properties directly into the f-k-algorithm, it is possible to estimate array-averaged parameters of the 3D-trajectories. Jurkevics (1988) showed that array-wide averaging of the covariance matrices (see Section 13.4.2.2) results in a more stable estimate of the seismic wave vector (Fig. 13.27). Jurkevics (1988) also demonstrated the insensitivity of this estimate to alignment problems within the array. With this algorithm it is also possible to average the polarization properties over certain frequency bands which is a further link to the averaging properties of the broadband f-k-analysis (see Chapter 9, Section 9.7).

A further method to incorporate three-component seismic recordings of an array is to compute the “waveform” semblance (e.g., Ohminato et al., 1998; Kawakatsu et al., 2000). This approach consists of a grid search over possible source locations and the simultaneous rotation of the 3D ground motion vector towards these hypothetical sources. The semblance value of the L direction (L-Q-T system defining the direction from the source to the station L, the plane Q perpendicular to L including the source and receiver and the plane T perpendicular to Q) is computed. Assuming a source which generates solely a compressional wave in L direction the energy-density on the orthogonal components should be zero. Finally the “waveform” semblance should be 1 if the signal is coherent on all array stations and if no energy is left on the two directions perpendicular to L. The “waveform” semblance should be zero if there exists only incoherent wave-groups and/or there is still signal on the components different to L. It must be emphasized that this approach is restricted to cases where path effects and the influence of the free surface has no or vanishing influence on the orientation of the particle motion, i.e., low-frequency near-field observations.

#### **13.4.4.3 Hypocenter determination using seismic arrays**

As described in the Section 13.4.4.1 it is possible to track seismic sources in space and time using seismic arrays. Unfortunately, the exact seismic velocity distribution of a volcano is unknown. This results in large uncertainties in estimating the location of volcanic-seismic sources. One possible solution is to use not only one array but a network of arrays distributed around the volcano to compute the backazimuth of the coherent arrivals for each array separately and to invert them for the epicenter of the signal. Applications of this technique can be found in La Rocca et al. (2000).

Another difficulty arises when the seismic array is located close to the source and/or the height differences between the array stations are not negligibly. In this case the usually assumed plane wave propagation is no longer a good approximation and therefore the results are biased to an unknown extent, because neither the influence of the topography nor the deviation of the wave front from a plane wave is exactly known. In this case, a better way to localize the seismic source is to apply a more complicated approach, which first uses f-k beamforming to detect coherent phases within the continuous seismic data records and then apply two station generalized cross-correlation techniques in order to estimate the time difference of arrivals between the two stations. Wassermann and Ohrnberger (2001) successfully applied this technique to localize VT and strong MP events recorded at Mt. Merapi without the need for interactively determined onsets.

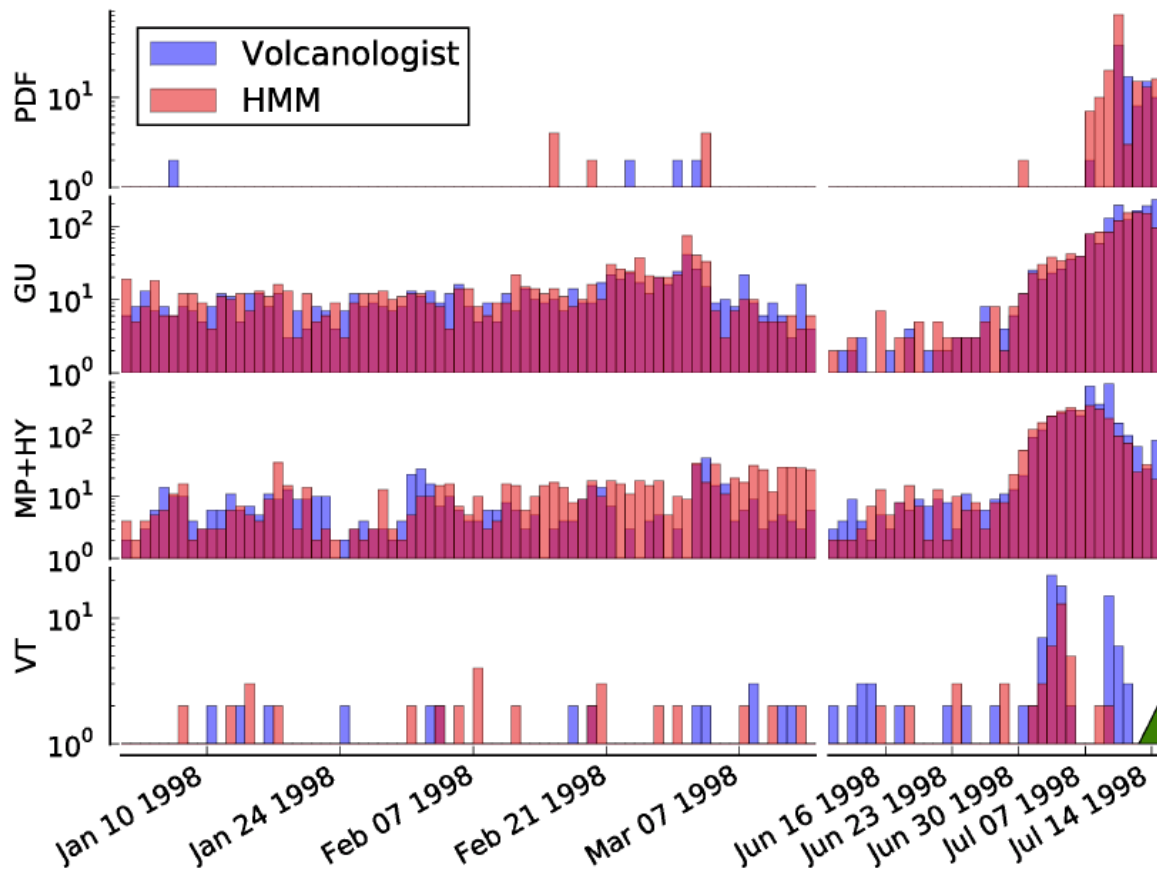
While this algorithm is only applicable to coherent and transient signals, algorithms exist which are based on the migration of coherent phases back to the source region (Almendros et al., 1999; Ohminato et al., 1998; Wassermann, 1997). Unfortunately, the computational load of these algorithms is high and their application is restricted to the “off-line” analysis of selected signals of special interest.

### **13.4.5 Automatic classification/detection**

When establishing a seismic network at an active volcano it is also possible to revise the classification scheme used (it would be possible even with a single station, but not very reliable). Besides the usually semi-automatic trigger and (visually) applied time-frequency analysis (see Section 13.4.1.1), we can also use wave-field properties obtained from the array analysis to enhance our discrimination quality significantly.

The advances in performing an automatic classification during the last 10 years are tremendous and can be separated in two major groups. Group one uses pre-triggered data, which are then classified by a combination of Self-Organizing-Maps (SOM) and Neural Networks (NN). See Scarpetta et al. (2005) and Esposito et al. (2008) for further information. The drawback of these methods is that the detection problem, i.e. to separate signal from noise, has to be solved in advance at least for the currently available algorithms.

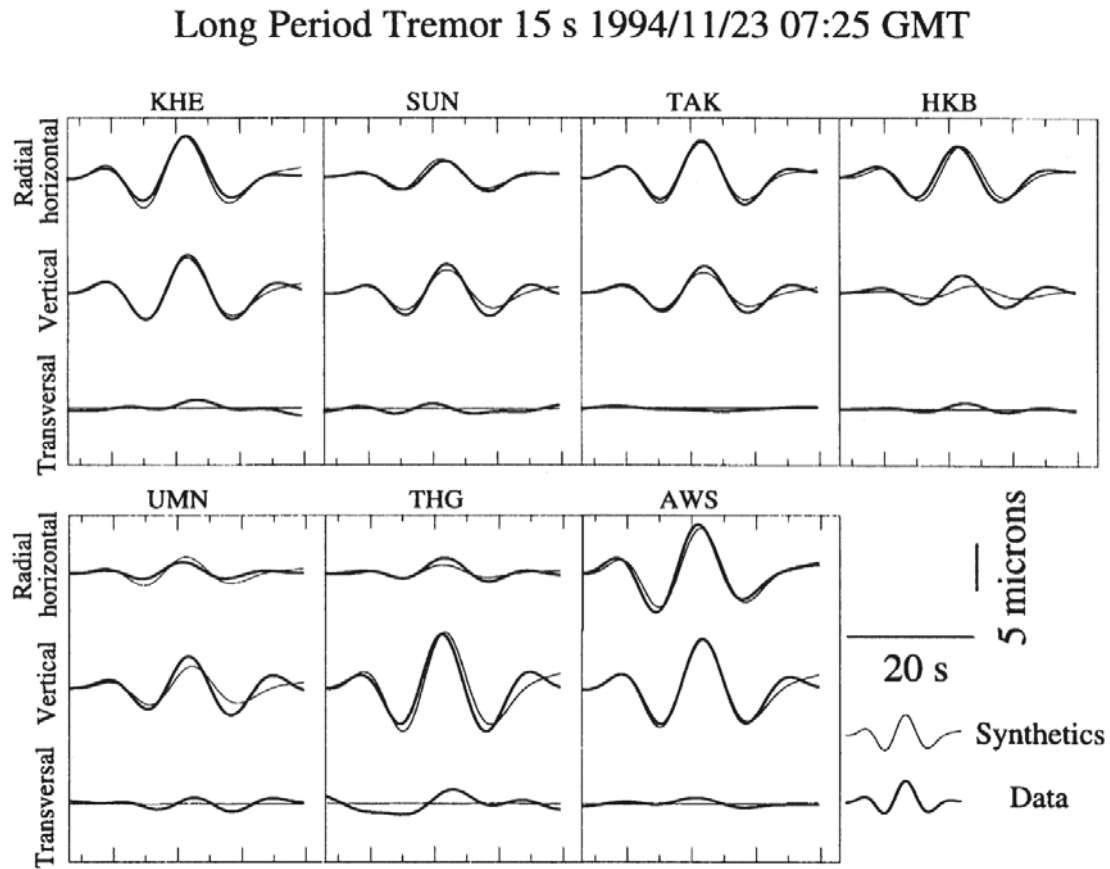
Ohrnberger (2001) was the first who applied Hidden Markov Model (HMM) based speech recognition techniques on parameters deduced from a continuous array analysis using data recorded at the Mt. Merapi volcano. Fig. 13.25 from Ohrnberger (2001) gives an example of the output of the continuous parameterization using seismic array techniques. The HMM approach solves the detection/classification problem in one single step by applying the double stochastic model which connects a Markov chain with the probability of observation parameters. Meanwhile a couple of applications exist (Beyreuther et al., 2008; Ibanez et al. 2009) where both groups are using the HTK-Toolkit for efficient computation of the classification probabilities. Fig. 13.34 gives an example of manual vs. automatic, HMM based detection and classification with data recorded in 1998 at Mt. Merapi volcano. Beyreuther and Wassermann (2011), also point out how the application of a slightly different stochastic model can improve the result significantly. In order to speed up the training and to ensure the flexibility of the classification, a guided training algorithm was developed as part of the EXUPERY project (see Section 13.6.5) by C. Hammer. This software is based on the python package ObsPy ([www.obspy.org](http://www.obspy.org), Beyreuther et al., 2010), can be incorporated into the EARTHWORM system (see Chapter 13.4.7) and is available via the project website [www.exupery-vfrs.de](http://www.exupery-vfrs.de).



**Fig. 13.34** Continuous classification using an automatic HMM based algorithm vs. human interpreter for seismic events from Mt. Merapi in 1998. The light red bars correspond to the classification results of the HMM classification, the blue bars correspond to the classifications of the local expert. The green triangle marks the onset of an eruption. The bars are transparent, thus a blue bar overlaid with a light red bar results in a purple bar (courtesy: M. Beyreuther).

### 13.4.6 Moment tensor inversion

In the recent years various approaches were made to investigate the dynamics of the different sources of the VLF and ULF signals using moment tensor analysis. While the estimation of the centroid moment tensor became a standard technique in earthquake seismology (e.g., NEIC and Harvard rapid moment tensor solutions), the application of this technique in volcano seismology is still restricted to specific applications or frequency ranges. The difficulties are manifold. First of all the influence of topography is neglected in the standard approaches, which results in large misfits of the computed synthetic Green's functions. Moreover, Ohminato et al. (1998) show that even when assuming a horizontal layered medium, the knowledge of the source location and velocity model with a high confidence is needed in order to apply this technique.



**Fig. 13.35** Data (thick) and synthetic (thin) seismograms calculated from a inversion of the seismic moment tensor for a single pulse of *long period tremor* at Aso volcano. The corresponding source mechanism consists of a large isotropic component (97%) in addition to a small deviatoric part (Legrand et al., 2000).

Compensated linear vector dipole solutions (CLVD) are often biased by the uncertainty of the assumed simplified velocity structure. However, there are some applications of moment tensor estimations with VLF and ULF signals which gave reliable results, indicating source mechanisms which deviate significantly from a pure double-couple solution commonly known of tectonic earthquake mechanisms (e.g., Fig. 13.35 from Legrand et al., 2000; Ohminato et al., 1998; Aoyama and Takeo, 2001; Saraò et al. (2001).

#### 13.4.6.1 Computation of Green's functions in 3D

The “day-to-day” increasing computer power combined with recent advances in computing synthetic seismograms (particle based methods: O’Brien and Bean, 2004; ADER-DG: Dumbser and Käser, 2006; Käser and Dumbser, 2006) makes it now possible to estimate Green’s function in 3D up to high frequencies for medium sized problems. Due to the still enormous computer cost, the application of these techniques is restricted to institutions with

sophisticated (parallel) computational infrastructure. One of the first attempts to include the topography of a volcano into the simulation process or Green's function calculation was published by Ripperger et al. (2003). In their simple stair case model the authors already showed the importance of topography as a prominent scatterer in the higher frequency parts of the spectra. For the definition of the Greens function please see Chapter 3 of this manual and the paper by Graves and Wald (2001).

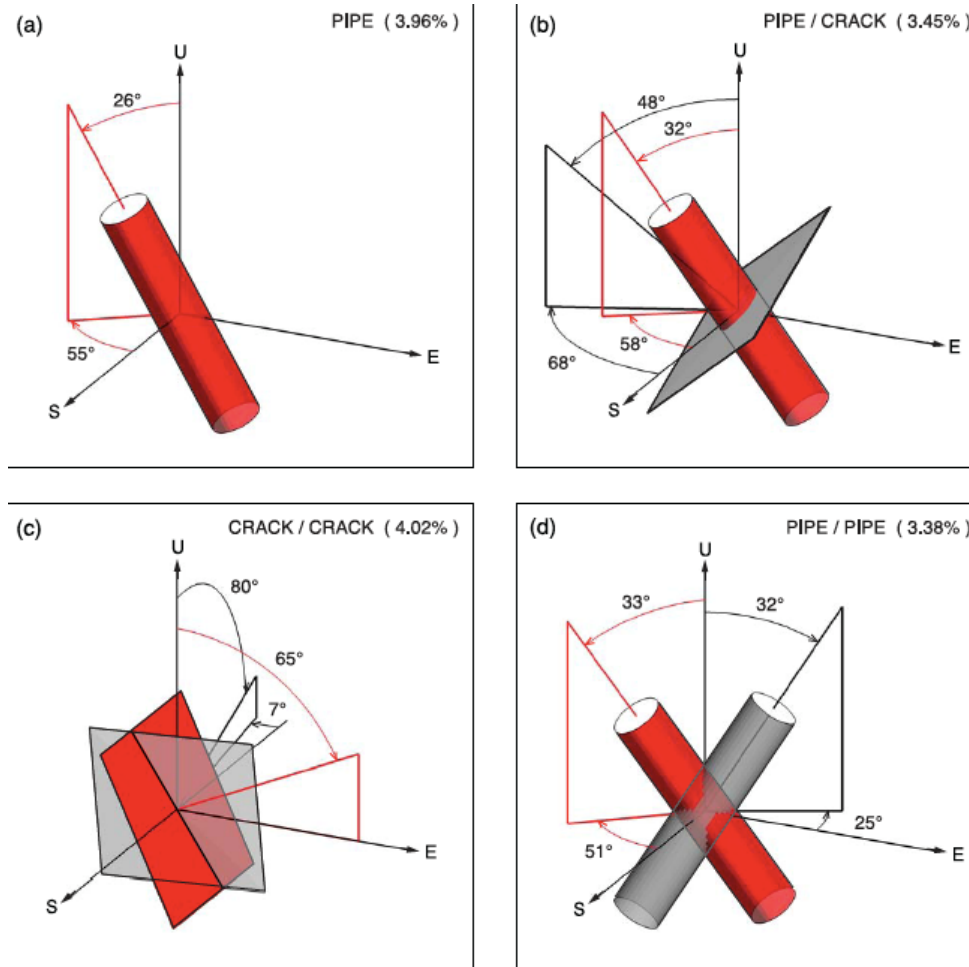
In a series of papers Bean et al. (2008), O'Brien and Bean (2004, 2009), Lokmer et al. (2007) and De Barros et al. (2011) impressively demonstrate the severe effects of topography, unknown velocity structure and source location on the inverted source mechanism. When not properly tested this will lead to completely wrong results where much of the error is projected in non-“double couple” parts of the inverted moment tensor. De Barros et al. (2011) also showed that the misfit or in other words “misleading forces” (see Section 13.4.5.2) can be reduced when observing volcanic signals in their near field and in a reduced frequency range. While scattering plays no important role for frequencies below, e.g., 0.2 Hz, the topography significantly change the Green's function and therefore, when not corrected for, causes severe errors in the resulting moment tensor. Additional sources of errors are uncorrected tilt signals especially when horizontal components of recorded seismograms are used and the unknown amount of strain-tilt coupling which is caused by local heterogeneities directly at the installation site of the deployed sensor (see Section 13.3.1)

#### **13.4.6.2 Source modeling of complex processes**

In theory nearly every seismic source known in earthquake seismology or related fields can successfully be modeled by an arbitrary moment tensor or by a combination of several moment tensors. However, when interpreting seismic signals of volcanic origin sometimes additional forces are needed in order to explain the recorded seismograms. Cesca and Dahm (2008) showed, that the 2005 paroxysmal eruption of Stromboli volcano can be modeled best by a combination of a moment tensor representing a CLVD (see chapter 3 for nomenclature) and a vertical orientated single force. The latter is often thought to represent the acceleration of the slug towards the surface. James et al. (2009) however show, that the single force is modeled best by a down flowing and thus accelerating viscous fluid.

A complete different explanation is given by De Barros et al. (2011). In numerical studies they can show, that the modeled single force often represent a sort of “trash can”, which includes all unknowns of the source process itself, the propagation media or instrumental problems.

Chouet et al. (2010), Chouet et al. (2008), Aster et al (2008) and De Barros et al. (2011) show that the combination of different geometrical bodies (cracks, cylinders, spheres), may lead to a successful minimized least squares representation of the seismic source (see Fig. 13.36). However, we should keep in mind that this representation only models the seismic part and may not explain the physics leading to a volcanic eruption correctly. Only few percent of the momentum during the ascent of magma or a volcanic eruption is transferred into elastic momentum, i.e., seismic waves. In order to shed more light into the process of volcanic eruption different measurements representing different physical aspects must be taken into account (see Section 13.5).



**Fig. 13.36** Dominant source component (red) and sub-dominant source component (grey) of a degassing burst at Kilauea volcano (Hawaii, USA). Different geometries are compared and the residual errors are shown in parentheses at the upper right (from Chouet et al., 2010)

### 13.4.6.3 Limitations of Moment Tensor Inversion

Many limitations were already mentioned while discussing the computation of Green's function and source modeling. Unknown velocity model, source time function, tilt motions on the horizontal components and/or tilt-strain coupling and our limited knowledge about the physics of ascending magmatic material will naturally lead to a rather blurred view of the source processes in a volcanic conduit.

Still unknown are also the influences of scattering and the violation of the commonly used point source assumption. While the latter may lead to a wrong representation of the forces involved, the scattering of the wavefield will gain more importance when the focus is put on the observation, processing and thus detailed modeling of higher frequencies. Possible ways out of this dilemma will lead to even more complex and costly (i.e., computational time) simulations and modeling of the recorded wavefield. The strong scattering usually present at active volcanoes, however, may form a natural frequency limit of interpreting the seismo-volcanic sources.

### 13.4.7 Automatic analysis and databases

During a seismic crisis or in the framework of a long-term seismic surveillance, it is not possible to apply all the analysis tools described above in a visually controlled, interactive manner. During the October 1996 volcanic crisis at Mt. Merapi nearly 5000 events/day occurred. This large number of events obviously precludes any on-line, interactive analysis of the seismic data.

There are various approaches to automate at least some parts of the routine analysis in a volcanic observatory (e.g., Patanè and Ferrari, 1999). The most prominent software package is called EARTHWORM (Johnson et al., 1995), developed mainly under the auspices of the U.S.G.S. Many of the techniques described above are implemented in this “real-time” environment, e.g., continuous spectral analysis, RSAM, SSAM, automatic event associations, hypocenter location and magnitude. Mainly designed for monitoring local earthquakes, its widespread use at volcano observatories has led to the development of new, volcano related modules and promises new tools in the future. The EARTHWORM system appears to be very flexible and capable of being adapted to special requirements at different volcanoes. However, the great flexibility of this software package entails rather complex and unwieldy setup procedures when establishing the system the first time.

Also quite widely used is the system SEISAN (Ottemöller et al., 2011, for the latest release: <http://seis.geus.net/software/seisan/>). In strict sense not an on-line system, SEISAN combines flexibility, open source character and a stable but simple interactive environment, which altogether form an attractive tool for the global volcanic community.

While writing the revised chapter of this manual, several new developments were made in the framework of automatic, near-realtime analysis of seismic signals at active volcanoes. Next to now established LINUX/UNIX/MAC and WINDOWS versions of EARTHWORM (Johnson et al. 1995, [www.isti.com](http://www.isti.com)) new systems are now available which may in future replace parts or the whole systems now currently running at volcanic observatories. E.g., seiscamp3, originally developed in the aftermath of the fatal 2004 Sumatra tsunami, is finding increasing applications within the global seismological and volcanological community ([www.seiscamp3.org](http://www.seiscamp3.org)). While the majority of the routines, especially the data retrieval (seedlink protocol: <http://geofon.gfz-potsdam.de/geofon//seiscamp3/>) is under GNU license, parts of the GUI, i.e., interactive parts are under binary license only. Currently seiscamp3 focuses on real-time analysis of seismic signals and thus needs a volcanological extension.

A volcanological extension of EARTHWORM and seiscamp3 offers the newly developed geophysical database SeisHub ([www.seishub.org](http://www.seishub.org)), which was partly developed during the Exupery-project (see Section 13.6.5). Using a combination of a SQL-database (incl. the command set) together with XML structured documents it forms a very flexible way to store and retrieve information with geophysical content. It is especially able to store all parameters and data described in Section 13.5, regardless of whether the data consists of time-amplitude values or sequences of 2D images. The SeisHub database extensively uses a package named ObsPy ([www.obspy.org](http://www.obspy.org)), a python based library for unified input/output and analysis techniques. With its possibilities to connect or include parameters from seiscamp3 and EARTHWORM as well the possibility to connect to data centers via arclink/fissures/seishub, ObsPy closes the gap between real-time algorithms and more specialized application needed in volcanology.



Also the World Organization of Volcano Observatories (WOVO; [www.wovo.org](http://www.wovo.org)) is currently developing a SQL/XML database hybrid (WOVOdat; [www.wovodat.org](http://www.wovodat.org)) able to store different parameters estimated or recorded at volcanic observatories around the world. The main focus of this data collection is to provide local observers with observations made at different volcanoes in order to enable a reliable estimation of unrest at “her/his local” volcano by comparing the surveillance parameters with that made at similar volcanoes.

## 13.5 Other monitoring techniques

As described at the beginning of this chapter, seismology is generally seen as the most reliable and diagnostic tool for monitoring a restless or erupting volcano. However, data from seismological surveillance alone are inadequate to understand and forecast eruptions. Modern monitoring systems will therefore combine seismology with other geophysical, geochemical, geodetic and geological techniques. In this section we focus on just a few of the various ground-based monitoring techniques which are closely related to seismology. We will not discuss the wide and fast-developing field of remote sensing in a volcanological context.

### 13.5.1 Infra sound

A method closely related to seismology consists in recording the infra sound produced by shallow volcanic processes. While the idea of recording infra sound is relatively old (e.g., Ripepe et al., 1996; Reed, 1987), the fields of application, instrumentation and theory is fast expanding. Most of the infra sound measurements are performed at volcanoes which show strombolian or hawaiianian activity. In this case the burst of large gas pockets (slug or annular flow) will produce large sonic signals which are recorded best in the infra sound range (Fig. 13.37).

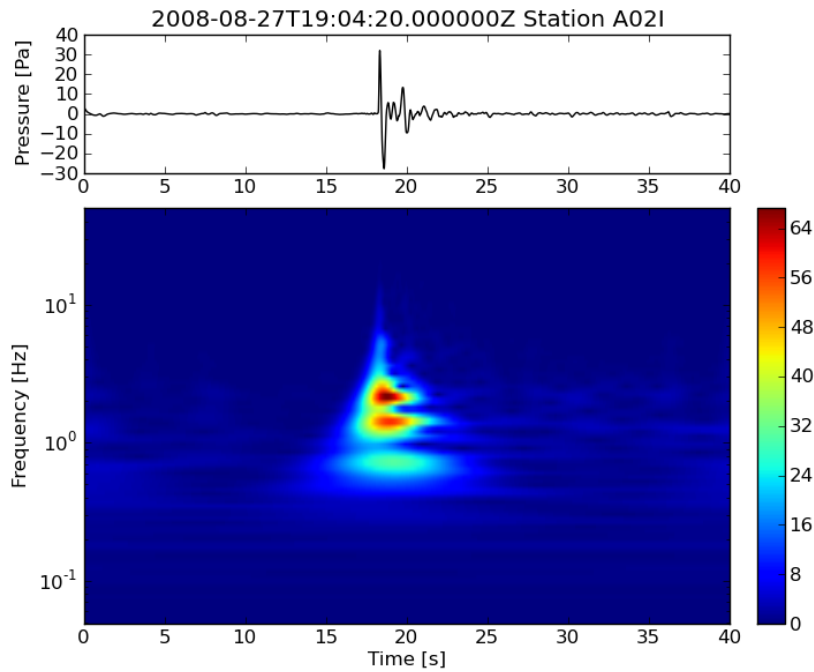
Johnson et al. (2008) and Gerst (2010) show how the recordings of infra sound can help to estimate the overpressures of the bursting gas slug. Gerst (2010) however, pointed out that infra sound recordings alone will give wrong estimates of overpressures and bubble burst dynamics. Using additional measures taken by a Doppler radar instrument Gerst (2010) was able to develop a physical model how the bubble is bursting at the magma air interface. While the dynamics of the bursting bubble in this model is controlled by the kinetic and potential energy of the expanding bubble shell, the overall budget is dominated by the thermal energy of hot lava ejecta (Gerst, 2010). The overpressures at Mt. Erebus volcano (Antarctica) remain small (3-7 bar) in relation to the violence of the eruptions. This is of great importance also for seismology at this type of volcano as it is then possible to better model the seismic coupling of the volcanic (fluid-gas) source to the elastic earth.

Also eruptions of more viscous volcanoes have been monitored using infra sound sensors. Caused by the different acting physics of ash jets and their associated sound the results are far less convincing than for their low viscous counterpart. Nonetheless, as a diagnostic of whether a volcano is erupting or the seismic signals are just reflecting an intrusion, infra sound recordings might help to discriminate and thus help to improve the early warning especially during bad weather conditions. Care should be taken, however, when installing an infra sound sensor (or sensor array). The reduction of wind noise or other unwanted signals is a difficult task, which might even act in favor of the installation of seismic broad band sensors instead. For details the reader is referred to Arrowsmith et al. (2010) for a very good overview of state

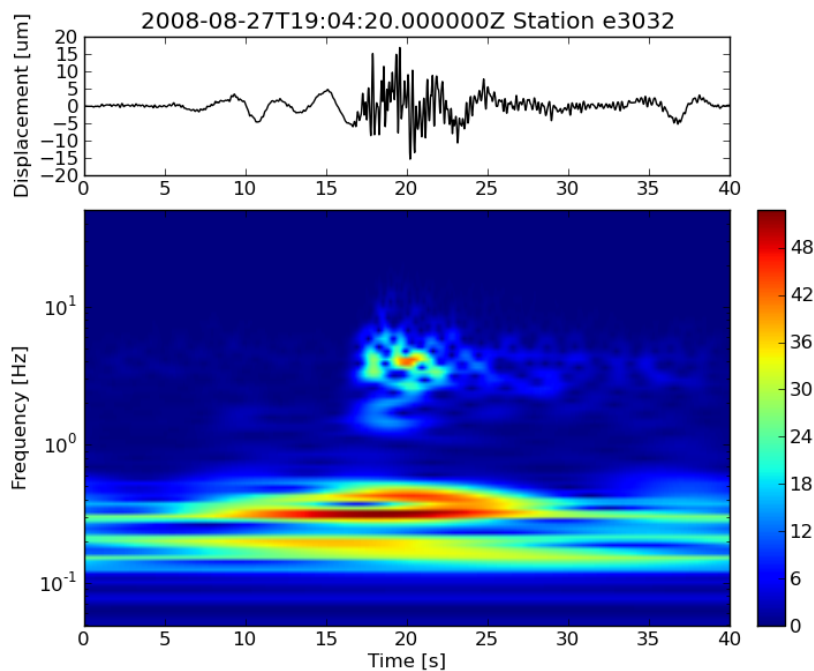


of the art research on seismo-acoustic wavefield and instrumentation.

a)



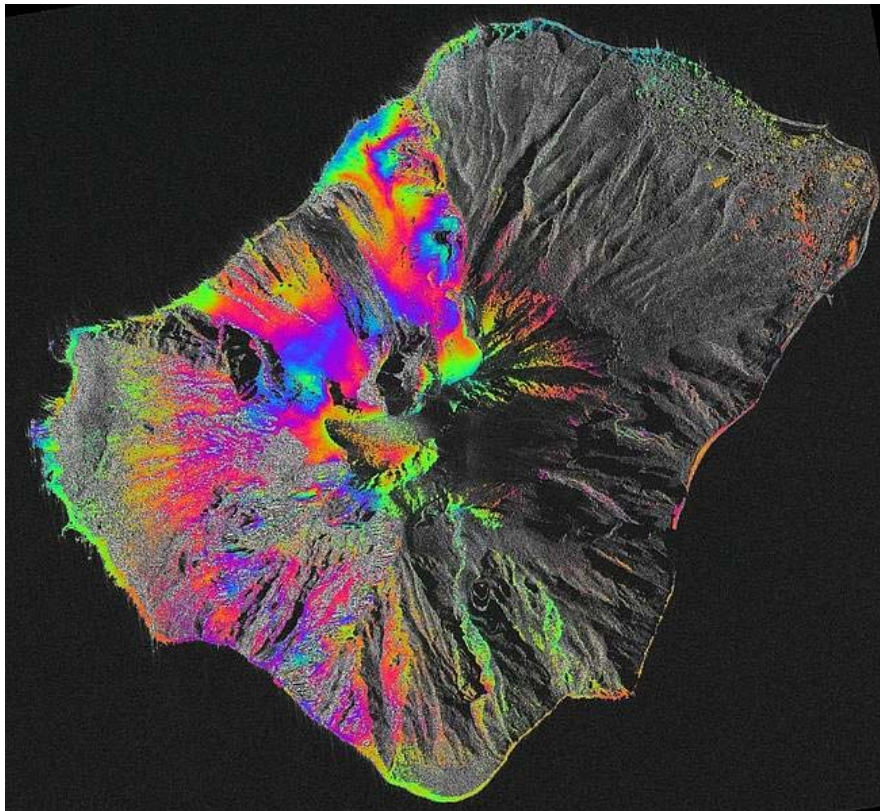
b)



**Fig. 13.37** Example of a strombolian type explosion recorded with **a)** an infra sound sensor and **b)** a seismometer at Mt. Yasur volcano (Vanuatu). The clear waveform of the infra sound signal is striking and supports the use and model of infra sound signals. However, the comparison with the seismic signal shows that, in this case, the latter already shows a significant amplitude at least 10 s before the onset of the acoustic signal.

### 13.5.2 Ground deformation (D-GPS, P-InSAR, strain, tilt)

Closely related to seismology is the monitoring of the deformation field caused by magma injection and/or hydrothermal pressurization within the volcanoes shallow or deep edifice. Deformation can be considered as an extension of seismology to lower, quasi-static frequencies, closing the gap of observed ground displacement to zero frequency. Modern techniques of monitoring the deformation signals of a restless volcano include borehole tiltmeters and/or strainmeters, electronic distance meter (EDM) and Global Positioning System (GPS) networks as well as airborne or ground based D-InSAR (Differential Synthetic Aperture Radar – Interferometry) measurements. Due to the increasing amount of positioning satellites (GPS, GLONAS) and accuracy, GPS will play an important role in the field of ground deformation monitoring over the coming decades. A relatively novel technique, which promises to have a large impact on the surveillance of active volcanoes are the airborne (airplane or satellite) or ground based differential InSAR measurements (Alba et al., 2008). While the airborne D-InSAR can cover a wide area with just one flight (see Fig. 13.38), the ground based D-InSAR has high repetition rate which makes the application of this technique very attractive during strong changes of the volcanoes edifice (Fig 13.39 for a non-volcanic example). For further reading on this exiting technique the reader is referred to text books of Hein (2004) or Hanssen (2001) and review articles by Bamler et al. 2009) and Bamler and Hartl (1998).



**Fig. 13.38** Composition of differential interferogram and the amplitude image from Stromboli volcano (Italy). The differential interferogram was calculated from the same pair of satellites and an additional SRTM based DEM. Differential phase with the coherence bigger than 0.72 are shown (courtesy: DLR Oberpfaffenhofen, Germany; from: [www.exupery-vfrs.de](http://www.exupery-vfrs.de)).

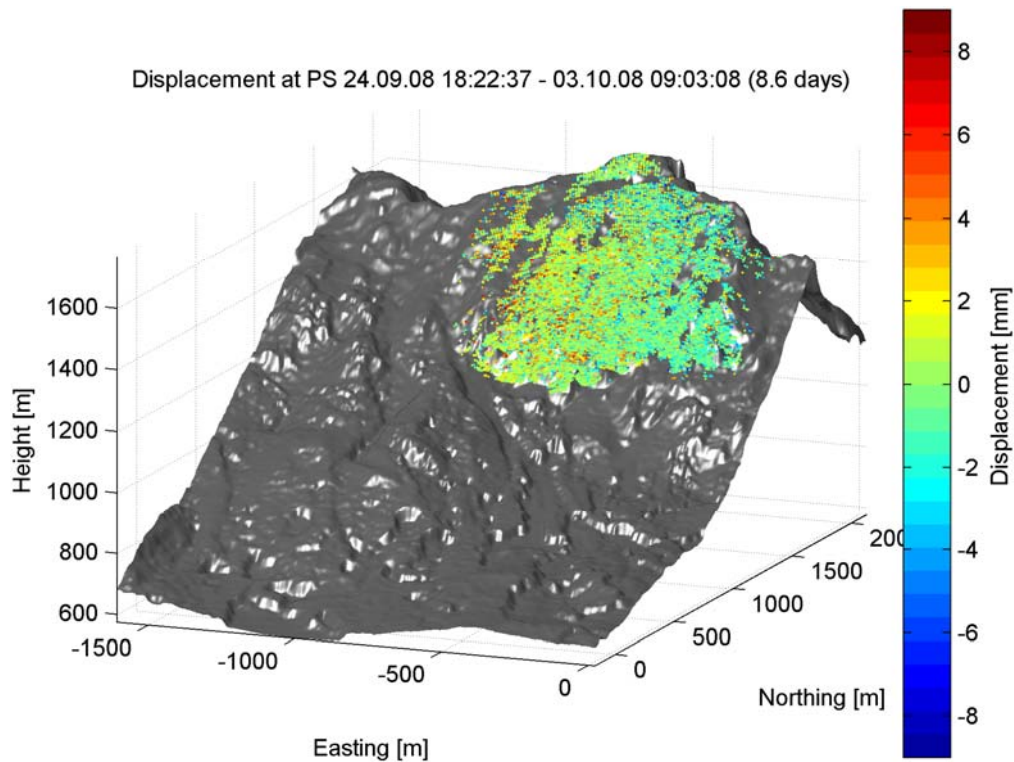
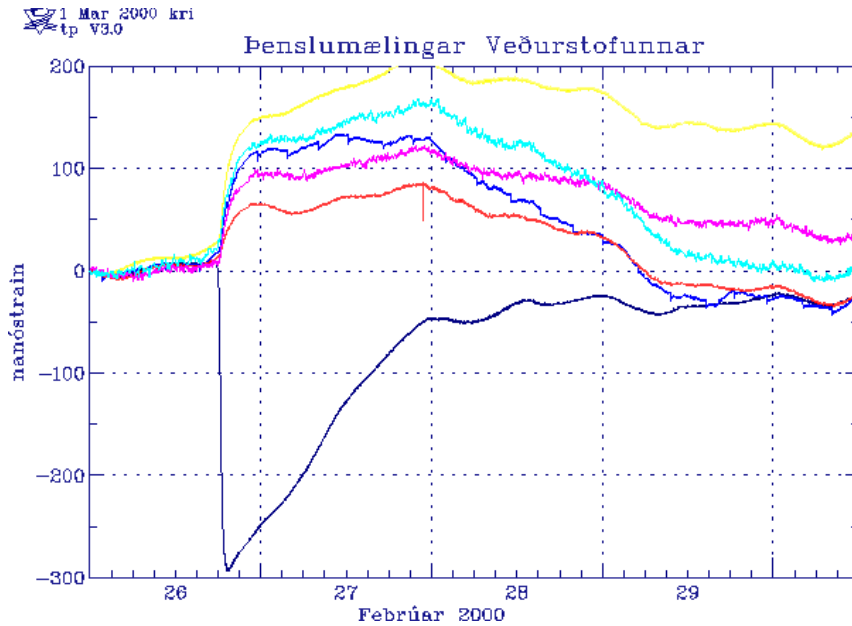


Fig. 13.39 Non-volcanic example of ground based D-InSAR measurement (Hochstaufen massif, Germany). While the covered area is small in comparison to flight based InSAR, the resolution and repetition frequency can be very high (courtesy: S. Rödelsperger TU Darmstadt, Germany).

A technique which is able to scan fast growing parts of the volcanoes edifice with even higher rates and precision, was originally designed to estimate the speed of particles expelled during strombolian eruptions (e.g. Hort and Seyfried, 1998; Hort et al., 2003). Vöge and Hort (2007) and Vöge et al. (2007) showed how this device, which consists mainly of a modified rain sensor commonly used in meteorology, could be adapted to monitor dome instabilities, which may lead to block and ash flows when this dome collapses. Recently Hort et al (2010) extended this method to also observe slow deformation processes by extracting specific phase information prior to eruptions at Stromboli and Santiaguito volcano.

Regardless what technique is used, the key point of monitoring ground deformation is the assumption that shallow or deep injection of large volumes of magma below a volcano will cause significant deformation of its surface. There were several successful approaches to forecast the 1980 eruption of Mt. St. Helens using deformation information (Murray et al., 2000) and at Hekla volcano, Iceland (Linde et al., 1993). The most recent 2000 eruption of Hekla volcano was accompanied by significant signals recorded by a cluster of strain meters located around the volcano (<http://hraun.vedur.is/ja/englishweb/heklanews.html#strain>). In addition, shortly before the eruption, increasing seismicity and volcanic tremor led to a precise forecast of the eruption (see Fig. 13.40). This can be seen as a perfect example of the complementarity of two different monitoring techniques. In addition there are many papers dealing with correlation between seismic signals and ground deformation at Kilauea volcano (Hawaii; e.g., Tilling et al., 1987)



**Fig. 13.40** The strainmeter data preceding the Hekla 2000 eruption are shown. As a result, three different phases could be defined. Firstly, a conduit was opened between 17:45 to 18:17. Secondly, a reduced rate of expansion of the same conduit at 19:20 could be detected and finally all stations showed an increase when the conduit was fully opened and magma was flowing directly beneath the volcano (courtesy of Icelandic Meteorological Office, <http://hraun.vedur.is/ja/englishweb/heklanews.html#strain>).

A further example of good correlation between measurable deformation and the appearance of seismic signals is known from Suffriere Hill volcano, Montserrat island, West Indies, where Voight et al. (1998) observed a coincidence between several swarms of Hybrid events with cyclic changes in the deformation signals. This coincidence is very important regarding the inversion of source mechanisms of this class of signals. This kind of deformation signal, in conjunction with magma intrusion into the volcanic edifice, is known to be very small and mainly related to the active part of the volcanic dome. At Mt. Merapi several clusters of borehole tiltmeters are installed at the flanks, but only very weak signals have been recorded (Rebscher et al., 2000). In contrast, strong deformation signals are visible at the volcanoes' summit stations (Voight et al., 2000). This might indicate that at Mt. Merapi no large-sized and shallow-situated magma chamber exists and that the volume of ascending magma during typical eruptive phases is small. However, tilt stations at the flanks of Mt. Merapi and other volcanoes with apparently small magmatic activity would be very useful for discrimination between the usual small magma intrusions and possible larger ascending volumes of magma (as in 2010), which will then produce a much more pronounced tilt signal at the flanks of the volcano.

This points to a dilemma of the surveillance of active volcanoes. Some measurement techniques will produce years of just baseline data, which might lead to a reduction of attention (e.g., in equipment maintenance, new investment, number of staff etc.) and instruments. However, if a larger, more threatening eruption is imminent these long-term measurements might be of crucial importance for reliable hazard assessment and risk mitigation.

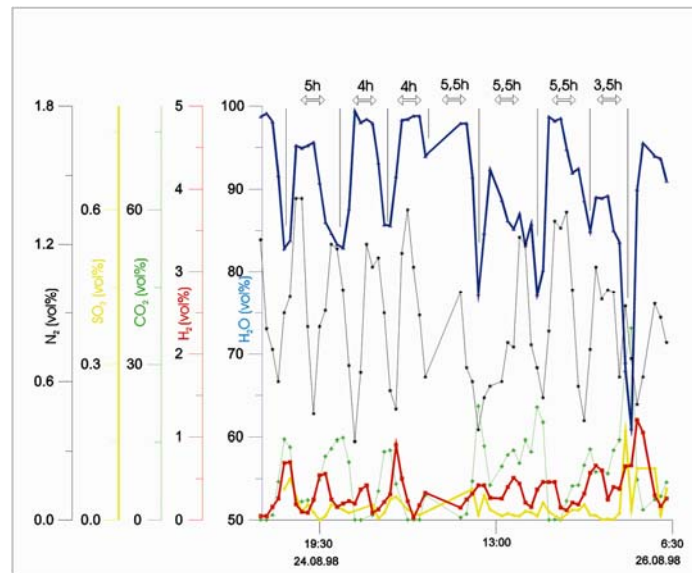


### 13.5.3 Micro-Gravimetry

The appearance of gravity changes at an active volcano also reflects possible inflation/deflation cycles of magmatic material. There is a complicated interaction between physical and geometric properties (i.e., density, volume, location) of the moving material and height changes caused by the deformation of the surface of a volcano. Therefore, the monitoring of gravity changes is a challenging task that should be carried out with great care. As height changes are a significant component of gravity changes, gravity monitoring should always be combined with high precision leveling (e.g., total stations -electronic distance meters attached to an optical theodolite- or GPS measurements). For a reliable and less ambiguous inversion of the gravity data, a good knowledge of the velocity/density structure of the volcano is needed. Models of the magmatic system from gravity data should be regarded with caution, unless the conclusions are also supported by other independent observations.

### 13.5.4 Gas monitoring

Another important parameter preceding a volcanic eruption is the volume, velocity, temperature and composition of the emitted gas from a volcanic vent or fumarole. Volatiles and released gases are seen as the most important driving forces for both an eruption and the source of volcanic signals (e.g., explosion quakes, LF, MP, volcanic tremor; Schick, 1988; Vergnolle and Jaupart, 1990). Different techniques of gas sampling are in use, ranging from routinely collected gas samples in a weekly or monthly manner to a continuous analysis (every 20 - 30 min) of the emitted gas using a gas-chromatograph (Zimmer and Erzinger, 2003). For example high sampling rate in continuous data at Mt. Merapi revealed surprisingly short period pulsations (with a duration of 3-5 hours, see Fig. 13.41) in the water to carbon-dioxide ratio as well as in the temperature of a fumarole (Zimmer and Erzinger, 2003).

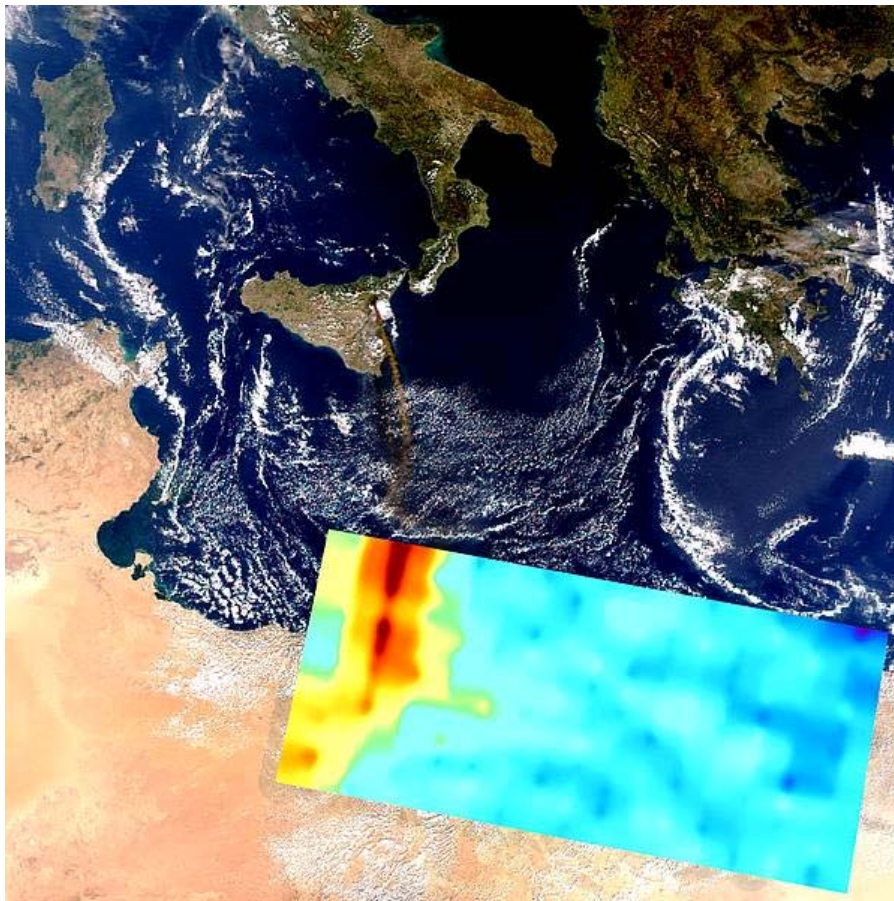


**Fig. 13.41** Variation of the gas composition at one of Mt. Merapi's fumaroles. The gas is automatically analyzed approximately every 30 min using a gas-chromatograph. The analysis shows a fast changing composition of the gas with a period of roughly 5 hr. This special fumarol field (Woro) was altered/destroyed during the 2006 and 2010 eruptions of Mt. Merapi (courtesy of M. Zimmer, GFZ-Potsdam).

However, no significant correlation between this pulsation and the related seismicity could be found. The rhythms in this pulsating gas source only changed when the number of very shallow MP-events also increased. This lack of correlation between fast sampled gas data and seismic signals at Mt. Merapi might be caused by the our imperfect knowledge of how to parameterize the seismicity and the gas composition, respectively. Several case studies of changes in the chemical composition of fumarolic gases including descriptions of the accompanying seismicity and ground deformation is given by Martini (1996).

Many other papers deal with the long-term variations of gas prior to a volcanic eruption, which makes this technique a useful tool for long term monitoring (e.g., Stix and Gaonac'h, 2000).

Sulfur dioxide measurements from satellites make it possible to monitor volcanic activity and eruptions on much larger, even global scales. E.g., the Global Ozone Monitoring Experiment (GOME-2; carried by the MetOp-A satellite launched in 2006) provides, next to other gases, SO<sub>2</sub> columns with a global coverage within 1.5 days and a spatial resolution of 40 x 80 km (Fig. 13.42)



**Fig. 13.42** Space-based observations allow near-real time detection of volcanic activity: SO<sub>2</sub> columns from GOME-2 as overlay on a MODIS (Moderate Resolution Imaging Spectroradiometer) imagery of Mt. Etna volcano, Italy (courtesy: DLR Oberpfaffenhofen; from: [www.exupery-vfrs.de](http://www.exupery-vfrs.de)).

The satellite based volcanic SO<sub>2</sub> emissions are determined from solar backscatter measurements in the ultra-violet spectral range. In so doing, the Differential Optical Absorption Spectroscopy (DOAS) is applied, a technique, which may also be used as a ground based installation. The ground based counterpart has again much higher repeat rates and is thus a very valuable technique when a volcano shows first signs of unrest. The main goal of the satellite based technique is to monitor and forecast the dispersion of volcanic plumes up to 3 days and to attribute the detected SO<sub>2</sub> to a particular volcano. The final goal is than to facilitate a global warning service using SO<sub>2</sub> data. Over the past couple of years DOAS measurements are starting to be replaced by so called SO<sub>2</sub> cameras (see e.g. Bluth et al, 2007; Mori and Burton, 2006), at least for the determination of the SO<sub>2</sub> concentrations. They allow a 2D measurement at a rather high temporal resolution (up to 10Hz).

### **13.5.5 Thermal and video recordings**

Monitoring of volcanic activity using radiometer or other imaging systems has been increasingly used over at least the last 25 years from space-borne or ground based remote sensing infrared sensors. The primary goal is the detection and analysis of volcanic hot spots and plumes at remote volcanoes. Modern sensors are now capable of resolving thermal anomalies even allowing the detection and monitoring of smaller effusive events. Over the last decade most observation from space of volcanic eruptions were carried out using data from AVHRR, GOES, MODIS, ASTER, BIRD, and Landsat-7 / ETM system. The different systems differ mainly in their spatial resolution ranging from 8×8 km for 6.47–7.02 μm band on GOES down to 15×15 m pixels for the VNIR bands on ASTER. Also the repeat times varies also strongly, with geostationary systems sending images every 15–30 min in contrast to the ASTER/MODIS system with repeat cycles of 16.5 days. At some volcanoes stationary, ground based infrared cameras proved to be useful to monitor changes in the volcanic system. Meanwhile cheap(er) and less sensitive devices are on the market which makes a remote installation at a sight with visibility to active vents possible (e.g., Stromboli, Popocatepetel etc.).

A lot more material can be found on the web page [www.exupery-vfrs.de](http://www.exupery-vfrs.de).

### **13.5.6 Meteorological parameters**

While not directly linked to the eruptive behavior of a volcano, monitoring meteorological conditions is important for the proper interpretation of observed parameters as well as for the anticipation of possible triggering of volcanic activity. Lahars, i.e., volcanic debris flow, are often triggered by heavy rainfall, which additionally weakens the unconsolidated volcanic material. At Mt. Merapi, small-size gravitational dome collapses take place more frequently during the tropical rainy season.

The influence of meteorological conditions on the installed monitoring equipment is manifold. Barometric pressure and temperature changes could cause severe disturbances on installed broadband seismometers and tiltmeters, respectively. While these influences can be reduced by proper installation of the sensors (see Section 13.6), it is never completely removed. Spectral analysis of both meteorological and surveillance parameters may help to identify possible disturbances of the installed sensors. As mentioned before, rainfall may trigger volcanic as well as seismic activity. A good monitoring station will therefore also have

a continuously recording rain gauge.

In order to better judge the influence of meteorological (and also tidal) effects on the sensors and/or the volcanic activity, continuous long-term meteorological recordings are needed. Frankly speaking this is a difficult and sometimes impossible task because of the harsh environments at many active volcanoes. However, the continuity of such measurement are among the most important functions of a volcano observatory.

## **13.6 Technical considerations**

In this chapter, some baseline information regarding the technical design of seismic surveillance network is given.

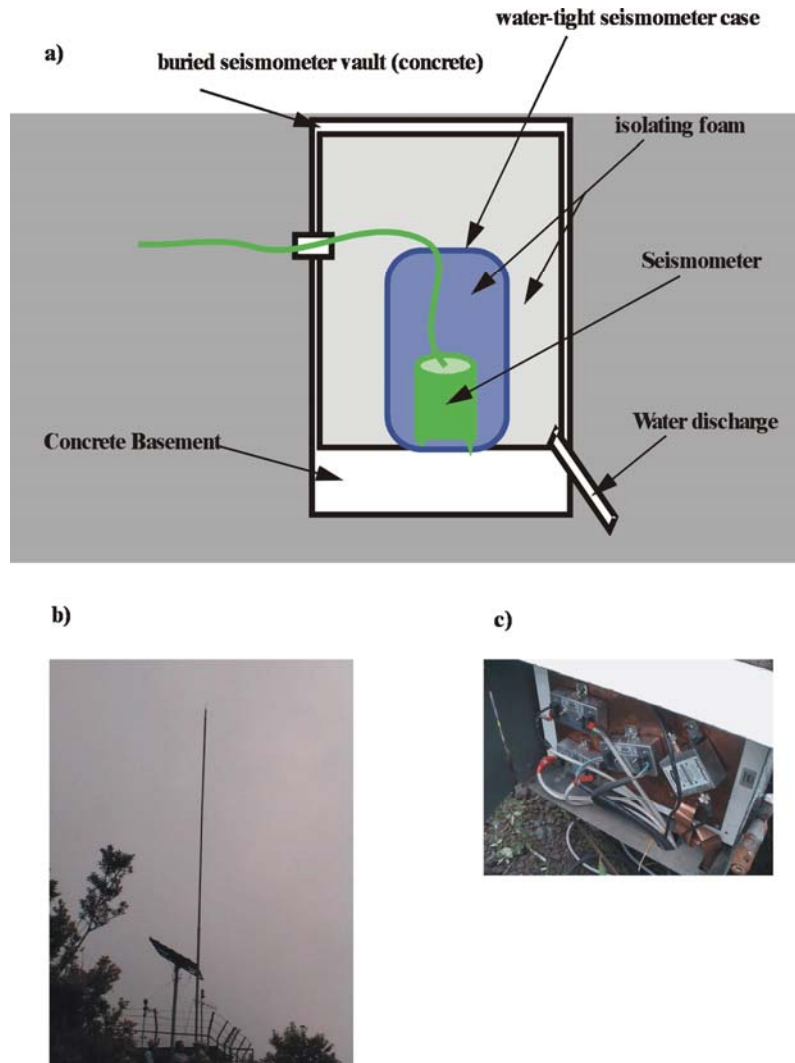
### **13.6.1 Site and vault**

After selecting a possible site (see Section 3.1) for a seismic station or a seismic array, significant care should be taken to protect the sensor from meteorological and other external effects. In Fig. 13.43a, a sketch of a possible installation scheme is shown. If a broadband sensor is to be deployed, even more care must be taken to protect this sensitive sensor from temperature and barometric pressure influences (see Chapter 5, Section 5.5 and Chapter 7, Section 7.4). Installations cover the sensor completely in sand in order to avoid instrument near air convection, which may lead to long period noise. Please inspect your instrument manual if this method will work for your instrument.

The weather conditions at volcanoes at even moderate altitudes can be very rough and rapidly changing. Protection against rain and lightning is the most important task when constructing a seismic station. All equipment should be placed in water tight casings, however, a water outlet should always be present. Most of the time it is harder to remove the water condensate from the vault than protect it against meteoric water.

Lightning protection is the most important and, unfortunately, the most difficult problem to solve (see Chapter 7, Section 7.4.2.5). Commonly, volcanoes have high resistivity surface layers (ash, lapilli etc.), making a proper grounding of the instruments nearly impossible. One of the optimal techniques to protect the equipment against lightning damage is to install a tower at the vicinity of the station with a mounted copper spire on top. The tower should be grounded as much as possible and be connected entirely with the ground of power-sensitive equipment. Furthermore, lightning protectors should be placed in front of any equipment to reduce the effect of high-voltage bursts (see Fig. 13.43b-c). Long cable runs should be avoided or changed to fibre optics. Using fibre optic cables for signal transmission has also the advantage that they are insensitive to electro-magnetic effects, which sometimes cause spike bursts on the transmitted signals.





**Fig. 13.43** **a)** sketch of a seismometer vault. The sensor should be placed in a water tight casing which is placed firmly on a concrete basement. In order to isolate the sensor from temperature and pressure changes due to air turbulences, the void space should be completely filled with insulating rubber foam or similar (see Chapter 7, Section 7.4.2). **b)** shows a lightning tower installed at Mt. Merapi, while **c)** shows additional lightning protectors for all sensitive equipment. The copper plate and all external devices (photo-voltaic modules) should be connected to the lightning tower.

### 13.6.2 Sensors and digitizers

Because the seismic signals produced by an active volcano cover a wide dynamic range the choice of the digitizer, i.e., the required dynamic range, should be carefully evaluated. Modern digitizers will sample the analog seismometer output with 24 bit resolution which results in a dynamic range of roughly 136 dB (depending on the sampling rate). A 16 bit A/D converter would usually be sufficient, but eruptive phases with various large amplitude signals will then saturate the digitizers' dynamic range (e.g., pyroclastic density flows, big explosions etc.). So called "gain-ranging" (i.e., the pre-amplification will be lowered if the signal is getting stronger) should be avoided because it will result in lower resolution and might mask small but important signals. Best suited are digitizers which sample with 24 bit

resolution but store the data depending on the recorded peak amplitude (i.e., 8 bits are stored if the signals are small and the activity is low, 16 bits when the activity is increasing and 32 bits if high activity occurs and the full 24 bit range is used).

If a network of seismic sensors is planned, several three-component short-period instruments (i.e., with 1 Hz corner frequency) will be sufficient (see Section 13.3.2). If the near-crater range is accessible, installing one or two broadband stations (i.e., 0.00833 to 0.05 Hz corner frequency) is a good choice. If large ( $M > 4$ ) earthquakes from an active volcano flank or nearby fault or subduction zone are possible, some broadband stations are preferred.

If an array or a network of arrays is to be installed, a mixture of three-component broadband and short-period one-component seismometers will be sufficient (especially when considering that there is no straight forward technique available which involves directly 3D seismic array data). A possible four station array would consist of a central three component station while the three satellite stations are equipped with vertical component sensors only.

### **13.6.3 Digital telemetry and satellite communication**

Most of today's established monitoring networks at volcanoes are designed for transmitting the data "on-line" to a central data center, generally the local volcano observatory. This might be the main technical difference between a short-term seismological experiment and the long-term monitoring of a volcano. From experience worldwide, establishing a reliable radio line is a difficult and time consuming task, which also is subject to changes when new telecommunication facilities are constructed nearby and possibly worsen the data communication.

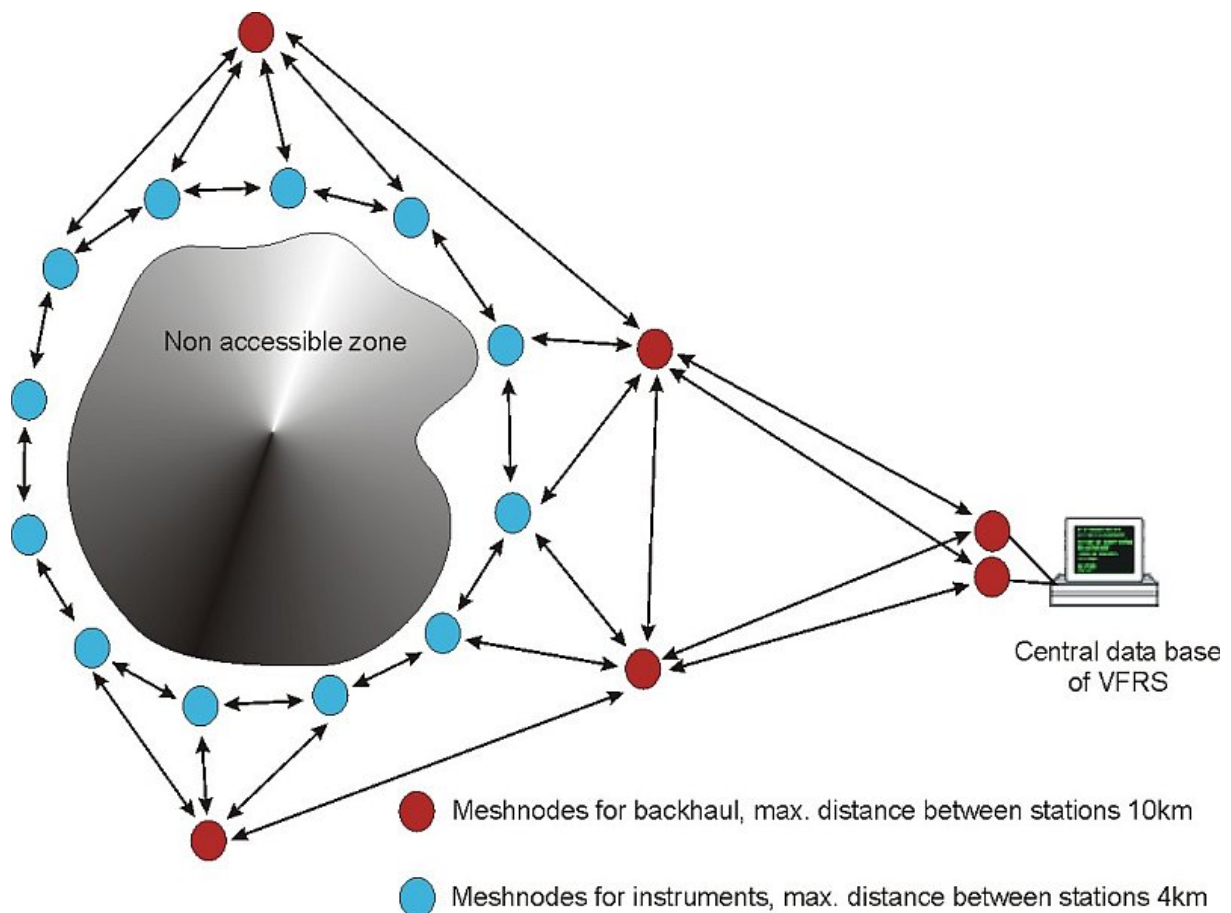
There are large differences between analog and digital radio transmission regarding data rate, dynamic range and sites to be selected and distance ranges to be covered. If a high resolution is required (e.g., using a 24 or 16 bit A/D converter at the sensor), the only way to exploit the full bandwidth of data is to transmit the signals with a digital radio modem. Meanwhile, several companies offer spread-spectrum modems which transmit in the frequency range of roughly 1 GHz and 2 GHz, respectively. The big advantage of these digital modems is the high data throughput (115,200 baud) and the low power consumption (transmitting power roughly 1 Watt). The high transmission frequency, however, is the main drawback of the digital radios. Commonly, the station and the data center must be in direct line of sight, with no hills, trees or other obstacles between them. This limitation should also be kept in mind when selecting a suitable seismic station site. The problem of obstacles can be circumvented when installing several repeaters on the way to the data center. Even so, the network design depends on intended radio lines (see also Chapter 7, Section 7.3 and Information Sheet IS 8.2).

A disadvantage of analog radio communication is the limited dynamic range and data throughput (usually below 38,400 baud). Most of the installed analog radio systems are barely able to transmit 12 bit and, therefore, the signals must be bandpass filtered (e.g., 1 - 20 Hz) before transmitting it. This is not acceptable when installing a broadband sensor. However, the radios are cheap and the typical frequency bands (100 MHz or 400 MHz) will enable a solid radio link even when the stations are slightly "out of sight".

Wireless LAN technology lately took over a large share in the communication business.

While in many countries the maximum applicable power output is restricted to 100 mW an intelligent use of different antennas for transmitting (low gain) and receiving (high gain) will allow distances of up to 10 km without the use of a repeater.

Even more suited for a communication network at active volcanoes are so called self organizing meshes. These ad hoc networks are able to organize their communication paths by themselves just by giving a central device the IP number of the transmitting station) see Fig. 13.44). If one of the nodes (transmitters) is lost, the network will re-organize itself to reach the stations by a different path. The drawback, however, is the rather complicated setting of the network as well as that potential station sites are chosen based on transmission properties rather than geological/geophysical evidences.



**Fig. 13.44** Schematic principle of operation of a Wireless-LAN mesh. In case one of the nodes is not operational the data from a particular station is transmitted using a different path to the data center (courtesy: M. Hort; from: [www.exupery-vfrs.de](http://www.exupery-vfrs.de))

Meanwhile also many different satellite communication devices (V-SAT, Sky-DSL) are available. Due to the still high power consumption and rather sensitive equipment the use of these devices is restricted to certain points within the network (e.g., connection of the observatory to the international community; special node of high throughput).

### 13.6.4 Power considerations

Most of the stations will be remote and no access to a power network will exist. Therefore, the first step is to calculate the expected power consumption of the seismic station. It strongly depends on the kind of digitizer and sensor used, whether the data are transmitted by radio or not and which options for local data storage are desired. Therefore the power consumption of the field equipment should be extensively tested in the lab before constructing the power supply at the site and deploying the instruments. Never trust the optimistic specifications given by the manufacturer!

Most likely the power will be delivered by photo-voltaic (PV) modules, where a variety of different systems is available. All components of a monitoring installation must fit together, including the capacity of batteries and solar charger. Care must be taken, when estimating the amount of solar-modules needed to supply the stations. As a rule of thumb, 10% of the nominal maximum voltage will be supplied by the panels on average (i.e., using a 50 Watt module just 5 W are available on average). Voltage will typically decrease when the panels are installed high up in the mountains. Clouds, snow or ashfall may further reduce the effective power output (i.e., 5% or even less). This significantly increases the number of PV-modules required. To give an example, let the station consume at least 20 W (including radio, some digitizers, SCSI disks for local storage etc.), you will need 400 W panel power in the worst case which is 8 x 50 W panels at this station!

New developments in manufacturing solar-cells increase the efficiency of PV-modules. Especially the CIS technology (Copper, Indium and Selen) in combination with different alignment of the cells makes them more robust against partial shadowing and more efficient in diffusive light condition. The higher costs per panel are balanced by fewer numbers of panels needed to supply power for the monitoring station.

Also the capacity of the battery must be adequate in case no solar power is produced. However, the battery should not be too big in capacity as the PV-modules must be able to recharge the battery in sufficient time. Whenever possible, alternative power sources should be used, such as robust wind generators. It should be kept in mind, however, that a nearby wind generator might increase the noise of a seismic sensor. In case the station is located near running water, a small hydro-power engine could be a good alternative.

A novel and promising approach in building independent power supply which also produce enough power to feed even sophisticated but high power consuming radar devices or satellite dishes originates from the development of fuel cells. Meanwhile there are several different systems of portable fuel cells available. The fuels ranges from classical hydrogen to methanol based systems, which are cheaper and also carry a higher energy to weight factor. Combining PV modules together with fuel cells reduces the usage of fuel by guaranteeing the unbroken power supply. We note however that methanol is a hazardous substance and sometimes difficult to transport on ships and planes.

Care should be taken in the choice of the (programmable) solar charger and the batteries used. Currently Absorbed Glass Mat (AGM) batteries are most suitable for application at a remote site. They are lighter and more easily charged and discharged than common lead-gel batteries but also completely sealed and thus maintenance free. In future lithium air batteries might play an important role in solving the current problems of running monitoring equipment at remote areas.

### 13.6.5 Data center

All data streams, including those from monitoring techniques in addition to seismic, should be collected, stored and archived in a central facility. In the age of high-performance low-cost PC's, few standard computers will be sufficient to satisfy all needs of data collection, backup systems, automatic and visual analysis. Because continuous recording of all relevant signals is preferred over triggered data, a good backup strategy is crucial for getting complete and long-term data. Continuous recording is indispensable for improving our knowledge about volcanic activity, the underlying physical mechanisms and the relevant parameters to be observed when aiming at improving eruption forecasting. With the advent of DVD disks and CD-ROMS a good solution would be to write images of data sets onto one of these media in a daily or regular manner. CD-ROMS in particular assure a good data safety to price ratio.

Much public domain software is available for either automatic (e.g., EARTHWORM; Johnson et al., 1995) or interactive analysis (SeismicHandler, SAC, IASPEI-Software, PITSA/GIANT). The Orfeus homepage is a good starting point when looking for suitable software and for further contacts ([www.orfeus-eu.org](http://www.orfeus-eu.org))

A very promising new application in the field of “real-time” analysis is distributed as seiscomp3 by the German Research Centre for Geosciences (GFZ-Potsdam). The main advantages are the new design concepts (C++; Python plugins etc) as well as an up-to-date graphical user interface for interactive analysis. In contrast to EARTHWORM, seiscomp3 is directly connected to a SQL-database which can store the analyzed (seismic) event data. The major drawback of this new package lies in the strong focus on seismic real time analysis, which is difficult to change.

A different strategy was chosen during the development of ObsPy ([www.obspy.org](http://www.obspy.org)) and SeisHub ([www.seishub.org](http://www.seishub.org)), which are not real-time packages but form a “near-real-time” system adapted for the need of small to medium sized observatories. All source code parts are completely licensed under GNU and may be altered and changed according to the need of the users.

Obspy/SeisHub together with EARTHWORM or seiscomp3 may serve as a good, flexible and easy extendable software solution for a volcano data center.

A special requirement of the data center is the availability of an “uninterruptable power supply” (UPS) to guarantee a loss-less data collection even if the power line of the observatory is broken. Depending on the quality of the power network, a generator could be a good solution to bypass blackouts which may last several hours.

Establishing a “quick-response” volcano observatory during a volcanic crisis of a long-dormant volcano need additional equipment and design criteria of the monitoring network to be deployed. All equipment, including the data center facilities, should be light weighted, robustly made and should have low power consumption. These demands very likely lead to down-grades in resolution and data through-put. A comprehensive description of one realization of mobile monitoring network, the *Mobile Volcano-Monitoring System*, is given by Murray et al. (1996).

Meanwhile several different new applications exist which are designed to react as quickly as possible to new volcanic crisis. An early European initiative EMEWS (European Mobile Early Warning System) was formed in order to define European standards in mobile volcano alarm systems.

A more national approach, which tries to include all possible sets of monitoring parameters, environmental parameters and network metadata is the Exupery-Volcano Fast Response System (VFRS; [www.exupery-vfrs.de](http://www.exupery-vfrs.de)). The goal of this initiative is to record and store all volcano relevant information in a newly developed database ([www.seishub.org](http://www.seishub.org)), to find new strategies of volcanic monitoring (e.g., ground base InSAR; space borne surveillance of active volcanoes) and to aid or guide the local observers and/or decision maker by creating a data visualization system based on geographic information (GIS) and the estimated monitoring parameters (via EARTHWORM/seiscomp3 – ObsPy/SeisHub). The resulting GIS is also freely available (executable as well as source code) at [www.exupery-vfrs.de](http://www.exupery-vfrs.de). Fig. 13.45 gives a snapshot of possible applications. One of the design criteria for the GIS was its platform independence and easy handling.

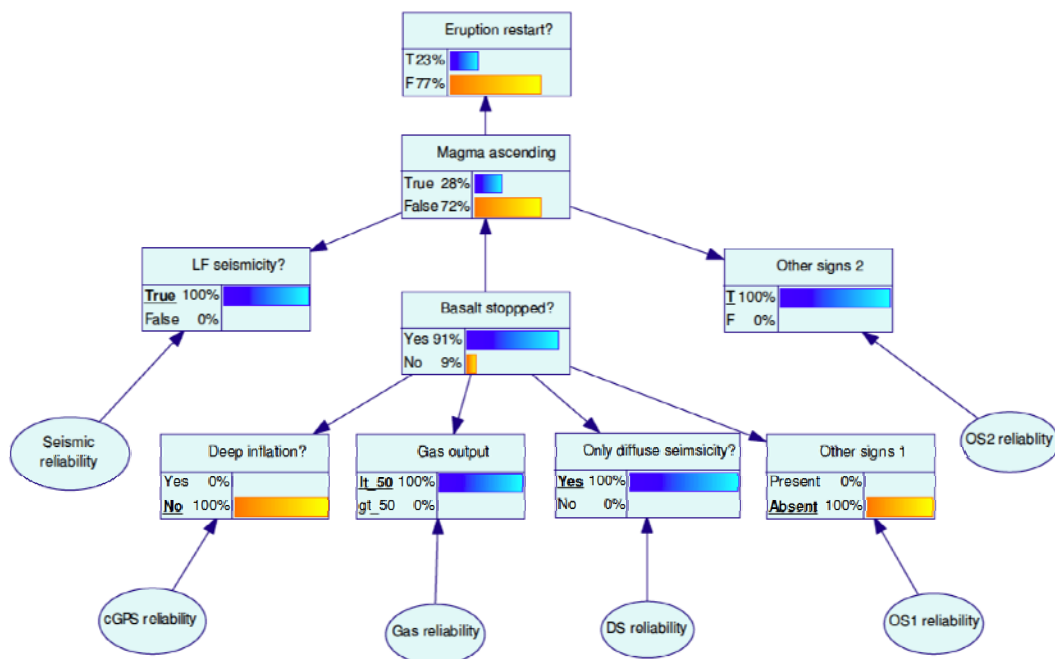


**Fig. 13.45** Snapshot of the Exupery GIS. The browser-based application can include ground-based parameter sets and data as well as display 2D InSAR sections and 3D models.

The more and more common storage of multi-parameter data together with environmental data makes the application of “data mining” much more attractive. First steps in this direction are the application of automatic alert level systems based on Bayesian Belief Networks (BBN), which were introduced in volcanology by Aspinall et al. (2003) and Marzocchi et al. (2008). Fig. 13.46 shows the graphical network representation of a BBN build in into the Exupery VFRS. The output of this network consists of a simple alert level estimation which



should only guide the local authorities in their decision. One of the key advantages of this probabilistic belief networks lies in the specification of uncertainties of parameters which lead to a decision. With the interactive tools now available (e.g., GeNie, <http://genie.sis.pitt.edu>) and open source library (smile, <http://genie.sis.pitt.edu>) an observer is able to compute the level of probability once a BBN has been constructed and interactively change the BBN output by putting weights on different nodes, when additional information (e.g., field observation, “expert knowledge”) is at hand.



**Fig. 13.46** Bayesian Belief Network used for automatic alert level estimation within the Exupery VFRS. This BBN is a modified version of a graphical network designed by Aspinall (courtesy of the author, from: [www.exupery-vfrs.de](http://www.exupery-vfrs.de)).

## Acknowledgments

I am grateful to the reviewers F. Klein, R. Scarpa and R. Tilling for their helpful comments and corrections to the first version, M. Hort and C. Bean for the revised manuscript, which improved the text significantly. I also thank P. Bormann for additional editing of this chapter. Without his continuous effort and labor this chapter never would have been realized. Furthermore leaving the figures and seismograms of E. Gottschämmer, S. Falsaperla, S. McNutt and M. Ohrnberger are highly appreciated. A lot material was taken from the Exupery project homepage ([www.exupery-vfrs.de](http://www.exupery-vfrs.de)).

## Recommended overview readings

Civetta, L., Gasparini, P., Luongo, G., and Rapolla, A. (Eds.), 1974. *Physical volcanology, developments in solid earth geophysics*, 6, Elsevier, Amsterdam.

Kawakatsu, H. and Yamamoto, M., 2007. Volcano Seismology, in *Treatise on Geophysics Vol.4*, pp. 389–420, ed. Schubert G., Elsevier, Oxford.

# Chapter 14

## Investigation of Site Response in Urban Areas by Using Earthquake Data and Seismic Noise

(Version January 2012; DOI: 10.2312/GFZ.NMSOP-2\_ch14)

Stefano Parolai

Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Department 2: Physics of the Earth,  
Section 2.1: Earthquake Risk and Early Warning, Telegrafenberg, 14473 Potsdam, Germany,  
E-mail: [parolai@gfz-potsdam.de](mailto:parolai@gfz-potsdam.de)

|                                                                                         | page |
|-----------------------------------------------------------------------------------------|------|
| <b>14.1 Introduction</b>                                                                | 1    |
| <b>14.2 Empirical methods (earthquake based)</b>                                        | 5    |
| 14.2.1 Reference site technique                                                         | 5    |
| 14.2.2.1 Standard Spectral Ratio method (SSR)                                           | 6    |
| 14.2.2.2 Generalized Inversion Technique (GIT)                                          | 6    |
| 14.2.2 Non-reference site techniques: H/V spectral ratio of earthquake data (EHV)       | 7    |
| 14.2.3 Vertical array data analysis                                                     | 8    |
| <b>14.3 Seismic noise measurements</b>                                                  | 10   |
| 14.3.1 Seismic Noise Horizontal-to-Vertical spectral ratio (NHV)                        | 10   |
| 14.3.1.1 Relationship between NHV peak frequency and thickness of the sedimentary cover | 12   |
| 14.3.1.2 NHV inversion                                                                  | 13   |
| <b>14.4 2-D array</b>                                                                   | 17   |
| 14.4.1 f-k based methods                                                                | 17   |
| 14.4.2 Spatial and Extended Spatial Autocorrelation method (SPAC and ESPAC)             | 20   |
| <b>14.5 Interferometry and tomography</b>                                               | 25   |
| <b>14.6 Conclusions, suggestions and related studies</b>                                | 26   |
| <b>Acknowledgment</b>                                                                   | 30   |
| <b>References</b>                                                                       | 30   |

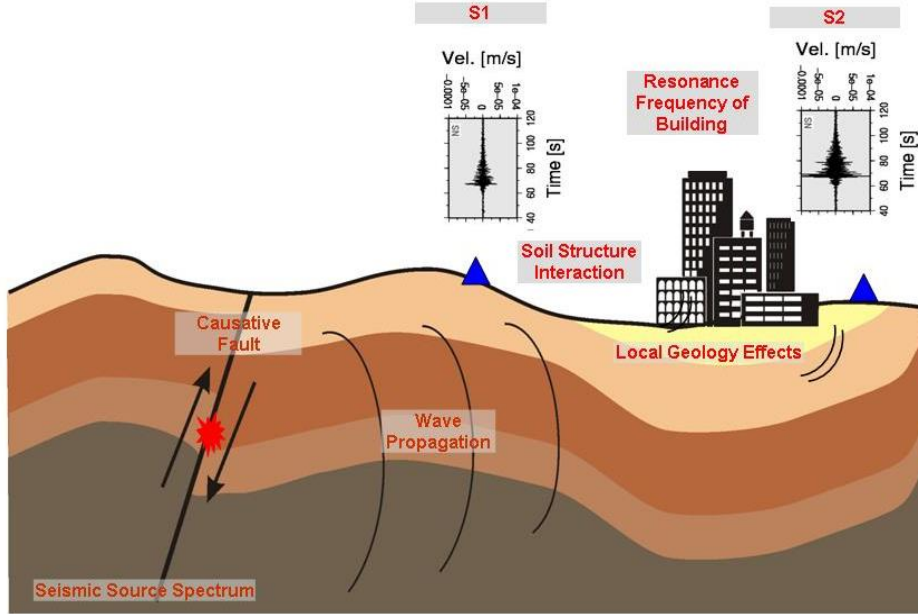
*“It is certainly not enough to have softwares available to all so that all are musicians.”*  
*Angelo Bergamini*

### 14.1 Introduction

The distribution of damages due to recent earthquakes have shown that the effects of local geology on ground shaking represent an important factor in engineering seismology. Figure 14.1, schematically shows that when an earthquake occurs, waves generated at the source propagate through the Earth (generally with decreasing amplitude with distance). When two



nearby sites are located on the top of different materials, the ground shaking they experience can be very different. These differences might be very important considering that (1) buildings can react to the input ground motion emphasising shaking depending on their frequency of oscillation, and (2) that they can even redistribute energy back to the Earth due to scattering and diffraction of seismic waves when interacting with their foundations and acting as a source while shaking.



**Fig. 14.1** Simplified representation showing seismic wave propagation after an earthquake. Blue triangles indicate stations located over different near surface geological material (S1 on a rock site and S2 on soft quaternary material) experiencing different ground motion.

The modification of the ground motion due to the Earth structure below the site can be related to different factors (e.g., Safak, 2001) (Fig. 14.2). In general, the main factor is the impedance contrast (see Chapter 2, subsection 2.5.4.1) between the soft sedimentary cover (with low density  $\rho_1$  and wave propagation velocity  $v_1$ ) and the bedrock (with high density  $\rho_2$  and wave propagation velocity  $v_2$ ). The impedance contrast  $c=(v_2 \rho_2)/(v_1 \rho_1)$  determines how strongly the waves at particular frequencies (the fundamental resonance frequency and the higher harmonics) are multi-reflected within the soft layer, thus trapping the wave energy within the layer (1-D effect). Quantitatively, and under the simplified assumption of vertical propagation of S waves, the modulus of the 1-D site amplification function  $H(f)$  can be described by the following relation:

$$|H(f)| = \left( \frac{(1+r)^2}{1+2r \cos(4\pi f \tau) + r^2} \right)^{\frac{1}{2}}, \quad (14.1)$$

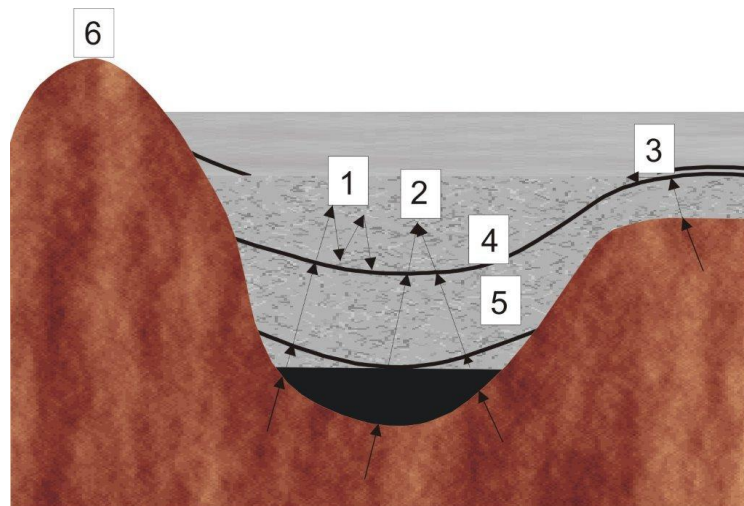
where  $\tau$  is the travel time of the S waves in the soft sedimentary layer and  $r=(\rho_2 v_2 - \rho_1 v_1)/(\rho_2 v_2 + \rho_1 v_1)$  is the reflection coefficient that is clearly related to the impedance contrast  $c$ . The maximum of  $H(f)$  is obtained when  $\cos(4\pi f \tau)$  is equal to -1 that is when  $f = 1/(4\tau)$ . This value of  $f$  is defined as the fundamental resonance frequency  $f_0$ . From the previous

consideration it follows the well know relationship  $f_0 = v_1/(4h_1)$  where  $h_1$  is the thickness of the soft sedimentary layer.

The site response (here represented by the shape of the Fourier spectra amplitude ratio between the surface recording and the input ground motion at the bedrock) is then determined by the S-wave velocity and the thickness of the sedimentary cover layer, which determine the frequencies that are affected by amplification, while the impedance contrast controls the amplification level. The quality factor  $Q$  (see Chapter 2, subsection 2.5.4.2) in the soft sedimentary cover influences the decay of the site response with increasing frequency.

Other effects like focusing, diffraction of body waves at the edges of sedimentary basins (2-D or 3-D effects) and scattering should also be taken into account. Sites located on pronounced (with respect to wavelength) topographic relief can also suffer significant amplification of ground motion. Importantly, in the case of strong ground motion the soil may behave non-linearly. In particular, the frequency of resonance of soil tends then to decrease and the damping to increase. Due to pore pressure increase and dilatancy high frequency spikes in ground motion are occasionally observed (Bonilla et al., 2005).

Thus, assessing the amplification/deamplification and lengthening of seismic ground motion due to superficial geology is of primary importance for earthquake microzonation studies. In the past few decades, the evaluation of site effects has attracted growing attention. In particular, due to the increasing importance of reliable seismic hazard assessment in large urban areas (e.g., Istanbul, Santiago de Chile), large efforts has been devoted to the development of tools for site effect estimation that allow large areas to be covered at limited costs.



**Fig. 14.2** Local features and effects that can modify the ground motion: 1) Resonance due to impedance contrasts; 2) Focusing due to subsurface topography; 3) Body waves converted to surface waves; 4) Water content; 5) Heterogeneity of the Earth medium; 6) Surface topography (Safak, 2001).

Site effects can be estimated by means of numerical simulations and experimental methods. The former range from simple 1-D numerical calculations to methods that consider complex phenomena and model geometry. In general, the experimental methods are subdivided into two categories: reference site and non-reference site techniques (see, e.g., Bard, 1995).

Borcherdt (1970) was the first to propose the reference site method (RSM) that consists in comparing records at nearby sites, using one as the reference site. It is assumed that records from the reference site (in general a station installed on outcropping hard rock) contain the same source and propagation effects as records from the other sites. Therefore, differences observed between the sites are explained as being due to local site effects. This method is generally applicable only to data from dense, local station deployments of arrays (see also Chapter 9 of this Manual). Andrews (1986) extended this method to large networks. He proposed to separate the contribution of the path, the source, and the site by means of an inversion scheme termed the generalized inversion technique (GIT). Since in general the site amplification at a reference (hard rock) site is set to unity, this approach provides results analogous to those obtained by the RSM method.

However, the major drawback of these methods is that a suitable reference site may not always be available (Steidl et al., 1996), because also rock sites might have their own response. Furthermore, being rock site often located on hills, topographic effects (due to the geometry of the Earth surface, the weathering of the shallow rock layers etc.) cannot be easily ruled out. In order to overcome this disadvantage, non-reference site techniques such as the horizontal-to-vertical (H/V) spectral ratio method are widely used, although they provide only partial information about the site response. Nogoshi and Igarashi (1970) first introduced this technique, but Nakamura (1989) developed it further as a tool for estimating the site response of S waves. Originally, it was proposed for interpreting ambient seismic noise measurements (NHV). Lermo and Chavez-Garcia (1993) applied the H/V technique to the S waves of earthquake recordings (EHV) and developed a theoretical background for the technique using numerical modelling of SV waves.

The comparison of site responses obtained with reference and non-reference site techniques (e.g. Field and Jacob, 1995; Bonilla et al., 1997; Riepl et al., 1998; Triantafyllidis et al., 1999; Parolai et al., 2000; 2001a; 2004) has shown that RSM and EHV usually provide site responses with similar shapes when the S-wave part of the seismogram is used. The fundamental resonance frequency of a site is consistently estimated by both methods. However, they can provide different levels of amplification, in particular at frequencies higher than the fundamental one. Parolai and Richwalski (2004) showed, by means of numerical simulations, that differences between the RSM and the EHV results are caused by the effect of waves that have been converted at the bedrock depth and are included in the analyzed record time windows. In particular, due to the S-to-P conversion, a deamplification at frequencies higher than the fundamental one is observed in the EHV ratio for S-waves.

NHV has been shown to provide (e.g., Haghshenas et al., 2008) a good estimate of the fundamental resonance frequency of the S waves, but it underestimates the level of amplification with respect to GIT and RSM. Also, by using the P-wave part of the seismograms, the RSM and EHV methods have been found to provide consistent results only in some cases.

In the following, the development, testing and application of techniques for site effect estimation is outlined with special attention to studies carried out in large urban areas. First, the most popular methods for site effects investigation in urban area are described. Second, the use of seismic noise as an optimal tool for investigating the subsoil structure in urban areas will be illustrated. Not described are here other subsoil investigation methods based on active sources such as SASW (Stokoe et al., 1994), MASW (Park et al., 1999) and REMI (Louie 2001). Finally, based on the experience gained in the application of empirical

techniques to large urban areas, recommendations are given on how to estimate best site effects in different environments.

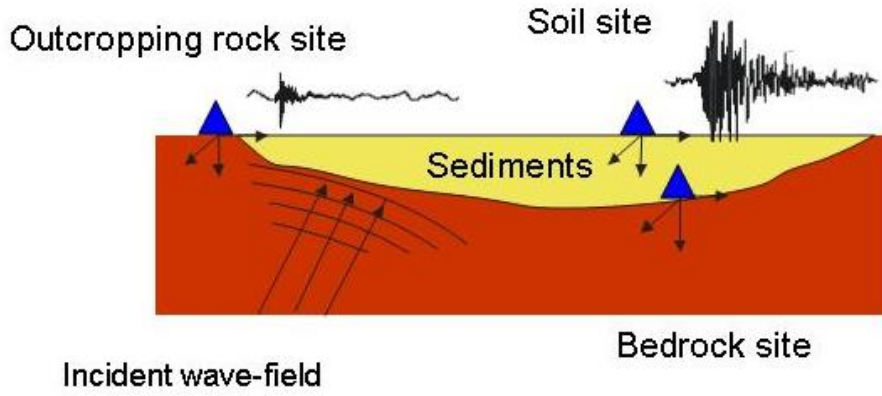
The intention of this chapter is to provide an overview of the work carried out during past and ongoing activities, with an emphasis on personal practical experience. Thus, this chapter may not appear to some readers to be a comprehensive review of the huge amount of significant literature that has been published in this field. If so, the author apologises for this.

## **14.2 Empirical methods (earthquake based)**

The empirical methods can be classified into two categories: reference site and non reference site techniques. While the former requires the availability of a station installed on a hard rock outcrop (the reference site), the latter assumes that one of the components of the ground motion (generally the vertical one) provides the input ground motion at the bedrock (see for more details 14.2.2). This assumption is mainly based on the empirical observation of similar amplitude of the horizontal and vertical component of ground motion in borehole recordings and on the presumed lack of amplification of the latter one.

### **14.2.1 Reference site techniques**

Reference site techniques can be used by taking the direct spectral ratio of the same component of ground motion recorded at two nearby stations (Standard Spectral Ratio SSR) (Fig. 14.3) or by deriving the site response via an inversion procedure (generalized inversion technique GIT). Generally, S-wave or P-wave signal windows are extracted from the seismograms starting before the phase arrival and ending either when a certain amount of energy is reached or after a pre-fixed amount of seconds. The selected time series are corrected for the instrumental response and the trend in the data. The two horizontal components can be analysed separately or combined. The vectorial sum or the root means square of the two horizontal components is often used in the analysis. Both ends of the selected signal window are tapered (for example with 5% cosine functions) and then the Fast Fourier Transform (FFT) calculated. Tapering is necessary to suppress spectral leakage before calculating the FFT (see IS 14.1). The resulting spectra are smoothed by using running time windows (preferably equally spaced on a logarithmic scale). The analyses are performed considering frequencies with a signal-to-noise (SNR) ratio larger than a certain threshold (generally at least 3), estimated by considering the Fast Fourier Transform (FFT) of a noise window as wide as the signal window.



**Fig. 14.3** Schematic description of the Reference Site Method. Blue triangles represent hypothetical seismological stations recording three component ground motion (directions are schematically indicated by the arrows).

#### 14.2.1.1 Standard Spectral Ratio method (SSR)

The Standard Spectral Ratio technique (SSR) consists of comparing records at nearby sites, using one as the reference site (Figure 14. 3). It is assumed that records from the reference site (in general a station installed on outcropping hard bedrock) contain the same source and propagation effects as records from the other sites that is:

$$\frac{U_{ij}(f, r)}{U_{ik}(f, r)} = \frac{S_i(f)Z(f)_j A_{ij}(f, r)}{S_i(f)Z(f)_k A_{ik}(f, r)} \quad (14.2)$$

where,  $U_{ij}(f, r)$  is the spectral amplitude of the ground motion observed at a recording site  $j$  for an event  $i$ ,  $S_i(f)$  is the source function,  $Z_j(f)$  is the site response,  $A_{ij}(f, r)$  is a function accounting for attenuation that includes the effect of geometrical spreading, intrinsic and scattering quality factor.  $f$  is the frequency,  $r$  the hypocentral distance and  $k$  indicates the reference station. Note, that similarly to what is schematically indicated in Figure 14.1, the ground motion recorded at one station is considered as the convolution of three terms, specifically the source  $S_i(f)$ , the path  $A_{ij}(f, r)$  and site response  $Z_j(f)$ . If the stations are nearby and the reference station is not affected by site effects, then  $Z_k = 1$  and  $A_{ij}(f, r) \approx A_{ik}(f, r)$ . Thus, the spectral ratio directly provides the site response at the non-reference stations. The major drawback of this method is that rock stations may also have their own response. Furthermore, a good reference site might be located too faraway from the target site therefore not allowing assuming similarity in the wave paths towards the two stations.

#### 14.2.1.2 Generalized inversion technique (GIT)

The natural logarithm of the spectral amplitude  $U_{ij}(f, r)$ , observed at a recording site  $j$  for an event  $i$ , can be represented by:

$$\ln U_{ij}(f, r) = \ln S_i(f) + \ln Z(f)_j + \ln A_{ij}(f, r), \quad (14.3)$$

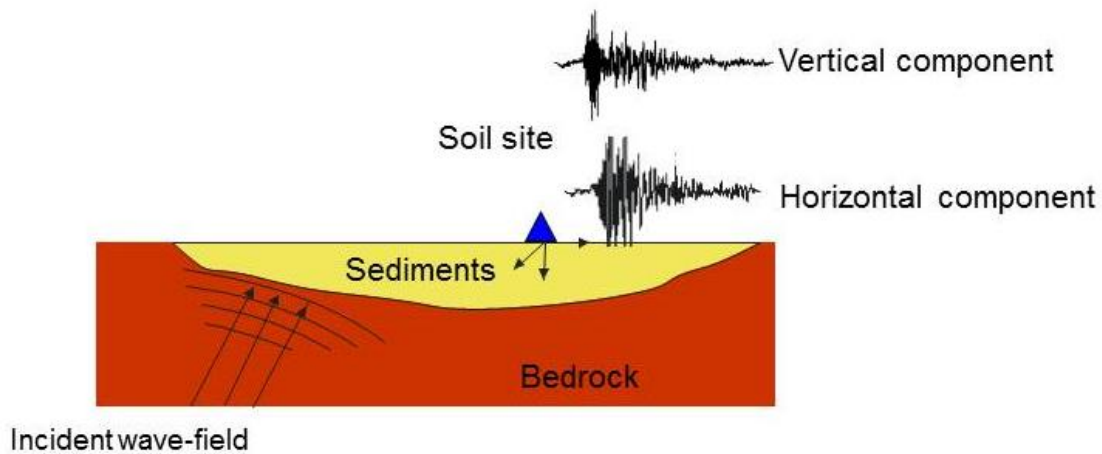
where, according to equation (14.2),  $S_i(f)$  is the source function,  $Z_j(f)$  is the site response,  $A_{ij}(f, r)$  is a function accounting for attenuation,  $f$  is the frequency and  $r$  the hypocentral distance.

For all events and stations, equation (14.3) describes a linear system of the form  $\mathbf{Ax} = \mathbf{b}$ , which can be solved by using suitable inversion algorithms. The inversion can even be performed in two steps following the procedure of Castro et al. (1990). In the first step, the inversion is carried out to separate the source and attenuation effects without introducing any parametric model to describe the attenuation term  $A_{ij}(f,r)$ . The trade-off between attenuation and source is resolved by requiring that attenuation assumes an a priori chosen value at a given distance, irrespective of frequency. This assumption is justified by the small effect that the total attenuation (intrinsic and scattering) has on the decay of the wavefield amplitude at short distances. In the second step the inversion is carried out for the source and site functions, using the observed data that have been corrected for the attenuation. The undetermined degree of freedom of the second inversion is resolved by constraining the logarithm of the site amplification of the reference station (generally a rock site) to 0, irrespective of frequency. In cases when a reference site cannot be identified (for example by a simple horizontal-to-vertical spectral ratio of earthquake data; see next section 14.2.2) the average response of the whole network is constrained to 0. Although this restriction affects the absolute value of the single site functions, it does not affect the source functions.

Parolai et al., (2000) applied the GIT technique to investigate the site response at stations of a small urban network, while Bindi et al., (2009) used the GIT for estimating 2-D or 3-D site effects in a small basin. Recent examples of application of the GIT approach include Moya and Irikura (2003), who performed the inversion using a reference event instead of a reference site, Oth et al. (2008 and 2009), who applied the technique to an area where a simple parameterization of attenuation was not possible, and Drouet et al., (2008), who introduced a non-linear inversion scheme, although this is valid only under assumptions made on the source spectral shape.

### **14.2.2 Non-reference site techniques: H/V spectral ratio of earthquake data (EHV)**

Non-reference site techniques allow to avoid the requirement of having a good hard rock reference station, either by making the assumption that the vertical component of the ground motion is not affected by amplification and that it is similar to the horizontal component at the bedrock (when earthquake recordings are used), or using seismic noise recordings generally dominated by surface waves (Fig. 14.4). While the assumption of the former cannot be true for frequencies higher than the fundamental one (Parolai and Richwalski, 2004), the second suffers from the lack of correlation between the estimated amplification level and that measured by earthquake data analysis. However, both methods can satisfactorily estimate the fundamental resonance frequency of a site.



**Fig. 14.4** Schematic description of the EHV method. The blue triangle represents a hypothetical seismological station recording three-component ground motion (directions are schematically indicated by the arrows). One horizontal component and the vertical component recordings are shown.

The EHV spectral ratio technique was introduced by Langston (1979) as a method to determine crustal and upper-mantle structure from teleseismic recordings. The basic assumption of the method is that the vertical component of motion is not influenced by the local structure, whereas the horizontal components contain the P-to-S conversions due to the geology underlying the station. Therefore, by deconvolving the vertical component from the horizontal, one could estimate the site response. The deconvolution is generally carried out by dividing the Fourier spectra of the horizontal component by the vertical one. However, the deconvolution operation is applied to data corrupted by noise and therefore, since this problem is ill-conditioned, small errors in the data could lead to solutions unacceptable from a physical point of view. Several regularization techniques are available to avoid artifacts in the results. The most common practice, although it is not properly a regularization approach, is to carry out the spectral ratio after having smoothed the spectra. The technique has been applied in many studies (e.g., Lermo and Chavez-Garcia, 1993; Field and Jacob, 1995; Lachet et al., 1996; Bonilla et al., 1997; Parolai et al., 2004; Parolai and Richwalski, 2004). In general, it was found that H/V ratios for S waves reveal the overall frequency dependence of the site response, even if the level of amplification, especially at frequencies higher than the fundamental one, can be different from that obtained by SSR analysis.

### 14.2.3 Vertical array data analysis

The increasing number of observations over the last few decades has greatly contributed towards the understanding of strong-motion site effects by the engineering and seismological communities, leading to advancements in the state of knowledge and modelling of in situ sediment response (Field et al., 1997). Nonetheless, finer resolution and more effective representation of the involved physics in the near surface medium require a good estimate of the input motion and of wave propagation in shallow layers. For this, by far, the best source of information is provided by downhole arrays, consisting of accelerometers recording ground motion at different depths and on the surface of the same site (see Chapter 7, Sub-chapter 7.4.6 on borehole strong-motion array installations).

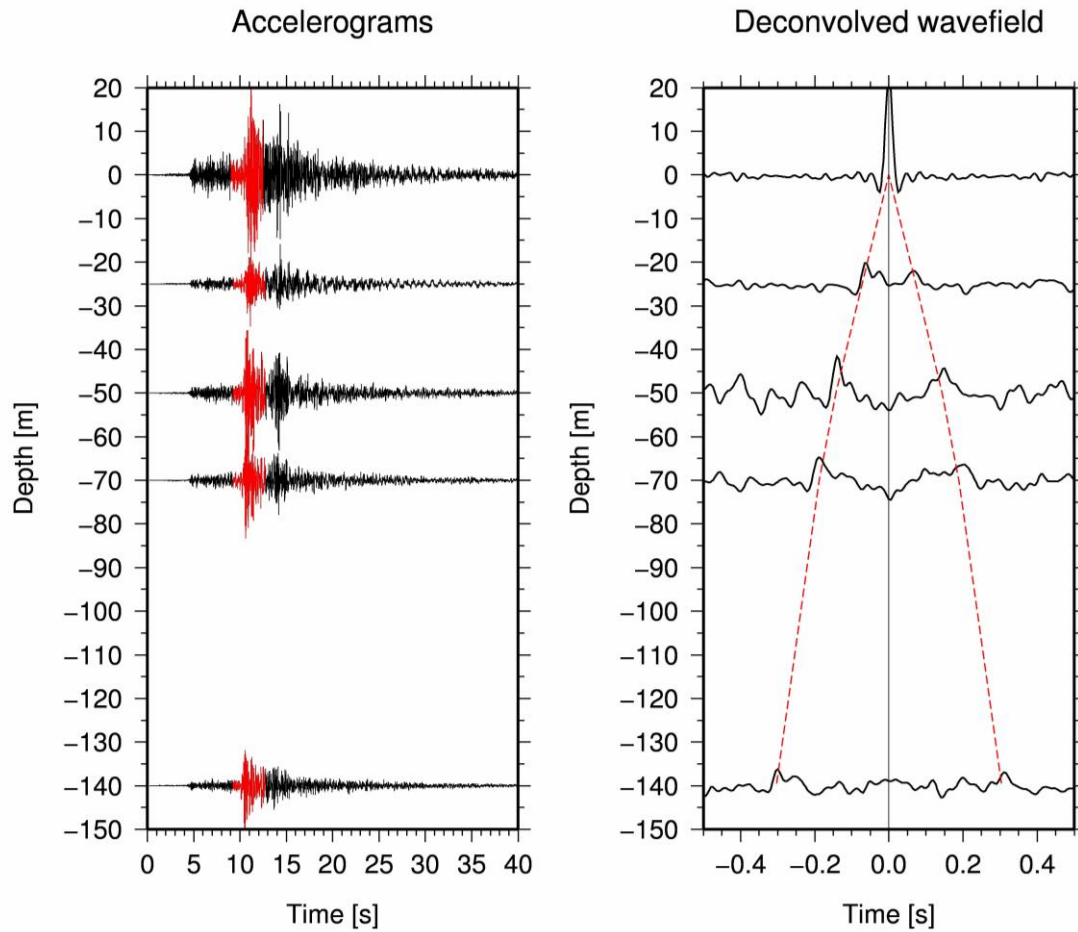


Downhole measurements are a valuable complement to in situ and laboratory geotechnical investigation techniques. Indeed, they provide critical constraints for both the interpretation methods for surface observations as well as information on the real material behaviour and overall site response over a wide range of loading conditions (Assimaki et al., 2008). The amount and quality of information from downhole arrays in seismically active areas is the key for improving our understanding of in situ soil behaviour, assessing the modelling and parametric uncertainties associated with employed methodologies for strong-motion site-response analysis, and for shallow geological investigations.

Several methods have been proposed for estimating wave propagation velocity and damping from boreholes recordings. Amongst others, Trampert et al., (1993), van Vossen et al., (2004) and van Vossen et al., (2005), focused their attention on the SH phase propagation, while Mehta et al. (2007a; 2007b), starting from the similarity between deconvolution and the cross correlation tool used in seismic interferometry (e.g., Lobkins and Weaver 2001; Schuster et al., 2004; Shapiro and Campillo 2004; Snieder et al., 2006; Halliday and Curtis, 2008a), showed that the deconvolution of waveforms recorded by a vertical array can provide useful insight into the wavefield propagation in the uppermost crustal layers,. Although the use of the Fourier spectra amplitude ratio (uphole to downhole or with a reference station) is well known in engineering seismology (e.g., Safak, 1997), the deconvolution of seismograms (retaining information about phase) and their interpretation in the time domain has found few applications (Trampert et al., 1993; van Vossen et al., 2004; van Vossen et al., 2005). Assimaki et al.,(2008) proposed an inversion procedure that aims at estimating the best borehole model in term of shear wave velocity, attenuation and density, by optimizing the correlation between observed and synthetic seismograms. Parolai et al. (2009) showed by using a deconvolution approach that the wave propagation in the shallowest layer can be imaged independently on the chosen signal window and that down-going waves resulting from the reflection at the free surface might affect the recordings down to more than 100 m depth (see also Fig. 4.33 in Chapter 4). Yamada et al. (2010) estimated non-linear effects during strong shaking applying a deconvolution approach. Mogi et al., (2010) studied non-linear soil behaviour using a normalised input-output normalisation method. Parolai et al., (2010) showed that a robust estimate of an average  $Q_s$ , that can be used for numerical simulations of ground motion, can be obtained by means of deconvolution of array recordings. Figure 14.5 shows an example of borehole recordings and of the relevant deconvolved wavefield.



It is worth noting how the ground motion is modified (in general amplified) with decreasing depth of recording, being largest at the surface. For this surface amplification effect see also Fig. 4.22 in Chapter 4 and Table 1 in EX 3.4. The deconvolved wavefield, obtained by analysing the signal window indicated in red in the left panel, clearly shows a peak in the a-causal (negative times) part of the signal propagating toward the surface with a velocity consistent with the S-wave velocity in the borehole (red dashed line). A peak corresponding to a downgoing wave is observed also at all the depths analyzed in the causal (positive times) part of the signal. But the downgoing wave peak has always smaller amplitude than the upgoing one due to the effect of attenuation.



**Fig. 14.5 Left-hand panel:** horizontal component of ground acceleration recorded at different depths. Red record traces indicate the window containing S waves. **Right hand panel:** the upgoing (left) and downgoing (right) S waves in the deconvolved wavefield. The S-wave traveltimes, computed from the velocity model derived with the Parolai et al. (2005) inversion scheme, are indicated by the broken red lines.

## 14.3 Seismic noise measurements

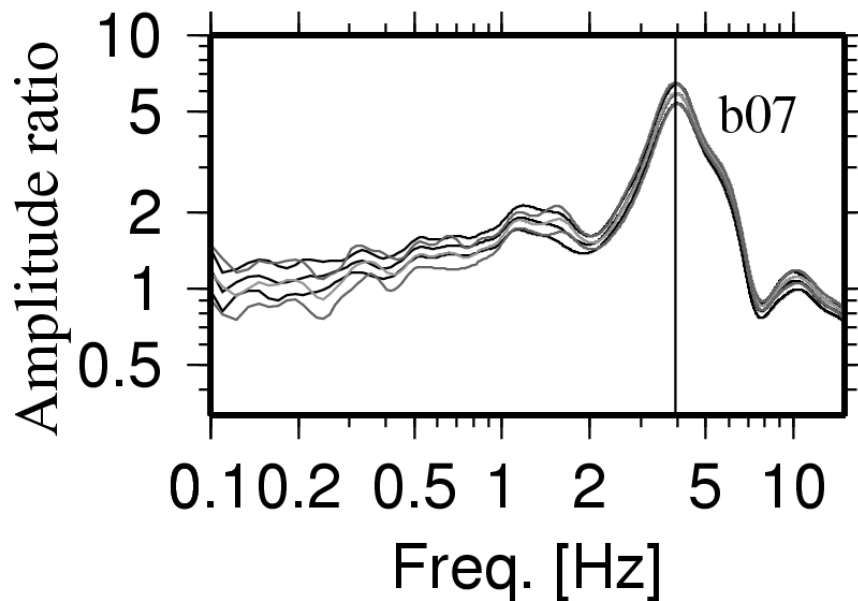
### 14.3.1 Seismic Noise Horizontal-to-Vertical spectral ratio (NHV)

Nakamura (1989) revised the Horizontal-to-Vertical (H/V) seismic noise spectral ratio technique (NHV), first proposed by Nogoshi and Igarashi (1970 and 1971). Since then, in the field of site effect estimation, a large number of studies have been published, which use this

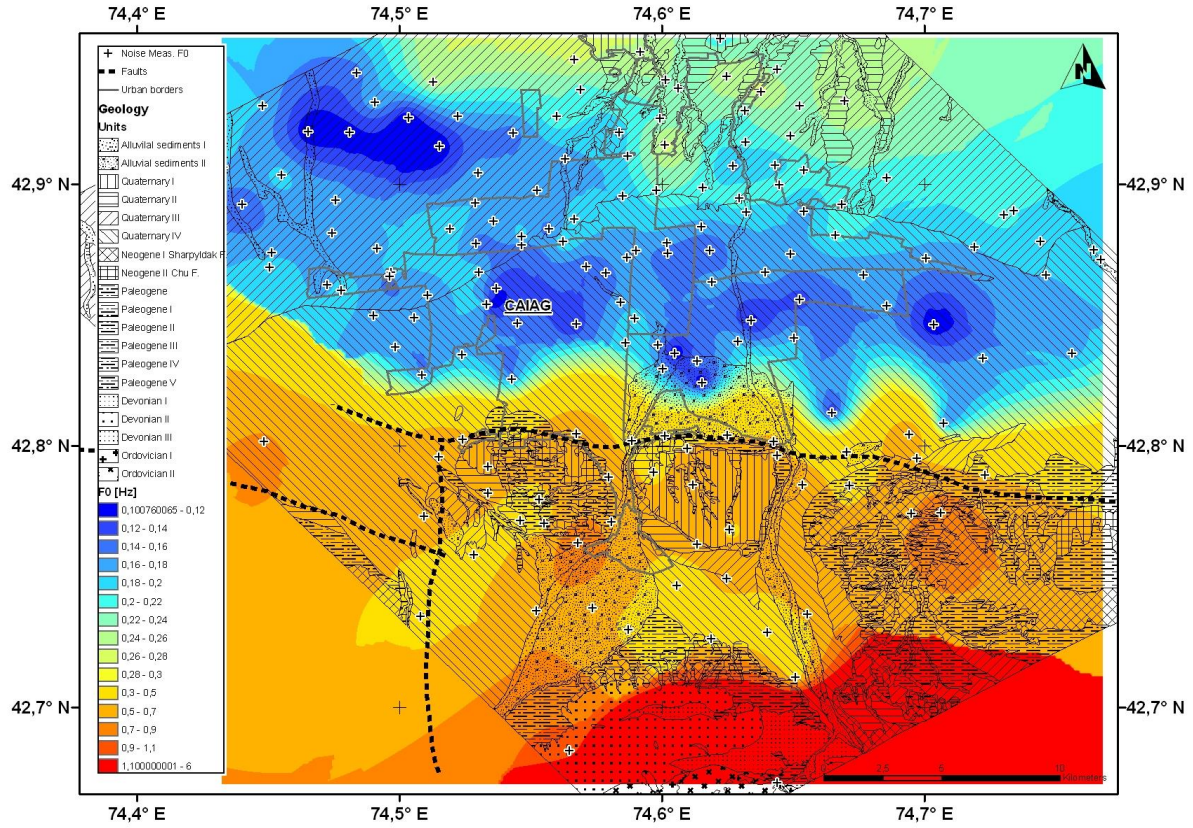
low cost, fast and therefore attractive technique (e.g. Field and Jacob, 1993; Lermo and Chavez-Garcia, 1994; Mucciarelli, 1998; Bard, 1999; Parolai et al., 2001b).

While in the Nakamura interpretation the NHV spectral ratio is directly related to the S-wave site response, most of the seismological/engineering community, based on the results of recent studies, is more inclined to consider predominant the role of surface waves (see e.g. Bard, 1999).

Most of the researchers focused their attention on the comparison of noise H/V spectral ratio and earthquake site response and agreed that the H/V spectral ratio of seismic noise provides a fair estimate of the fundamental resonance frequency of a site (Fig. 14.6). However, attempts to provide standards for the analysis of seismic noise have only recently been carried out (Bard, 1999; SESAME, 2003; Picozzi et al., 2005a). When using the NHV technique particular attention must be paid to the used seismic instruments and sensors (Guiller et al., 2007; Strollo et al., 2008a and b). Resonance frequency maps for large urban areas were recently provided by Parolai et al. (2001b) (Cologne, Germany), Duval et al. (2001) (Caracas, Venezuela), Picozzi et al. (2008) (Istanbul, Turkey), Bonnefoy-Claudet et al. (2009), Pilz et al. (2009 and 2010) (Santiago de Chile, Chile), Parolai et al. (2010b) (Bishkek, Kyrgyzstan, Figure 14.7), after validation with earthquake data.



**Fig. 14.6** Example of NHV ratio. The peak allows an estimate of the fundamental resonance frequency of the site (redrawn from Parolai and Galliana-Merino, 2006).



**Fig. 14.7** Fundamental resonance frequency map of Bishkek (Kyrgyzstan). The crosses indicate sites where single station noise measurements were carried out. The map was obtained after interpolation between these points. A geological legend and sketch of the area is also shown.

#### 14.3.1.1 Relationship between NHV peak frequency and thickness of the sedimentary cover

Recent studies (Yamanaka et al., 1994; Ibs-von Seht and Wohlenberg, 1999; Delgado et al., 2000 a and b; Parolai et al., 2002; D'Amico et al., 2008) showed that noise measurements can be used to map the thickness of soft sediments. Quantitative relationships between this thickness and the fundamental resonance frequency of the sedimentary cover as determined from the peak in the NHV spectral ratio were calculated for different basins in Europe (e.g. Ibs-von Seht and Wohlenberg, 1999; Delgado et al., 2000 a).

The approach is based on the assumption that in the investigated area, lateral variations of the S-wave velocity are minor and that it mainly increases with depth following a relation like

$$v_s(z) = v_{s0}(1 + Z)^x \quad (14.4)$$

where  $v_{s0}$  is the surface shear-wave velocity,  $Z = z/z_0$  (with  $z_0 = 1$  m) is the depth relative to  $z_0$  and  $x$  describes the depth dependence of velocity. Taking this into account and considering the well-known relation among  $f_r$  (the resonance frequency), the average S-wave velocity of soft sediments  $\bar{V}_s$ , and the layer thickness  $h$ ,

$$f_r = \bar{V}_s / 4h, \quad (14.5)$$

the dependency between thickness and  $f_r$  can be written as

$$h = \left[ v_{s0} \frac{(1-x)}{4f_r} + 1 \right]^{1/(1-x)} \quad (14.6)$$

where  $f_r$  is to be given in Hz,  $v_{s0}$  in m/s and  $h$  in m. Moreover, empirical relationships between  $f_r$  and  $h$  in the form

$$h = a f_r^b \quad (14.7)$$

can also be derived, generally applying grid search procedures.

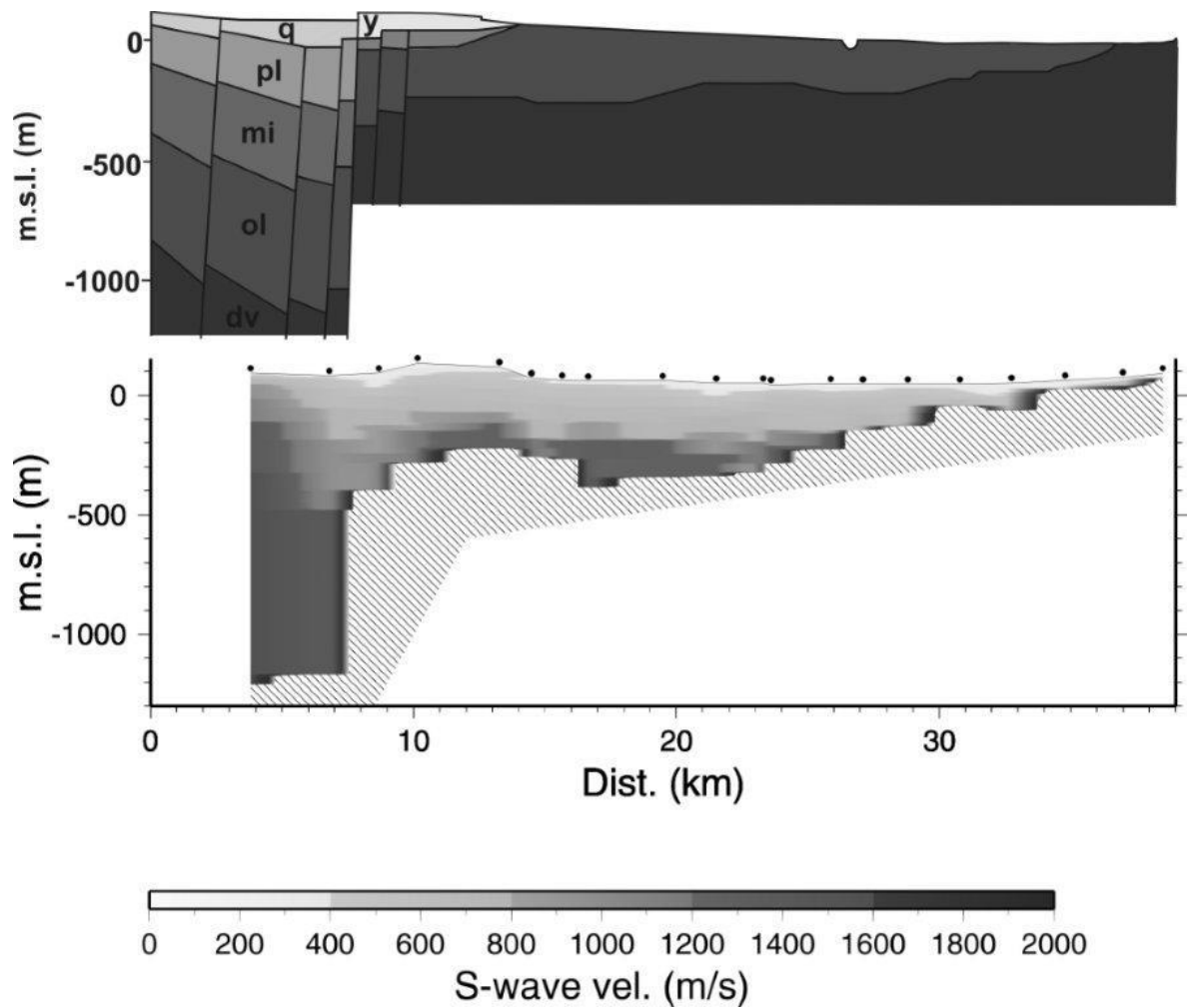
The above approximate interpretation can be easily extended to the case of a two layer sedimentary cover (D'Amico et al., 2008). Despite the fact that relatively large errors affect the depth estimates provided by this approach (D'Amico et al., 2004) it can be considered as a useful proxy for more costly exploratory surveys.

#### 14.3.1.2 NHV inversion

Recently, the possibility of retrieving the S-wave velocity structure below a site from single station measurements based on NHV ratio computation was tested by Fäh et al. (2001). They suggested a new method for calculating NHV ratios employing a frequency-time analysis (FTAN). Moreover, after having shown that there is a good agreement between the NHV ratio and the theoretical ellipticity curves of the fundamental mode Rayleigh wave, they proposed to invert the NHV curve to derive directly the S-wave subsoil structure. The NHV was corrected for the contamination by SH and Love waves by simply reducing it by a factor  $\sqrt{2}$ , independent of frequency, under the assumption of an equal contribution to the seismic noise wavefield of Love and Rayleigh waves. The inversion, due to the non-linear nature of the problem, was based on a genetic algorithm (GA) (Fäh et al., 2001, 2003). The inversion is carried out for a fixed number of layers and a-priori defined ranges of the geophysical properties (S- and P-wave velocity, density and thickness) of the layers. An initial starting population of individuals is generated through a uniform distribution in the parameter space. The model that, amongst all those generated, allows the best reproduction of the observed NHV, is chosen as the best model.

Parolai et al. (2006) adopted this technique to derive S-wave velocity profiles in the Cologne area. Their inversions were carried out by fixing the total thickness of the sedimentary cover in order to avoid problems of trade-off between the total thickness and the S-wave velocity (Scherbaum et al., 2003; Arai and Tokimatsu, 2004). For each site, different inversions were carried out, varying the thickness of the sedimentary layer considering the uncertainty in its estimate. The best models obtained from each inversion were averaged.

In Parolai et al. (2006) the inversion of NHV curves was extended to 20 of the sites measured by Parolai et al. (2001b) and a 2-D S-wave velocity model was derived by means of interpolating between the derived 20 depth profiles. Figure 14.8 shows the resulting 2-D S-wave velocity model (bottom) together with the geological cross-section. The agreement between the geological structure obtained by interpretation of boreholes data and surface geology, and the S-wave velocity model is very good. Compared to the average velocity relationship previously derived for the whole area (Parolai et al., 2002; see also section 14.3.1.1), lateral variations in the velocity structure are clearly visible.



**Fig. 14.8** Top: Geological cross section of the sedimentary cover in Cologne (redrawn after Von Kamp, 1986) y = Landfills, q = Quaternary, pl = Pliocene, mi = Miocene, ol = Oligocene, dv = Devonian. Bottom: 2-D S-wave velocity model interpolated from 1-D S-wave velocity profiles calculated for 20 selected sites where the NHV was inverted. The striped pattern indicates the Devonian bedrock inverted (modified after Parolai et al., 2006).

Therefore, in this case of well constrained bedrock depth, it could be shown that the sedimentary cover is fairly regularly layered, and the NHV inversion a suitable method for quickly mapping 3-D S-wave velocity structures. The vertical resolution of the profiles was also found sufficient to provide site responses (Parolai et al., 2006; Parolai et al., 2007) by means of numerical simulations, in agreement with the empirical ones.

Recently, improvement in the forward calculation of NHV spectral ratios were proposed by Arai and Tokimatsu (2000, 2004) and applied in a joint inversion scheme of NHV and dispersion curves by Parolai et al. (2005), Picozzi and Albarello (2007) and D'Amico et al. (2008).

Arai and Tokimatsu (2000) showed that NHV spectral ratios can be better reproduced if the contribution of higher modes of Rayleigh waves and Love waves is also taken into account. They suggest to calculate the NHV spectral ratio as

$$(\text{NHV})_s = (P_{\text{HS}}/P_{\text{VS}})^{1/2} \quad (14.8)$$

where the sub-index  $s$  stands for surface waves, and  $P_{\text{VS}}$  and  $P_{\text{HS}}$  are the vertical and horizontal powers of surface waves (Rayleigh and Love), respectively.

The vertical power of the surface waves is only determined by the vertical power of Rayleigh waves ( $P_{\text{VR}}$ ), while the horizontal power involves the contribution of both Rayleigh ( $P_{\text{HR}}$ ) and Love waves ( $P_{\text{HL}}$ ). The following equations can therefore be used (Harkrider, 1964):

$$P_{\text{VS}} = P_{\text{VR}} = \sum_j (A_{\text{Rj}}/k_{\text{Rj}})^2 \left\{ 1 + (\alpha^2/2)(u/w)_j^2 \right\} \quad (14.9)$$

$$P_{\text{HS}} = P_{\text{HR}} + P_{\text{HL}} \quad (14.10)$$

$$P_{\text{HR}} = \sum_j (A_{\text{Rj}}/k_{\text{Rj}})^2 (u/w)_j^2 \left\{ 1 + (\alpha^2/2)(u/w)_j^2 \right\} \quad (14.11)$$

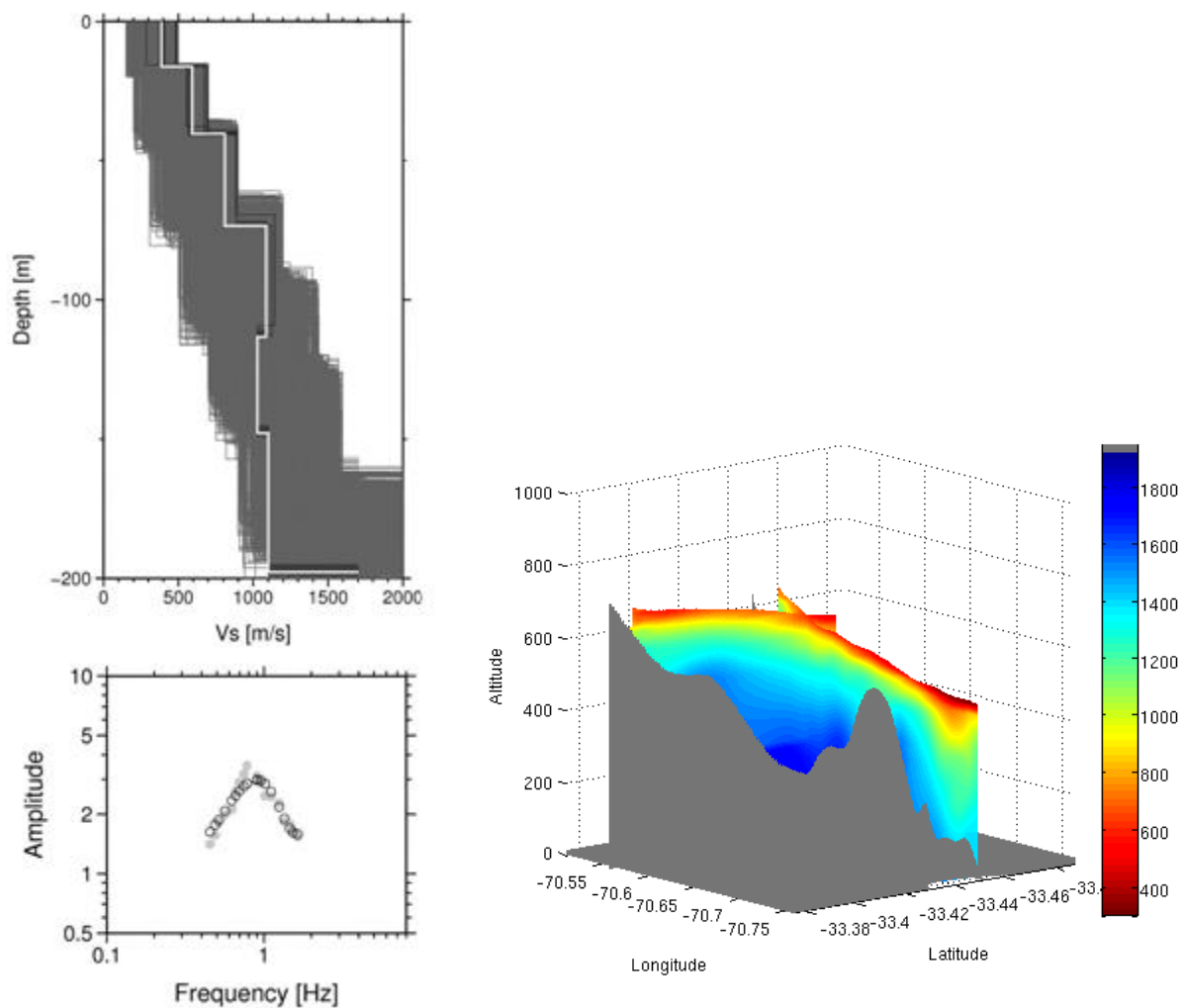
$$P_{\text{HL}} = \sum_j (A_{\text{Lj}}/k_{\text{Lj}})^2 (\alpha^2/2) \quad (14.12)$$

where  $A$  is the medium response,  $k$  is the wavenumber,  $u/w$  is the H/V ratio of the Rayleigh mode at the free surface,  $j$  is the mode index, and  $\alpha$  is the H/V ratio of the loading horizontal and vertical forces  $L_{\text{H}}/L_{\text{V}}$ .

Picozzi et al. (2005b) showed that varying  $\alpha$  over a large range did not significantly changed the NHV shape. Therefore, they used  $\alpha=1$ .

A basic problem of these inversion procedures concerns the choice of the frequency band to be used for the inversion of the NHV curve. As an example, the NHV values around the maximum have been discarded by Parolai et al. (2006) since the effect of smoothing of the spectra biases the amplitude estimation around this frequency band. Instead, they have been taken into account by Picozzi and Albarello (2007) and D'Amico et al., (2008), who tried to exploit the effect of higher modes around the fundamental resonance frequency value. Recent theoretical studies (Lunedei and Albarello, 2009; Albarello and Lunedei, 2009) indicate that the NHV curve around the fundamental resonance frequency  $f_0$  (i.e., around the NHV maximum) can be significantly affected by the damping profile in the subsoil and by the distribution of sources around the receiver. In particular, they showed that sources located within few hundreds of meters from the receiver can generate seismic phases that strongly affect the shape of the NHV curve around and below  $f_0$ . This implies that, unless a large source-free area exists around the receiver, inversion of the NHV shape (around and below  $f_0$ ) carried out by using forward models based on surface waves only, might provide biased results. However, these conclusions still need to be confirmed by comparison with real data.

Recently, Pilz et al. (2010) obtained a 3-D S-wave velocity model of the Santiago de Chile basin by inverting with this method the NHV ratio calculated in Pilz et al. (2009). In order to reduce the trade-off between thickness of the sedimentary cover and S-wave velocity they constrained the former to values derived by previously carried out gravimetric measurements. An example of the results is shown in Figure 14.9.



**Figure 14.9** **Left:** Example of S-wave velocity profile derived by NHV inversion. The best model (white line), the models lying within the 10% of the minimum misfit (black lines) and all the tested models (gray lines) are shown. The bottom panel shows the observed (gray circles) and calculated (empty circles) for the best models NHV. **Right:** The 3-D model of the Santiago de Chile basin obtained by interpolating between the S-wave velocity-depth profiles like the one in the upper left corner of this figure. The color scale indicates the S-wave velocities in m/sec.

## 14.4 2-D arrays

Seismic arrays were originally proposed at the beginning of the 1960s as a new type of seismological tool for the detection and identification of nuclear explosion (e.g., Bormann, 1966; Frosch and Green, 1966). Since then, seismic arrays have been applied at various scales for many geophysical purposes. At the seismological scale, they were used to obtain refined velocity models of the Earth's interior (e.g., Birtill and Whiteway, 1965; Whiteway, 1966; Kværna, 1989; Káráson and van der Hilst, 2001; Ritter et al., 2001; Krüger et al., 2001). Reviews on array applications in seismology can be found in Chapter 9 of this Manual, in Douglas (2002) and in Rost and Thomas (2002). At smaller scales, since the pioneering work of Aki (1957), seismic arrays have been used for the characterization of surface wave



propagation, and the extraction of information on the shallow subsoil structure (i.e., the estimation of the local S-wave velocity profile). Especially in the last decades, due to the focus of seismologists and engineers on estimating the amplification of earthquake ground motion as a function of local geology, the improvements in the quality of instrumentation and computing power, interest in analyzing seismic noise recorded by arrays has grown (e.g., Horike, 1985; Hough et al., 1992; Ohori et al., 2002; Okada, 2003; Scherbaum et al. 2003; Parolai et al., 2005; Tada et al., 2006; Asten, 2006; Köhler et al., 2007). A review on array application for site classification can be found in Foti et al. (2011).

### 14.4.1 f-k based methods

The phase velocity of the surface waves can be extracted from noise recordings by using different methods. Here we will illustrate the two most frequently used methods for f-k (frequency-wavenumber) analysis: the Beam-Forming Method (BFM) (Lacoss et al., 1969) and the Maximum Likelihood Method (MLM) (Capon, 1969).

The estimate of the f-k spectra  $P_b(f, k)$  by the BFM is given by:

$$P_b(f, k) = \sum_{l,m=1}^n \phi_{lm} \exp\{ik(X_l - X_m)\} , \quad (14.13)$$

where  $f$  is the frequency,  $k$  the two-dimensional horizontal wavenumber vector,  $n$  the number of sensors,  $\phi_{lm}$  the estimate of the cross-power spectra between the  $l^{\text{th}}$  and the  $m^{\text{th}}$  recordings, and  $X_l$  and  $X_m$ , are the coordinates of the  $l^{\text{th}}$  and  $m^{\text{th}}$  sensors, respectively.

The MLM gives the estimate of the f-k spectra  $P_m(f, k)$  as:

$$P_m(f, k) = \left( \sum_{l,m=1}^n \phi_{lm}^{-1} \exp\{ik(X_l - X_m)\} \right)^{-1} . \quad (14.14)$$

Capon (1969) showed that the resolving power of the MLM is higher than that of the BFM, however, the MLM is more sensitive to measurement errors.

From the peak in the f-k spectrum occurring at coordinates  $k_{xo}$  and  $k_{yo}$  for a certain frequency  $f_0$ , the phase velocity  $c_0$  can be calculated by:

$$c_0 = \frac{2\pi f_0}{\sqrt{k_{xo}^2 + k_{yo}^2}} . \quad (14.15)$$

An extensive description of these methods can be found in Horike (1985) and Okada (2003).

The estimate  $EP_b$  and  $EP_m$  of the true  $P_b$  and  $P_m$  f-k spectra may be considered in the convolution of the true functions with a frequency window function  $W_f$  and the wavenumber window functions  $W_B$  and  $W_M$  for the BFM and MLM, respectively (Lacoss *et al.*, 1969). The first window function  $W_f$  is the transfer function of the tapering function applied to the signal time windows (Kind, 2005). The function  $W_B$  is termed differently by various authors (e.g., “*spatial window function*” by Lacoss et al., 1969, and “*beam-forming array response function*” by Capon, 1969), and hereafter simply defined as Array Response Function (ARF). The ARF depends only on the distribution of stations in the array and has for the wavenumber vector  $k_o$  the form (Horike, 1985)



$$W_B(k, k_o) = \frac{1}{n^2} \sum_{l,m=1}^n \exp\{i(k - k_o)(X_l - X_m)\}. \quad (14.16)$$

Simply speaking, it represents a kind of spatial filter for the wavefield. The main advantage of the MLM with respect to the BFM involves the use of an improved wavenumber window  $W_M$ . Namely, for a wavenumber  $k_o$  this window function may be expressed in the form

$$W_M(f, k, k_o) = \left| \sum_{j=1}^N A_j(f, k_o) \right| W_B(k, k_o), \quad (14.17)$$

where

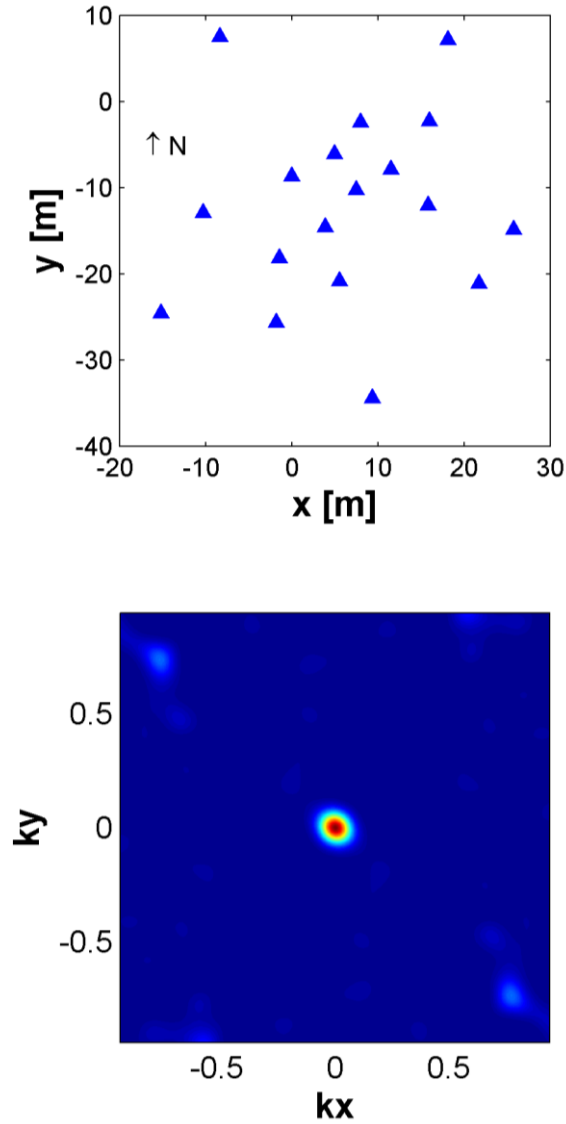
$$A_j(f, k_o) = \frac{\sum_{l=1}^N q_{jl}(f, k_o)}{\sum_{j,l=1}^N q_{jl}(f, k_o)} \quad (14.18)$$

and  $q_{jl}$  represents the elements of the cross-power spectral matrix. It is evident that  $W_M$  depends not only on the array configuration, through the function  $W_B$ , but also on the quality (i.e., signal-to-noise ratio) of the data (Horike, 1985). In fact, the wavenumber response is modified by using the weights  $A_j(f, k_o)$ , which depend directly on the elements  $q_{jl}(f)$ . In practice,  $W_M$  allows the monochromatic plane wave travelling at a velocity corresponding to the wavenumber  $k_o$  to pass undistorted, while it suppresses, in an optimum least-square sense, the power of those waves travelling with velocities corresponding to wavenumbers other than  $k_o$  (Capon, 1969). Or, in other words, coherent signals are associated with large weights of  $A_j$  and their energy is emphasized in the f-k spectrum. On the contrary, if the coherency is low, the weights  $A_j$  are small and the energy in the f-k spectrum is damped (Kind et al., 2005). This automatic change of the main-lobe and side-lobe structure minimizes the leakage of power from the remote portion of the spectrum and has a direct positive effect on the  $P_m$  function, and consequently on the following velocity analysis. However, considering the dependence of  $W_M$  on  $W_B$ , it is clear that the array geometry is a factor which has a strong influence on both  $EP_b$  and  $EP_m$ . In fact, similarly to every kind of filter, several large side lobes located around the major central peak can remain in the f-k spectra (Okada, 2003) and cause serious biases in the velocity and back-azimuth estimates. In particular, side-lobe height and main-lobe width within  $W_B$  control the leakage of energy and resolution, respectively (Zywicki, 1999).

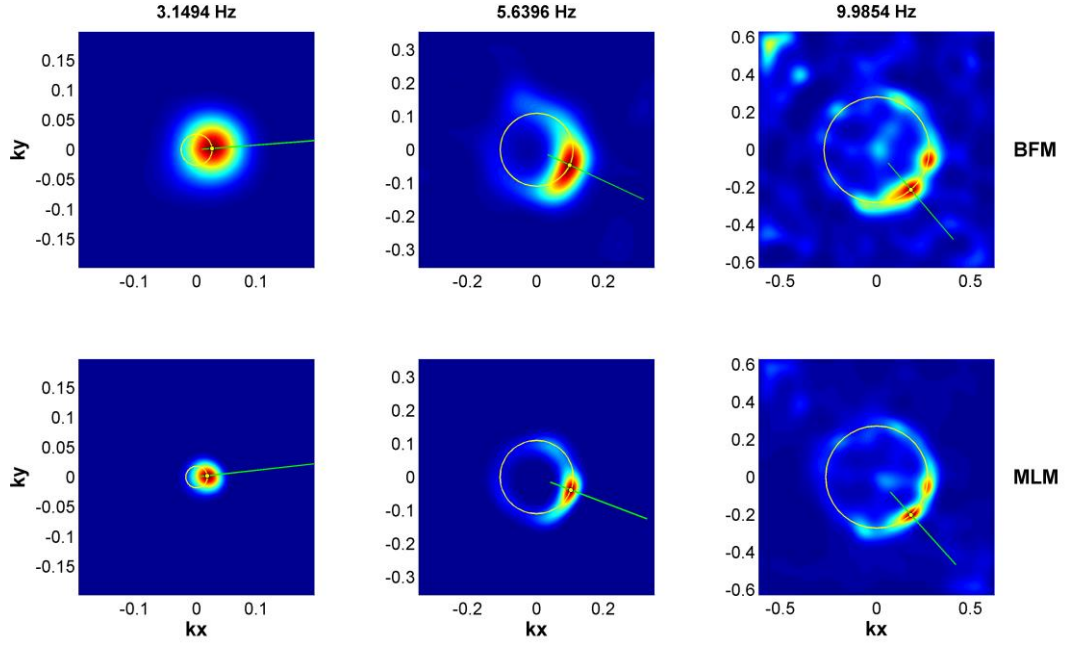
As a general criterion, the error in the velocity analysis due to the presence of spurious peaks in the f-k spectra may be reduced using distributions of sensors for which the array response approaches a two-dimensional  $\delta$ -function. For that reason, it is considered a good practice to undertake a preliminary evaluation of the array response when the survey is planned. Irregular configurations of even only a few sensors should be preferred, because they allow one to obtain a good compromise between a large aperture, which is necessary for sharp main peaks in the  $EP_b$  and  $EP_m$ , and small inter-sensor distances, which are needed for large aliasing periods (Kind et al., 2005).

Recently, Picozzi et al. (2010a) showed that removing the effect of the array response from the frequency wavenumber images by using the Richards-Lucy regularization method improve the phase velocity estimation.

Fig.14.10 shows an example of 2-D-array configuration and its respective array responses. Figure 14.11 shows an example of f-k analysis results for 3.15 Hz, 5.6 Hz, and 9.9 Hz for the array configuration of Fig. 14.10. White dots indicate the position of the maximum used to estimate the phase velocity while the white circles join points with the same k values.



**Fig. 14.10 Top:** Example of a micro-array configuration. The blue triangles indicate the station positions. **Bottom:** The relevant micro-array response estimated for the frequency of 15 Hz.



**Fig. 14.11** Example of results from f-k analysis for 3.15 Hz, 5.6 Hz, and 9.9 Hz. White dots indicate the position of the maximum used to estimate the phase velocity. The circles join points with the same  $k$ . In this case at frequencies of 5.6 and 9.9 Hz the velocity of propagation of waves can be estimated and the source position identified (red color in the maps). (Comment: There are as many white dots as circles. Thus both has to be plural)

#### 14.4.2 Spatial and Extended Spatial Autocorrelation method (SPAC and ESAC )

Aki (1957 and 1965) showed that phase velocities in sedimentary layers can be determined using a statistical analysis of ambient noise. He assumed that noise represents the sum of plane waves propagating without attenuation in a horizontal plane in different directions with different powers, but with the same phase velocity for a given frequency. He also assumed that waves with different propagation directions and different frequencies are statistically independent. Then, a spatial correlation function can be defined as

$$\phi(r, \lambda) = \langle u(x, y, t)(x + r \cos(\lambda), y + r \sin(\lambda), t) \rangle \quad (14.19)$$

where  $u(x, y, t)$  is the velocity observed at point  $(x, y)$  at time  $t$ ;  $r$  is the inter-station distance;  $\lambda$  is the azimuth and  $\langle \rangle$  denotes the ensemble average. An azimuthal average of this function is given by

$$\phi(r) = \frac{1}{\pi} \int_0^\pi \phi(r, \lambda) d\lambda. \quad (14.20)$$

For the vertical component, the power spectrum  $\phi(\omega)$  can be related to  $\phi(r)$  via the zero<sup>th</sup> order Hankel transform:

$$\phi(r) = \frac{1}{\pi} \int_0^{\infty} \phi(\omega) J_0\left(\frac{\omega}{c(\omega)} r\right) d\omega, \quad (14.21)$$

where  $\omega$  is the angular frequency,  $c(\omega)$  is the frequency-dependent phase velocity, and  $J_0$  is the zero order Bessel function. The space-correlation function for one angular frequency  $\omega_0$ , normalized to the power spectrum, will be of the form

$$\phi(r, \omega_0) = J_0\left(\frac{\omega_0}{c(\omega_0)} r\right). \quad (14.22)$$

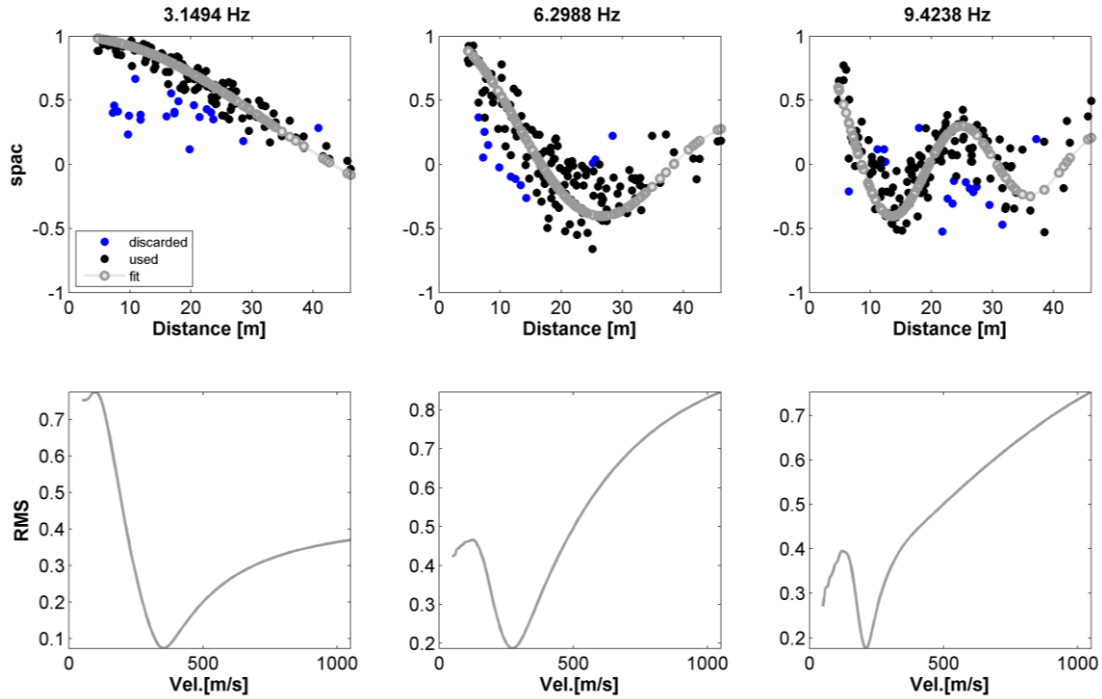
By fitting the azimuthally averaged spatial correlation function obtained from measured data to the Bessel function, the phase velocity  $c(\omega_0)$  can be calculated. A fixed value of  $r$  is used in the spatial autocorrelation method (SPAC). However, Okada (2003) and Ohori et al. (2002) showed that, since  $c(\omega)$  is a function of frequency, better results are achieved by fitting the spatial-correlation function at each frequency to a Bessel function, which depends on the inter-station distances (extended spatial autocorrelation, ESAC). For every couple of stations the function  $\phi(\omega)$  can be calculated in the frequency domain by means of (Malagnini et al., 1993; Ohori et al., 2002; Okada, 2003):

$$\phi(\omega) = \frac{\frac{1}{M} \sum_{m=1}^M \text{Re}( {}_m S_{jn}(\omega) )}{\sqrt{\frac{1}{M} \sum_{m=1}^M {}_m S_{jj}(\omega) \sum_{m=1}^M {}_m S_{nn}(\omega)}}, \quad (14.23)$$

where  ${}_m S_{jn}$  is the cross-spectrum for the  $m$ th segment of data, between the  $j^{\text{th}}$  and the  $n^{\text{th}}$  station;  $M$  is the total number of used segments. The power spectra of the  $m^{\text{th}}$  segments at station  $j$  and station  $n$  are  ${}_m S_{jj}$  and  ${}_m S_{nn}$ , respectively.

The space-correlation values for every frequency are plotted as a function of distance, and an iterative grid-search procedure can then be performed using equation (14.22) in order to find the value of  $c(\omega_0)$  that gives the best fit to the data. The tentative phase velocity  $c(\omega_0)$  is generally varied over large intervals (e.g., between 100 and 3000 m/s) in small steps (e.g., 1 m/s). The best fit is achieved by minimizing the root mean square (RMS) of the differences between the values calculated with equation (14.21) and (14.22). Data points, differing more than two standard deviations from the value obtained with the minimum-misfit velocity, can be removed before the next iteration of the grid-search. Parolai et al. (2006) used this procedure and allowed for a maximum of three grid-search iterations. An example of the application of this procedure is shown Fig. 14.12.

Several studies, (e.g. Parolai et al., 2009, Foti et al., 2011) showed the reliability of the procedure in deriving a robust estimate of the dispersion curve that lead to S-wave velocity profiles consistent with those obtained by other geophysical methods.

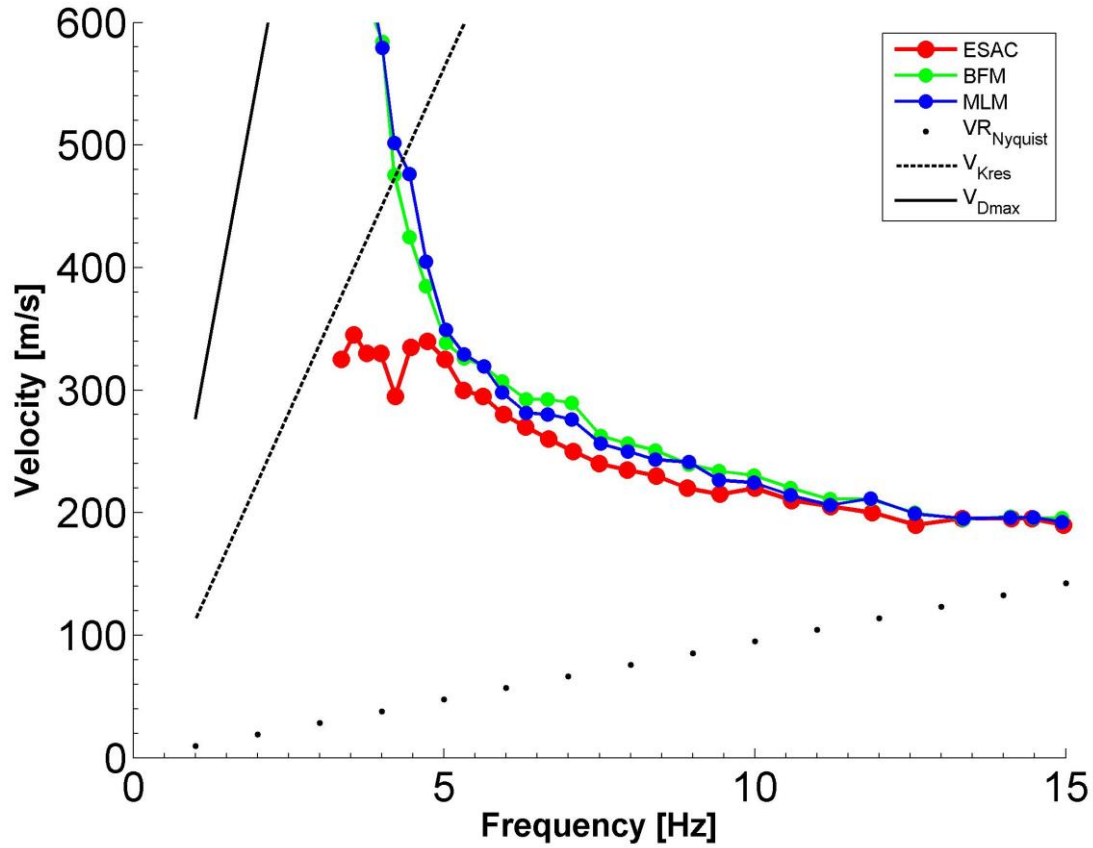


**Fig. 14.12 Top:** Measured space-correlation function values for different frequencies (black and blue circles) and the best-fitting Bessel function (gray circles). Discarded values (blue circles) lie two standard deviations outside the curve. **Bottom:** The respective RMS error versus phase velocity curves.

The ESAC method can be adopted to derive the phase velocities for all frequencies composing the Fourier spectrum of the data. Figure 14.12 shows on top three examples of the space-correlation values computed from the data together with the best fitting Bessel functions and below the corresponding RMS errors as function of the tested phase velocities, exhibiting clear minima. For high frequencies, the absolute minimum might sometimes corresponds to the minimum velocity chosen for the grid search procedure. This solution is then discarded, because a smooth variation of the velocity between close frequencies is required. At frequencies higher than a certain threshold the phase velocity might increase linearly. This effect is due to spatial aliasing which limits the upper bound of the usable frequency band. It depends on the S-wave velocity structure at the site and the minimum inter-station distance. At low frequencies the RMS error function clearly indicates the lower boundary for acceptable phase velocities, but might not be able to constrain the higher ones (then a plateau can appear in the RMS curves). The frequency above which phase differences cannot be resolved any more depends on the maximum inter-station distance and the S-wave velocity structure below the site: a wide range of velocities will then explain the observed small phase differences. Zhang et al. (2004) clearly pointed out this problem in Equation (3a) of their article.

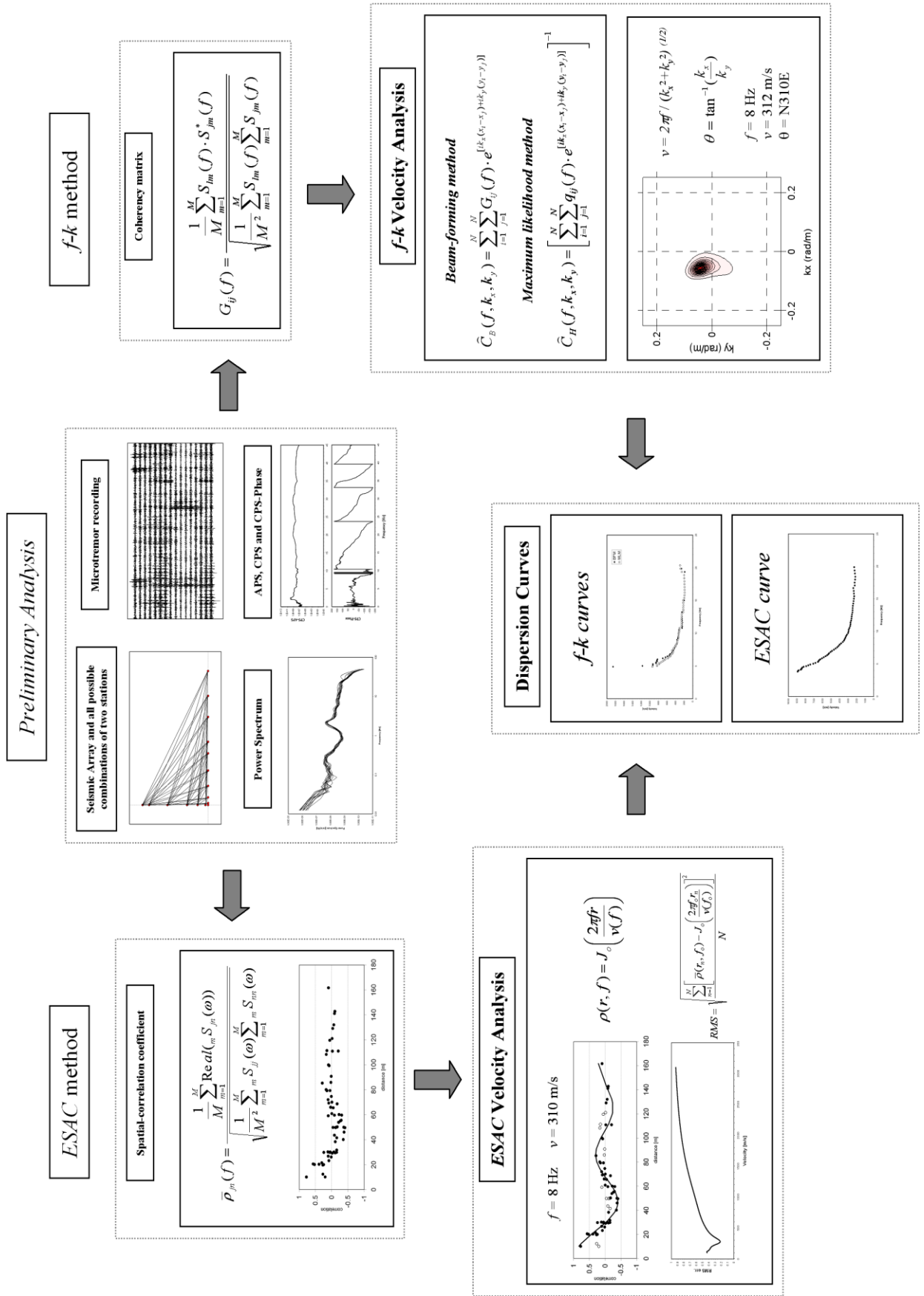
Figure 14.13 gives an example of the final dispersion curve. The typical behaviour of f-k methods, namely of BFM to yield higher phase velocities than the ESAC method at low frequencies (Okada 2003), is clearly to be seen. Yet, Okada (2003) showed that by increasing the array size, the f-k velocities at low frequencies become similar to those estimated by ESAC, indicating that the latter is more suitable for providing reliable dispersion curves over

a wider frequency range, especially toward lower frequencies (and thus with greater penetration depths).



**Fig. 14.13** Phase velocity dispersion curves obtained by ESAC (reds line), BFM (green line) and MLM (blue line) analysis. The dotted line indicate the theoretical aliasing limit calculated as  $4 f_{dmin}$ , where  $d_{min}$  is the minimum interstation distance. The factor 4 is used instead of the generally used factor 2, because the minimum distance in the array is appearing only once. The dashed gray line indicates the lower frequency threshold of the analysis based on  $2\pi f \Delta k$  where  $\Delta k$  is calculated as the half-width of the main peak in the array response function. The continuous line indicates the lower frequency threshold of the analysis based on the equation  $3 f_{dmax}$ , where  $d_{max}$  is the maximum interstation distance in the array

Figure 14.14 summarises the procedure used with the ESAC and f-k methods.



**Fig. 14.14** Summary of the procedure used with the ESAC and *f-k* methods (from Picozzi, 2006)

## 14.5 Interferometry and tomography

Recent theoretical studies have shown that the cross-correlation of diffuse wavefields can provide an estimate of the Green's functions between receivers (Weaver and Lobkis, 2001 and 2004; Snieder, 2004; Wapenaar, 2004; Wapenaar and Fokkema, 2006). Using coda waves of seismic events (Campillo and Paul, 2003) and long seismic noise sequences (Shapiro and Campillo, 2004), it was confirmed that it is possible to estimate the Rayleigh wave component of Green's functions between two stations by the cross-correlation of simultaneous recordings. This method is now generally referred to as seismic interferometry. These results allowed the first attempts of surface wave tomography at regional scales (e.g., Shapiro and Campillo, 2004; Sabra et al., 2005; Shapiro et al., 2005; Gerstoft et al., 2006; Yao et al., 2006; Cho et al., 2007; Lin et al., 2007; Yang et al., 2007) using seismic noise recordings from broad-band seismic networks. Generally, for these kinds of studies, waves at frequencies well below 1 Hz were used to image the crust and the upper-mantle structure. A comprehensive review of the seismic interferometry method can be found in Curtis et al. (2006).

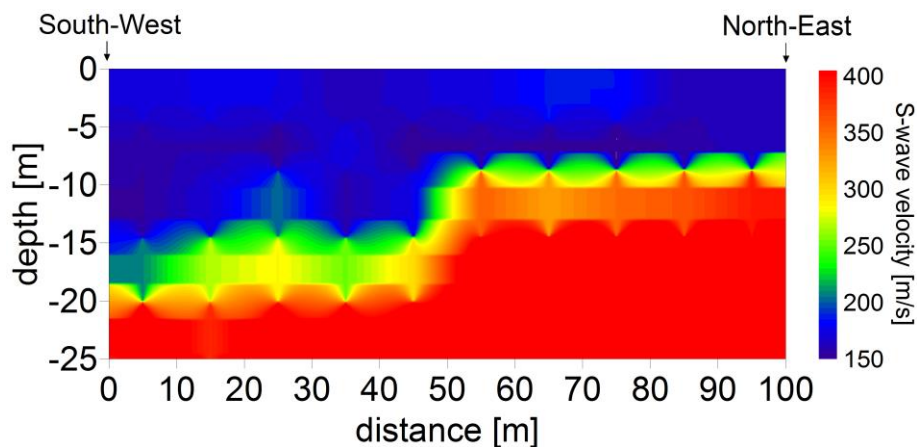
Seismic noise interferometry can also be applied to frequencies larger than 1Hz. Schuster (2001) and Schuster et al. (2004) demonstrated the possibility of forming an image of the subsurface using the cross-correlation of seismic responses from natural and man-made sources at the surface or in the subsurface. Furthermore, within the context of exploration geophysics, Bakulin and Calvert (2004 and 2006) first proposed a practical application of seismic interferometry, showing that it is possible in practice to create a virtual source at a subsurface receiver location in a well. Other recent applications for the high-frequency range have been proposed by Dong and Schuster (2006) and Halliday et al. (2007) for surface-wave isolation and removal in active-source surveys. Among the several reasons that have stimulated the application of seismic noise interferometry to high frequencies, one is the possibility of applying this technique to suburban settings (Halliday et al., 2008), and another to exploit it for engineering seismology purposes, which require knowledge of the subsurface structure from depths of a few meters to several hundred meters.

The application of seismic noise interferometry to high frequencies is not merely a change of scale, since it involves also important questions still under discussion within the research community. For example, the effects of the high spatial and temporal variability in the distribution of noise sources occurring at high frequencies are still under investigation (Halliday and Curtis, 2008a and b), as well as the relationship between the wavelength of interest and inter-station distances. Several authors (e.g., Chavez-Garcia and Luzon, 2005; Chavez-Garcia and Rodriguez, 2007; Yokoi and Margaryan, 2008) showed, for a small scale experiment at a site with a homogeneous subsoil structure, the equivalence between the results obtained by cross-correlation in the time domain and the SPatial Auto-Correlation analysis (SPAC) method (Aki, 1957). However, it is worth noting that for non-homogeneous subsoil conditions, the SPAC method suffers a severe drawback. Indeed, as a general rule, such a method is used to retrieve the shallow soil structure below a small array of sensors by means of inversion of surface wave dispersion curves extracted by seismic noise analysis. Yet, the inversion is performed under the assumption that the structure below the site is nearly 1-D. Therefore, if the situation is more complicated (2-D or 3-D structure), then the SPAC method can only provide a biased estimate of the S-wave velocity structure. In contrast, one can expect that, similarly to what is obtained over regional scales, local heterogeneities will affect the noise propagation between sensors, and hence can be retrieved by analysing the Green's function estimated by the cross-correlation of the signals recorded at two different stations. For this reason, passive seismic interferometry is also believed to be a valuable tool for studying complex structures and performing surface wave tomography also for smaller



spatial scales of investigation. Picozzi et al. (2008) verified the suitability of seismic interferometry for seismic engineering and microzonation purposes. In fact, after having first evaluated the possibility of retrieving reliable and stable Green's functions within the limitations of time and instrumentation that bound standard engineering seismological experiments (for example, in urban microzonation studies, the number of deployed sensors is generally not larger than 20 and the acquisition time does not last more than a few hours), they applied the seismic interferometry technique to recordings from a 21-station array installed in the Nauen test site (Germany) (Yaramanci et al., 2002; and [http://www.geophysik.tu-berlin.de/menue/forschung/testfeld\\_nauen/uebersicht](http://www.geophysik.tu-berlin.de/menue/forschung/testfeld_nauen/uebersicht)). They showed that passive seismic interferometry is a valuable tool for the characterization of near-surface geology since the travel times estimated from the Green's functions analysis for different frequencies were inverted to derive the laterally varying 3-D surface wave velocity structure below the array, and with a few further steps the S-wave velocity structure as well. This represents a major innovation, due to the frequency range investigated and the scale of the experiment,

Figure 14.15 shows a 2-D cross section highlighting lateral velocity variations as inferred by seismic tomography from the application of seismic interferometry to seismic noise (Picozzi et al., 2008).



**Fig. 14.15:** S-wave velocity section extending southwest to northeast in the centre of the study area of Picozzi et al. (2008), derived by seismic noise tomography.

Furthermore, recent technical development (Picozzi et al., 2010b), show that in future it will be possible to overcome the problem of the high costs associated with of the application of this technique, caused by the large number of instruments required by using ad-hoc developed low-cost sensors.

## 14.6 Conclusions, suggestions and related case studies

Estimating site effects in large urban areas is a challenging task for seismologists and engineers. On the one hand, detailed site investigations are necessary for optimal design of buildings and foundations, while on the other hand urban planning and seismic risk mitigation actions could rely also on more general information about site amplification. Furthermore, the area that should be covered by related investigations for megacities might require large

investments and a long duration of the experiments. The rapid growth and expansion of populated areas, especially in developing countries, might even render the studies already outdated soon after their completion. A reasonable compromise between collection of empirical data at acceptable cost and within realistic time-frames can only be based on a mixed approach that combines the acquisition of high-quality (and thus high-cost) information at few sites with the application of rapid low-cost methods that allow for an extrapolation of the information over wide areas. With this aim, seismic microzonation of large urban areas can be carried out on the basis of site effect studies, which should involve:

- 1) Collection of all already available geological, geotechnical, and geophysical data.
- 2) Installation of a temporary seismic network for recording of earthquake data (about 3 months in duration). Local, regional and teleseismic events provide the necessary inputs for site effect estimation. The different frequency content of the recordings allows to cover the whole frequency range of engineering interest. The stations should be installed at characteristic sites taking into account the surface geology, topography and the expected depth variations of the bedrock surface.
- 3) Seismic noise recordings at single stations. Measurements should be carried out for at least 30 minutes at each site following the recommendation of SESAME (2003) ([http://sesame-fp5.obs.ujf-grenoble.fr/SES\\_Home\\_Description.htm](http://sesame-fp5.obs.ujf-grenoble.fr/SES_Home_Description.htm)) and Picozzi et al. (2005a). In particular attention should be paid to the instruments used (Guiller et al., 2007; Strollo et al., 2008a and b). After calibration through comparison with earthquake data the fundamental resonance frequency estimated by NHV can be used for preparing maps. If an estimation of the thickness of the sedimentary layers is available, NHV can be inverted to derive S-wave velocity profiles. These, in turn, can be used to calculate 1-D site response or to generate 2-D and 3-D models that provide a basis for computing ground motion scenarios.
- 4) S-wave velocity profiles can be estimated using seismic noise arrays. .

Although these noise-based methods cannot be considered in general a substitute of investigation techniques dealing with active sources (that might provide more detailed information about the subsoil structure) they have the advantage to be easily applicable in urban areas (where method based on active sources might be impossible or suffer problems) and allow for investigating the subsurface structure down to large depths in a cost-effective way.

Tables 14.1 and 14.2 summarize essential points to be considered in the acquisition and analysis of earthquake data and seismic noise data, respectively, when carrying out such investigations. In the case that non-linear soil behaviour is likely this should be investigated by drilling of shallow boreholes and collection of undisturbed samples at a limited number of representative sites. If available funding allows for long-term monitoring, and improvement of estimates of site response for different levels of ground motion, the installation of vertical array of accelerometers should be considered. A similar combined approach was recently followed for different large urban areas. Parolai et al. (2004) installed a network of 40 seismological stations in the urban area of Cologne (Germany). Moreover they performed single station noise analysis at nearly 400 sites (Parolai et al., 2001b) and estimated the S-wave velocity at two sites by means of seismic noise array data analysis (Parolai et al. 2005; Parolai et al. 2006).

Parolai et al. (2010b) installed a network of 20 seismological stations in Bishkek and carried out single station noise measurements at nearly 200 sites (Figure 15.7).

**Table 14.1 Suggestions for collection and analysis of earthquake data**

|                                                                   |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|-------------------------------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Acquisition</b>                                                | <ul style="list-style-type: none"> <li>- A sample rate of at least 100 s.p.s is suggested</li> <li>- Continuous recording allows to collect seismic noise and to record also regional and teleseismic events that might be masked by high frequency noise</li> <li>- Short period sensors might provide a reasonable compromise between the data quality and the temporary (low quality generally) installation.</li> </ul>                                                                                                                                                                                                                                                                                                                                    |
| <b>Data analysis: window selection</b>                            | <p>The window length can be selected following different criteria:</p> <ul style="list-style-type: none"> <li>- fixed window lengths (be careful that all the significant part of ground motion is included);</li> <li>- window length selected on the base of energy criteria, e.g., including only part of signal containing 80% or 90% of the energy.</li> </ul> <p>Remember that frequency resolution is related to the total window length used.</p> <p>Since generally windows are tapered at both ends before the Fourier transform, the starting of the window should be selected in order to avoid that tapering is affecting the signal of interest. If, for example S waves are of interest, the window should start before the S-wave arrival.</p> |
| <b>Data analysis: Baseline correction, Tapering and smoothing</b> | <p>Removing the average of the signal might be necessary before calculating the Fast Fourier transform (FFT).</p> <p>Tapering at both ends is generally performed using windows. The percentage of tapering can vary. A generally used value is 5% of the signal length.</p> <p>The FFT of the signal should be smoothed, in order to avoid numerical instabilities. However, smoothing should allow to preserve the main features in the spectra. When spectra are spanning more than one decade it is preferable to smooth them using relative bandwidth windows.</p>                                                                                                                                                                                        |

**Table 14.2 Suggestions for collection and analysis seismic noise data**

|                                                                                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|------------------------------------------------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Acquisition</b>                                                                             | <ul style="list-style-type: none"> <li>- A sample rate of at least 100 s.p.s is recommended for single station noise measurements. Depending on the minimum inter-station distance, in the case of arrays, even larger sampling rates are suggested.</li> <li>- Short-period sensors might provide a reasonable compromise between high data quality and temporary installations of generally low data quality.</li> <li>- Please, check if the level of seismic noise is higher than the internal noise of the sensor+acquisition system in the frequency range of interest (see Chapter 5, section 5.5.6).</li> </ul>                                                                        |
| <b>Data analysis: window selection</b>                                                         | <p>The windows should be selected keeping in mind the frequency band of interest. The maximum usable frequency is limited by the Nyquist frequency due to the sampling rate. The minimum frequency that can be analyzed is determined by the length of the selected signal window. However, stable results are expected only for frequencies that have a reasonable number of cycles (at least 10) within the selected signal window.</p> <p>In the case that the frequency range of interest spans between 0.1 and 10 Hz, window lengths of 20-60 seconds are suggested.</p>                                                                                                                  |
| <b>Data analysis: Baseline correction, tapering, smoothing and number of analysed windows.</b> | <p>Removing the average of the signal might be necessary before calculating the FFT.</p> <p>Tapering at both ends is generally performed using windows. The percentage of tapering can vary. A generally used value is 5% of the signal length.</p> <p>The FFT of the signal should be smoothed, in order to avoid numerical instabilities. However, smoothing should allow to preserve the main features in the spectra. When spectra are spanning more than one decade it is preferable to smooth them using relative bandwidth windows.</p> <p>When the frequency range of interest is 0.1-10 Hz at least 20 minutes of noise (divided in windows of 20-60 seconds) should be analysed.</p> |

Similarly, Picozzi et al. (2009), estimated the fundamental resonance frequency in western Istanbul (Turkey) by means of nearly 200 single station noise measurements. The reliability of the results was assessed by comparison with results from earthquake data obtained by analysing the recordings of the Istanbul Earthquake Rapid Response System (IERRS) operated by the Kandilli Observatory and Earthquake Research Institute of Bogazici University. Furthermore, eight 2-D array measurements of seismic noise were performed in the metropolitan area with the aim of obtaining a preliminary geophysical characterization of the different sedimentary covers. Comparison of the theoretical site response from an estimated S-wave velocity profile with the empirical one derived using earthquake recordings confirmed the suitability of the use of the low-cost seismic noise techniques for the study of seismic site effects.

Pilz et al. (2009) installed a temporary seismic network in Santiago de Chile (~6 million inhabitants). They estimate site amplification using the standard spectral ratio (SSR) method and proposed a map of the fundamental resonance frequency by analysing ~150 single station seismic noise recordings. Furthermore, by taking advantage of the available knowledge on bedrock depth from gravimetric measurements they derived a 3-D model of the basin that can be used for calculating realistic ground motion scenarios.

All these studies showed that the approach described above is feasible for covering large urban areas. at reasonable costs. Open questions still remain with respect to the estimation of a first order classification of soil that can be used, with a sufficient degree of reliability, at regional or even global scales.

## Acknowledgment

I am grateful to Dino Bindi, Matteo Picozzi, Marco Pilz and Tobias Boxberger for suggestions and comments that helped in improving this manuscript. Their help in drawing figures was also strongly appreciated. The manuscript also benefitted from valuable comments and suggestions by Peter Bormann, Adrien Oth and Pierre-Yves Bard.

## References

- Aki, K. (1957). Space and time spectra of stationary stochastic waves with special reference to microtremors. *Bull. Earthqu. Res. Inst.*, **35**, 415-456.
- Aki, K. (1965). A note on the use of microseisms in determining the shallow structure of the Earth's crust. *Geophysics*, **30**, 665-666
- Albarelo, D., and Lunedei, E. (2009). Alternative interpretations of horizontal to vertical spectral ratios of ambient vibrations: new insights from theoretical modeling. *Bull. Earthq. Eng.*; doi: 10.1007/s10518-009-9110-0
- Andrews, D. J. (1986). Objective determination of source parameters and similarity of earthquakes of different sizes. In: Das, S., Boatwright, J., and Scholz, C. H. (Editors), *Earthquake Source Mechanics*, American Geophysical Union, Washington D.C., 259–268.
- Arai, H., and K. Tokimatsu (2000). Effects of Rayleigh and Love waves on microtremor H/V spectra, paper presented at *12th World Conference on Earthquake Engineering*, N. Z. Soc. for Earthquake Eng., Auckland, N. Z.
- Arai, H., and K. Tokimatsu (2004). S-wave velocity profiling by inversion of microtremor H/V spectrum, *Bull. Seism. Soc. Am.* **94**, 53–63.

- Assimaki, D., Li, W., Steidl, J. H., and Tsuda, K. (2008), Site amplification and attenuation via downhole array seismogram inversion: a comparative study of the 2003 Miyagi-Oki aftershock sequence. *Bull. Seism. Soc. Am.* **98**, 301-330.
- Asten, M. W. (2006). On bias and noise in passive seismic data from finite circular array data processed using SPAC method. *Geophysics*, **71**, V153-V162, doi:10.1190/1.2345054.
- Bakulin, A. and Calvert, R. (2004). Virtual source: new method for imaging and 4D below complex overburden. In: *Proceedings of 74th Annual International Meeting* (expanded abstracts), *Society of Exploration Geophysicists*, 2477–2480.
- Bakulin, A. and Calvert, R. (2006). The virtual source method: theory and case study. *Geophysics*, **71**, S1139–S1150.
- Bard, P.-Y. (1995). Effects of surface geology on ground motion: recent results and remaining issues. In: *10th European Conference on Earthquake Engineering*, Duma (Editor), Balkema, Rotterdam, 305–323.
- Bard, P.-Y. (1999). Microtremor measurements: A tool for site effect estimation? In: Irkura, K., Kudo, K., Okada, H., and Sasatani, T. (Eds.) (1999). The effects of surface geology on seismic motion. *Proc. 2<sup>nd</sup> International Symposium in Yokohama, Japan, 1-3 Dec. 1998*, A. A. Balkema, Rotterdam, ISBN 90 5809 030 2, 1251-1279.
- Bindi, D., Parolai, S., Cara, F., Di Giulio, G., Ferretti, G., Luzi, L., Monachesi, G., Pacor, F., and Rovelli, A. (2009). Site amplifications observed in the Gubbio Basin, Central Italy: Hints for lateral propagation effects. *Bull. Seism. Soc. Am.*, **99**, 741-760.
- Birtill, J. W., and Whiteway, F. E. (1965). The application of phased arrays to the analysis of seismic body waves. *Phil. Trans. R. Soc. of London, Math. and Phys. Sciences A-258*, No. 1091, 421-493.
- Borcherdt, R. D. (1970). Effects of local geology on ground motion near San Francisco Bay. *Bull. Seism. Soc. Am.*, **60**, 29–61.
- Bonilla, L. F., Steidl, J. H., Lindley, G. T., Tumarkin, A. G., and Archuleta, R. J. (1997). Site amplification in the San Fernando Valley, California: variability of site effect estimation using S-wave, coda, and H/V methods. *Bull. Seism. Soc. Am.*, **87**, 710–730.
- Bonilla L.F., Archuleta R.J., and Lavallée, D. (2005). Hysteretic and dilatant behavior of cohesionless soils and their effects on nonlinear site response: Field data observations and modelling. *Bull. Seism. Soc. Am.*, **95**, 2373 - 2395.
- Bonnefoy-Claudet S., Baize, S., Bonilla, L. F., Berge-Thierry, C., Pasten, C. R., Campos, J., Volant, P., Verdugo, R. (2009). Site effect evaluation in the basin of Santiago de Chile using ambient noise measurements. *Geophys. J. Int.*, **176** (3), 925–937.
- Bormann, P. (1966). Recording and interpretation of seismic events (principles, present state and tendencies of development) (in German). *Publ. Inst. Geodyn., Jena, Akademie Verlag*, Berlin, Vol. 1, 158 pp.
- Campillo, M. and Paul, A. (2003). Long-range correlations in the seismic coda. *Science*, **299**, 547–549.
- Capon, J. (1969). High-resolution frequency-wavenumber spectral analysis. *Proc. IEEE.*, **57**, 1408-1419.
- Castro, R. R., Anderson, J. G., and Singh, S. K. (1990). Site response, attenuation and source spectra of S-waves along the Guerrero, Mexico, subduction zone. *Bull. Seism. Soc. Am.* **80**, 1481–1503
- Chavez-Garcia, F. J. and Luzon, F. (2005). On the correlation of seismic microtremors. *J. Geophys. Res.*, **110**, B11313, doi:10.1029/2005JB003671
- Chavez-Garcia, F. J. and Rodriguez, M. (2007). The correlation of microtremors: empirical limits and relations between results in frequency and time domains. *Geophys. J. Int.*, **171**, 657–664.

- Cho, K. H., Herrmann, R. B. and Ammon, C. J. and Lee, K. (2007). Imaging the upper crust of the Korean Peninsula by surface-wave tomography. *Bull. Seism. Soc. Am.*, **97**, 198–207, doi:10.1785/0120060096.
- Curtis, A., Gerstoft, P., Sato, H., Snieder, R. and Wapenaar, K. (2006). Seismic interferometry - turning noise into signal. *Leading Edge*, **25**, 1082–1092.
- D’Amico V., Picozzi M., Albarello D., Naso G. and Tropenscovino S. (2004). Quick estimate of soft sediments thickness from ambient noise horizontal to vertical spectral ratios: a case study in southern Italy”. *J. Earthq. Engineering*, **8**, 6, 895-908.
- D’Amico V., Picozzi M., Baliva F. and Albarello D. (2008). Ambient Noise Measurements for Preliminary Site-Effects Characterization in the Urban Area of Florence. *Bull.Seism.Soc.Am.*, **98**, 3, 1373-1388, doi: 10.1785/0120070231
- Delgado, J., C. Lopez Casado, A. C. Estevez, J. Giner, A. Cuenca and S. Molina (2000a). Mapping soft soils in the Segura river valley (SE Spain): a case study of microtremors as an exploration tool, *J. Appl. Geophys.* **45**, 19–32.
- Delgado, J., C. Lopez Casado, J. Giner, A. Estevez, A. Cuenca, and S. Molina. (2000b). Microtremors as a geophysical exploration tool: applications and limitations, *Pure Appl. Geophys.* **157**, 1445–1462. 1841.
- Dong, S., He, R. and Schuster, G. (2006). Interferometric prediction and leastsquares subtraction of surface waves, in *Proceedings of 76th Annual International Meeting* (expanded abstracts), Society of Exploration Geophysicists, 2783–2786.
- Douglas, A., (2002). Seismometer arrays – Their use in earthquake and test ban seismology, in *Handbook of Earthquake and Engineering Seismology*, edited by W. H. K. Lee, H. Kanamori, P. C. Jennings, and C. Kisslinger, 357-367, Academic Press, San Diego, California.
- Drouet S., S. Chevrot, F. Cotton, and A. Souriau (2008). Simultaneous Inversion of Source Spectra, Attenuation Parameters, and Site Responses: Application to the Data of the French Accelerometric Network, *Bull. Seism. Soc. Am.*, **98**, 198 - 219.
- Duval A.-M., S. Vidal, J.-P. Méneroud, A. Singer, F. De Santis, C. Ramos, G. Romero, R. Rodriguez, A. Pernia, N. Reyes and C. Griman (2001). Caracas, Venezuela, Site Effect Determination with Microtremors, *Pure and Applied Geophysics* **158**, 2513-2523
- Fäh D, Kind F, Giardini D. (2001). A theoretical investigation of average H/V ratios. *Geophys. J. Int.* **145**, 535-549
- Fäh, D., Kind, F. and Giardini, D. (2003). Inversion of local S-wave velocity structures from average H/V ratios, and their use for the estimation of site-effects. *J. Seismol.*, **7**, 449–467.
- Field, E., and Jacob, K. H. (1993). The theoretical response of sedimentary layers to ambient seismic noise. *Geophys. Res. Lett.*, **20**, 24, 2925-2928.
- Field, E. H., and Jacob, K. H. (1995). A comparison and test of various site response estimation techniques, including three that are non reference- site dependent. *Bull. Seism. Soc. Am.*, **86**, 991–1005.
- Field, E. H., Johnson, P. A., Beresnev, I. A., and Zeng, Y. (1997). Nonlinear ground-motion amplification by sediments during the 1994 Northridge earthquake. *Nature* **390**, 599-602.
- Foti, S., Parolai, S., Albarello, D., Picozzi, M. (2011). Application of surface-wave methods for seismic site characterization. *Survey in Geophysics*, DOI 10.1007/s10712-011-9134-2.
- Frosch, R. A., and Green, P. E. (1966). The concept of the large aperture seismic array, *Proc. R. Soc., London, Ser. A*, **290**, 368-384.
- Gerstoft P., Sabra, K. G., Roux, P., Kuperman, W. A. and Fehler, M. C. (2006). Green’s functions extraction and surface-wave tomography from microseisms in southern California. *Geophysics*, **71**, 23–32.
- Guiller B., Atakan, K., Chatelain, J.-L., Havskov, J., Ohrnberger, Cara, F., Duval, A.-M., Zacharopoulos, S., Teves-Costa, P., and the SESAME Team (2007). Influence of

- instruments on the H/V spectral ratios of ambient vibrations. *Bull. Earth. Eng.*, **6**, 3-31. doi:10.1007/s10518-007-9039-0
- Haghshenas, E., Bard, P.-Y., Theodulidis, N., and SESAME WP04 Team (2008). Empirical evaluation of microtremor H/V spectral ratio. *Bull. Earthquake Eng.*, doi: 10.1007/s10518-007-9058-x
- Halliday D. F., and Curtis, A., (2008a). Seismic interferometry, surface waves, and source distribution. *Geophys. J. Int.*, doi:10.1111/j.1365- 246X.2008.03918.x.
- Halliday D. F., and Curtis, A., (2008b). Seismic interferometry of scattered surface waves in attenuative media. *Geophys. J. Int.*, **178**, 419–446, doi: 10.1111/j.1365-246X.2009.04153.x.
- Halliday D.F., Curtis A., Robertsson J.O.A. and van Manen D.-J. (2007). Interferometric surface-wave isolation and removal, *Geophysics*, **72**, A67–A73.
- Halliday D.F., Curtis A. and Kragh E., (2008). Seismic surface waves in a suburban environment - active and passive interferometric methods. *Leading Edge*, **27**, 210–218.
- Harkrider, D. G, (1964). Surface waves in multilayered elastic media. I. Rayleigh and Love waves from buried sources in a multilayered elastic half-space. *Bull. Seism. Soc. Am.*, **54**, 627-679.
- Horike, M. (1985). Inversion of phase velocity of long period microtremors to the S-wave velocity structure down to the basement in urbanized areas. *J. Phys. Earth* **33**, 59-96.
- Hough S.E., Seeber L., Rovelli A., Malagnini L., DeCesare A., Selvaggi G., and Lerner-Lam A. (1992). Ambient noise and weak motion excitation of sediment resonances: results from the Tiber Valley, Italy. *Bull. Seism. Soc. Am.*, **82**, 1186-1205.
- Ibs von Seht, M., and Wohlenberger, R. (1999). Microtremor measurements used to map thickness of soft soil sediments. *Bull. Seism. Soc. Am.*, **89**, 250-259.
- Káráson, H., and van der Hilst, R.D. (2001). Tomographic imaging of the lowermost mantle with differential times of refracted and diffracted core phases (PKP, Pdiff), *J. Geophys. Res.*, **106**, 6569-6587.
- Kind, F., Fäh, D., Giardini, D. (2005). Array measurements of S-wave velocities from ambient vibrations, *Geophysical Journal International*, **160**. 114-126.
- Köhler, A., Ohrnberger, M., Scherbaum, F., Wathelet, M., and Cornou, C. (2007). Assessing the reliability of the modified three-component spatial autocorrelation technique. *Geophys J. Int.*, **168**, 779-796; doi:10.1111/j.1365-246X.2006.03253.x
- Krüger, F., Baumann, M., Scherbaum, F., and Weber, M. (2001). Mid mantle scatterers near the Mariana slab detected with a double array method. *Geophys. Res. Lett.*, **28**, 667-670.
- Kværna, T. (1989). On exploitation of small-aperture NORESS type arrays for enhanced P-wave detectability. *Bull. Seism. Soc. Am.*, **79**, 888-900.
- Lachet, C., Hatzfeld, D., Bard, P.-Y., Theodulidis, N., Papaioannou, C., and Savvaidis, A. (1996). Site effects and microzonation in the city of Thessaloniki (Greece): comparison of different approaches. *Bull. Seism. Soc. Am.*, **86**, no. 6, 1692–1703.
- Lacoss, R. T., Kelly, E.J., and Toksöz, M.N. (1969). Estimation of seismic noise structure using array. *Geophysics*, **29**, 21-38.
- Langston, C. A. (1979). Structure under Mount Rainier, Washington, inferred from teleseismic body waves, *J. Geophys. Res.* **84**, 4749–4762.
- Lermo, J. F., and Chavez-Garcia, F. J. (1993). Site effect evaluation using spectral ratios with only one station. *Bull. Seism. Soc. Am.* **83**, 1574-1594.
- Lermo, J. F. and Chavez-Garcia, F. J., (1994). Are microtremors useful in site response evaluation?" *Bull. Seism. Soc. Am.*, **84**, 135, 1364.
- Lin, F.-C., Ritzwoller, M.H., Townend, J., Bannister, S. and Savage M.K. (2007). Ambient noise Rayleigh wave tomography of New Zealand. *Geophys. J. Int.*, **170**, 649–666, doi:10.1111/j.1365–246X.2007.03414x.



- Lobkins, O. I., and R. L. Weaver (2001). On the emergence of the Green's function in the correlation of a diffuse field. *Journal of Acoustical Society of America*, **110**, 3011-3017.
- Louie, J. N. (2001). Faster, better: shear-wave velocity to 100 meters depth from refraction microtremor arrays. *Bull. Seismol. Soc. Amer.*, **91**, 2, 347-364.
- Lunedei, E., and Albarello, D. (2009). On the seismic noise wave field in a weakly dissipative layered Earth. *Geophys. J. Int.*;doi: 10.1111/j.1365-246X.2008.04062.x
- Malagnini, L., Rovelli, A., Hough, S.E., and Seeber, L. (1993), Site amplification estimates in the Garigliano Valley, Central Italy, based on dense array measurements of ambient noise. *Bull. Seism. Soc. Am.*, **83**, 1744-1755.
- Mehta, K., Snieder, R., and Grazier, V. (2007a). Extraction of near-surface properties for a lossy layered medium using the propagator matrix. *Geophys. J. Int.*, **169**, 2171-280.
- Mehta, K., Snieder, R., and Grazier, V. (2007b). Downhole receiver function: a case study. *Bull. Seism. Soc. Am.*, **97**, 1396-1403.
- Mogi H., S. M. Shrestha, H. Kawakami, and Shinya Okamura (2010). Nonlinear Soil Behavior Observed at Vertical Array in the Kashiwazaki-Kariwa Nuclear Power Plant during the 2007 Niigata-ken Chuetsu-oki Earthquake. *Bull. Seism. Soc. Am.*, **100**, 762-775.
- Moya A., and Irikura, K. (2003). Estimation of Site Effects and Q Factor Using a Reference Event. *Bull. Seism. Soc. Am.*, **93**: 1730 - 1745.
- Mucciarelli, M. (1998). Reliability and applicability of Nakamura's technique using microtremors: an experimental approach. *J. Earthq. Engrg.*, **2**, 625-638.
- Nakamura, Y. (1989). A method for dynamic characteristics estimation of subsurface using microtremor on the ground surface. *Q. Rep. Railway Tech. Res. Inst.*, **30**, 25-33.
- Nogoshi, M., and Igarashi, T. (1970). On the propagation characteristics of microtremors, *J. Seism. Soc. Japan*, **23**, 264-280 (in Japanese with English abstract).
- Nogoshi, M., and Igarashi, T. (1971). On the amplitude characteristics of microtremor (part 2). *J. Seism. Soc. Japan*, **24**, 26-40 (Japanese with English abstract).
- Ohori, M., Nobata, A., and Wakamatsu, K. (2002). A comparison of ESAC and FK methods of estimating phase velocity using arbitrarily shaped microtremor arrays. *Bull. Seism. Soc. Am.*, **92**, (6), 2323-2332.
- Okada, H. (2003). The microtremor survey method. *Geophysical monograph series*, **12**, American Geophysical Union, Washington.
- Oth, A., Bindi, D., Parolai, S., and Wenzel, F. (2008). S-wave attenuation characteristics beneath the Vrancea region in Romania: New insights from the inversion of ground-motion spectra. *Bull. Seism. Soc. Am.*, **98**, 2482-2497.
- Oth, A., Parolai, S., Bindi, D., Wenzel, F. (2009). Source Spectra and Site Response from S Waves of Intermediate-Depth Vrancea, Romania, Earthquakes. *Bull. Seism. Soc. Am.*, **99**, 235-254.
- Park C. B., Miller R. D., and Xia J. (1999). Multichannel analysis of surface waves. *Geophysics*, **64**, 800-808.
- Parolai, S., and Richwalski, S. M. (2004). The importance of converted waves in comparing H/V and RSM site responses. *Bull. Seism. Soc. Am.*, **94**, 304-313.
- Parolai, S., and Galiana-Merino, J. J. (2006): Effect of transient seismic noise on estimates of H/V spectral ratios. *Bull. Seism. Soc. Am.*, **96**, 1, 228-236.
- Parolai, S., Bindi, D., and Augliera, P. (2000). Application of the Generalized Inversion Technique (GIT) to a microzonation study: numerical simulations and comparison with different site-estimation techniques. *Bull. Seism. Soc. Am.*, **90**, no. 2, 286-297.
- Parolai, S., Trojani, L., Monachesi, G., Frapiccini, M., Cattaneo, M., and Augliera, P. (2001). Hypocenter location accuracy and seismicity distribution in the Central Apennines (Italy). *J. Seism.*, **5**, 243-261.

- Parolai, S., Bindi, D., and Troiani, L. (2001a). Site response for the RSM seismic network and source parameters in the Central Apennines (Italy). *Pure Appl. Geophys.*, **158**, 695–715.
- Parolai S, Bormann, P, and Milkereit, C. (2001b). Assessment of the natural frequency of the sedimentary cover in the Cologne area (Germany) using noise measurements. *J. Earthq. Engrg.*, **5**, 541-564.
- Parolai S, Bormann, P, Milkereit, C. (2002). New relationships between  $V_s$ , thickness of sediments, and resonance frequency calculated by the H/V ratio of seismic noise for the Cologne area (Germany). *Bull. Seism. Soc. Am.*, **92**, 2521-2527.
- Parolai S, Richwalski, S. M, Milkereit, C, and Bormann, P. (2004). Assessment of the stability of H/V spectral ratio from ambient noise and comparison with earthquake data in the Cologne area (Germany). *Tectonophysics*, **390**, 57-73.
- Parolai, S., Picozzi, M., Richwalski, S. M., and Milkereit, C. (2005). Joint inversion of phase velocity dispersion and H/V ratio curves from seismic noise recordings using a genetic algorithm, considering higher modes. *Geoph. Res. Lett.*, **32**, doi: 10.1029/2004GL 021115
- Parolai, S., Richwalski, S. M., Milkereit, C., and Fäh, D. (2006). S-wave velocity profiles for earthquake engineering purposes for the Cologne area (Germany). *Bull. Earth. Engineering*, **4**, 65-94.
- Parolai, S., Grünthal, G., and Wahlström, R., (2007). Site-specific response spectra from the combination of microzonation with probabilistic seismic hazard assessment - An example for the Cologne (Germany) area. *Soil Dynamics and Earthquake Engineering*, **27**, 1, 49-59.
- Parolai, S., Bindi, D., Durukal, E., Grosser, H., and Milkereit, C. (2007). Source parameter and seismic moment-magnitude scaling for northwestern Turkey. *Bull. Seim. Soc. Am.*, **97**(2), 655-660.
- Parolai, S., Ansal, A., Kurtulus, A., Strollo, A., Wang, R., and Zschau, J. (2009). The Ataköy vertical array (Turkey): insights into seismic wave propagation in the shallow-most crustal layers by waveform deconvolution. *Geophys. J. Int*; doi:10.1111/j.1365-246X.2009.04257. x
- Parolai S., Bindi, D., Ansal, A., Kurtulus, A., Strollo, A. and Zschau, J. (2010a). Determination of shallow S-wave attenuation by down-hole waveform deconvolution: a case study in Istanbul (Turkey). *Geophys. J. Int.*, doi: 10.1111/j.1365-246X.2010.04567.x
- Parolai S., Orunbaev, S., Bindi, D., Strollo, A., Usupaev, S., Picozzi, M., Di Giacomo, D., Augliera, P., D'Alema, E., Milkereit, C., Moldobekov, B., and Zschau, J. (2010b). Site effect assessment in Bishkek (Kyrgyzstan) using earthquake and noise recording data. *Bull. Seism. Soc. Am.*, **100**, 3068-3082.
- Picozzi, M. (2006). Joint inversion of phase velocity dispersion and H/V ratio curves from seismic noise recordings. *PhD. Thesys, University of Siena*, 170 pp.
- Picozzi, M., Parolai, S., and Albarello, D. (2005a). Statistical analysis of noise horizontal to-vertical spectral ratios (HVSr). *Bull. Seism. Soc. Am.*, **95**, 1779, 1786.
- Picozzi, M., Parolai, S., and Richwalski, S. M. (2005b). Joint inversion of H/V ratios and dispersion curves from seismic noise: Estimating the S-wave velocity of bedrock. *Geoph. Res. Lett.*, **32**, doi: 10.1029/2005GL022878.
- Picozzi, M., and Albarello, D. (2007). Combining genetic and linearized algorithms for a two-step joint inversion of Rayleigh wave dispersion and H/V spectral ratio curves. *Geophys. J. Int.*, **169**, 189–200.
- Picozzi, M., Parolai, S., Bindi, D., Strollo, A. (2008). Characterization of shallow geology by high-frequency seismic noise tomography. *Geophysical Journal International*, **176**, 1, 164-174.
- Picozzi, M., Strollo, A., Parolai, S., Durukal, E., Özel, O., Karabulut, S., Zschau, J., Erdik, M. (2009). Site characterization by seismic noise in Istanbul, Turkey. *Soil Dynamics and Earthquake Engineering*, **29**, 3, 469-482.

- Picozzi, M., Parolai, S., and Bindi, D. (2010a): Deblurring of frequency-wavenumber images from small-scale seismic arrays. *Geophys. J. Internat.*; doi: [10.1111/j.1365-246X.2009.04471.x](https://doi.org/10.1111/j.1365-246X.2009.04471.x)
- Picozzi, M., Milkereit, C., Parolai, S., Jaekel, K.-H., Veit, I., Fischer, J., Zschau, J. (2010b). GFZ Wireless Seismic Array (GFZ-WISE), a wireless mesh network of seismic sensors: New perspectives for seismic noise Array investigations and site monitoring. *Sensors*, **10**(4), 3280-3304.
- Pilz, M., Parolai, S., Leyton, F., Campos, J., and Zschau, J. (2009). A comparison of site response techniques using earthquake data and ambient seismic noise analysis in the large urban areas of Santiago de Chile. *Geophys. J. Internat.*, **178**, 2, 713-728.
- Pilz, M., Parolai, S., Picozzi, M., Wang, R., Leyton, F., Campos, J., and Zschau, J. (2010). Shear wave velocity model of the Santiago de Chile basin derived from ambient noise measurements: a comparison of proxies for seismic site conditions and amplification. *Geophys. J. Internat.*, **182**, 355-367.
- Riepl, J., Bard, P.-Y., Hatzfeld, D., Papaioannou, C., and Nechtschein, S. (1998). Detailed evaluation of site response estimation methods across and along the sedimentary valley of Volvi (EURO-SEISTEST). *Bull. Seism. Soc. Am.*, **88**, no. 2, 488-502.
- Ritter, J. R. R., Jordan, M., Christensen, U., and Achauer, U. (2001). A mantle plume below the Eifel volcanic fields, Germany. *Earth Planet. Sci. Lett.*, **186**, 7-14.
- Rost, S., and Thomas, C. (2002). Array Seismology: Methods and applications. *Rev. Geophys.*, **40** (3), 1008, doi: [10.1029/20000RG000100](https://doi.org/10.1029/20000RG000100).
- Sabra, K.G., Gerstoft, P., Roux, P., Kuperman, W.A., and Fehler, M.C. (2005). Surface wave tomography from microseisms in Southern California. *Geophys. Res. Lett.*, **32**, L14311, doi:10.1029/2005GL023155.
- Safak, E. (1997). Models and methods to characterize site amplification from a pair of records. *Earthquake Spectra*, EERI, **13**, 97-129.
- Safak, E. (2001). Local site effects and dynamic soil behaviour. *Soil Dynamics and Earthquake Engineering*, Elsevier Science Ltd., Vol. 21, pp.453-458.
- Scherbaum, F., Hinzen, K.G., and Ohrnberger, M. (2003). Determination of shallow shear wave velocity profiles in the Cologne, Germany, area using ambient vibrations. *Geophys. J. Int.*, **152**, 597-612.
- Schuster, G. T. (2001). Theory of daylight/interferometric imaging: tutorial. In: *Proceedings of 63rd Meeting, European Association of Geoscientists and Engineers*, Session: A32 (extended Abstracts).
- Schuster, G. T., Yu, J., Sheng, J., and Rickett, J. (2004). Interferometric/daylight seismic imaging. *Geophys. J. Int.*, **157**, 838-852.
- SESAME group (2003). Final report on measurements guidelines, LGIT Grenoble, CETE Nice, WP02, H/V technique: experimental conditions, [http://sesame-fp5.obs.ujf-grenoble.fr/Delivrables/D08-02\\_Texte.pdf](http://sesame-fp5.obs.ujf-grenoble.fr/Delivrables/D08-02_Texte.pdf)
- Shapiro, N. M., and Campillo, M. (2004). Emergence of broadband Rayleigh waves from correlations of ambient seismic noise. *Geophys. Res. Lett.*, **31**, L07614, doi:10.1029/2004GL019491.
- Shapiro, N. M., Campillo, M., Stehly, L., and Ritzwoller, M. (2005). High resolution surface wave tomography from ambient seismic noise. *Science*, **307**, 1615-1618
- Snieder, R. (2004). Extracting the Green's function from the correlation of coda waves: a derivation based on stationary phase. *Phys. Rev. E*, **69**; doi:10.1103/PhysRevE.69.046610.
- Snieder, R. (2006). The theory of coda wave interferometry. *Pure Appl. Geophys.*, **163**, 455-473.
- Snieder, R., Sheiman, J., and Calvert, R. (2006). Equivalence of the virtual-source method and wave-field deconvolution in seismic interferometry. *Physical Review E*, **73**, 066620.

- Steidl, J. H., Tumarkin, A. G., and Archuleta, R. (1996). What is a reference site? *Bull. Seism. Soc. Am.*, **86**, no. 6, 1733–1748.
- Stokoe K. H. II, Wright, S. G., Bay, J.A., Roesset, J.M.(1994). Characterization of geotechnical sites by SASW method. In: R.D. Woods (Ed.), *Geophysical Characterization of Sites*, 15-25.
- Strollo A., D. Bindi,, S. Parolai,, K.-H. Jäckel (2008a). On the suitability of 1 s geophone for ambient noise measurements in the 0.1–20Hz frequency range: experimental outcomes, *Bulletin of Earthquake Engineering*, **6**, 1, 141-147.
- Strollo, A., Parolai, S., Jäckel, K.-H., Marzorati., S., and Bindi, D. (2008b). Suitability of short-period sensors for retrieving reliable H/V peaks for frequencies less than 1 Hz. *Bull. Seism. Soc. Am.*, **98**, 2, 671-681.
- Tada, T., Cho, I., Shinozaki, Y. (2006). A two-radius circular array method: inferring phase velocity of Love waves using microtremor records. *Geophys. Res. Lett.*, **33**, L10303; doi:10.1029/2006GL025722.
- Trampert, J., Cara, M., and Frogneux, M. (1993). SH propagator matrix and Qs estimates from borehole- and surface-recorded earthquake data. *Geophys. J. Int.*, **112**, 290-299.
- Triantafyllidis, P., Hatzidimitriou, P. M., Theodulidis, M., Suhadolc, P., Papazachos, C., Raptakis, D., and Lontzetidis, K. (1999). Site effects in the city of Thessaloniki (Greece) estimated from acceleration data and 1D local soil profiles. *Bull. Seism. Soc. Am.*, **89**, 521–537.
- van Vossen, R., Trampert, J., and Curtis, A. (2004). Propagator and wave-equation inversion for near-receiver material properties. *Geophys. J. Int.*, **157**, 796-812.
- van Vossen, R., Curtis, A., and Trampert, J. (2005). Subsonic near-surface P-velocity and low S-velocity observation using propagator inversion. *Geophysics*, **70**, R15-R23.
- von Kamp, H. (1986). Geologische Karte von Nordrhein-Westfalen 1:100.000. *Geologisches Landesamt Nordrhein-Westfalen*, Germany.
- Wapenaar, K. (2004). Retrieving the elastodynamic Green's function of an arbitrary inhomogeneous medium by cross correlation. *Phys. Rev. Lett.*, **93**, 25 4301–1-4.
- Wapenaar, K., and Fokkema, J. (2006). Green's function representations for seismic interferometry. *Geophyscis*, **71**, SI33–SI44.
- Weaver, R., and Lobkis, O. (2001). On the emergence of the Green's function in the correlations of a diffuse field. *J. acoust. Soc. Am.*, **109**, 2410.
- Weaver, R. L., and Lobkis, O. I. (2004). Diffuse fields in open systems and the emergence of the Green's function. *J. acoust. Soc. Am.*, **116**, 2731–2734.
- Whiteway, F.E. (1966). The use of arrays for earthquake seismology. *Proc. R. Soc. London, Ser. A*, **290**, 328-348.
- Yamada, M., Mori, J., and Ohmi, S. (2010). Temporal Changes of Subsurface Velocities during Strong Shaking as Seen from Seismic Interferometry. *J. Geophys. Res.*, **115**, B03302, doi:10.1029/2009JB006567.
- Yamanaka, H., Takemura, M., Ishida, H., and Niwa, M. (1994). Characteristics of long-period microtremors and their applicability in the exploration of deep sedimentary layers. *Bull. Seism. Soc. Am.*, **84**, 1831–1841.
- Yang, Y., Ritzwoller, M. H., Levshin, A.L., and Shapiro, N. M. (2007). Ambient noise Rayleigh wave tomography across Europe. *Geophys. J. Int.*, **168**, 259–274, doi:10.1111/j.1365–246X.2006.03203.x
- Yao, H., Van Der Hilst, R. D., and De Hoop, M.V. (2006). Surface-wave array tomography in SE Tibet from ambient seismic noise and two station analysis: I - phase velocity maps. *Geophys. J. Int.*, **166**, 732– 744.
- Yaramanci, U., Lange, G., and Hertrich, M. (2002). Aquifer characterisation using Surface NMR jointly with other geophysical techniques at the Nauen/Berlin test site *J. Appl. Geophys.*, **50**, 47–65.

- Zhang, S. H., Chan, L. S., and Xia, J. (2004). The selection of field acquisition parameters for dispersion images from multichannel surface wave data. *Pure Appl. Geophys.*, **161**, 185–201
- Zywicki, D. J. (1999). Advanced signal processing methods applied to engineering analysis of seismic surface waves. PhD thesis at Georgia Institute of Technology

| Topic   | Tapering of windowed time series                                                                                                                                                                   |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Authors | Marco Pilz and Stefano Parolai, Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Helmholtzstr. 7, 14473 Potsdam, <a href="mailto:pilz@gfz-potsdam.de">pilz@gfz-potsdam.de</a> |
| Version | February 2012; DOI: 10.2312/GFZ.NMSOP-2_IS_14.1                                                                                                                                                    |

As described in Chapter 14, a largely used method of seismic data analysis, which is often applied to study geophysical processes, consists in the spectral analysis of seismic signals after having carried out a Fourier transform of the time series.

However, for the analysis of noise data the recordings of some tens of minutes are usually subdivided into much shorter time windows (usually 30 s or 60 s). For earthquake recordings the spectral estimation of specific phases within a seismogram, particular those recorded at local and regional distances, can be difficult due to the difficulties in isolating a particular phase. On the contrary, mathematical principles require infinite long time series for performing a Fourier transform and, hence, such windowing will cause the Fourier transform to develop non-zero values predominantly at lower frequencies (commonly called spectral leakage, i.e. some frequencies tend to leak to other frequencies).

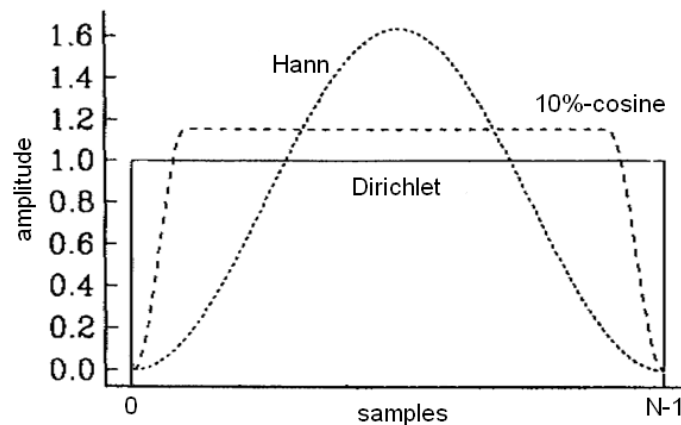
Therefore, it is standard practice to multiply the data windows by a taper before performing a Fourier transform. The taper consist in a function smoothly decaying to zero near the ends of each window, aimed at minimizing the effect of the discontinuity between the beginning and end of the time series. Although spectral leakage cannot be prevented it can be significantly reduced by changing the shape of the taper function in a way to minimize strong discontinuities close to the window edges.

Different types of tapers are shown in Figure 1. In seismic data analysis cosine tapers are often used, being both effective and easy to calculate, though bell-shaped and triangle functions are sometimes applied, too. In a mathematical form, the cosine taper can be written as

$$c(t) = \begin{cases} \frac{1}{2} \left( 1 - \cos \frac{\pi}{a} t \right) & \text{for } 0 \leq t \leq a \\ 1 & \text{for } a \leq t \leq (1-a) \\ \frac{1}{2} \left( 1 - \cos \frac{\pi}{a} (1-t) \right) & \text{for } (1-a) \leq t \leq 1 \end{cases}$$

with time  $t$  and taper ratio  $a$ . The cosine window represents an attempt to smoothly set the data to zero at the boundaries while not significantly reducing the level of the windowed transform.

Such tapering will lead to the following effects: It reduces the leakage of spectral power from a spectral peak to frequencies far away and it coarsens the spectral resolution by a factor  $1 / (1 - a)$  for the above cosine tapers.



**Figure 1** Comparison between the Dirichlet sharp-edged rectangle window function with two common taper functions.

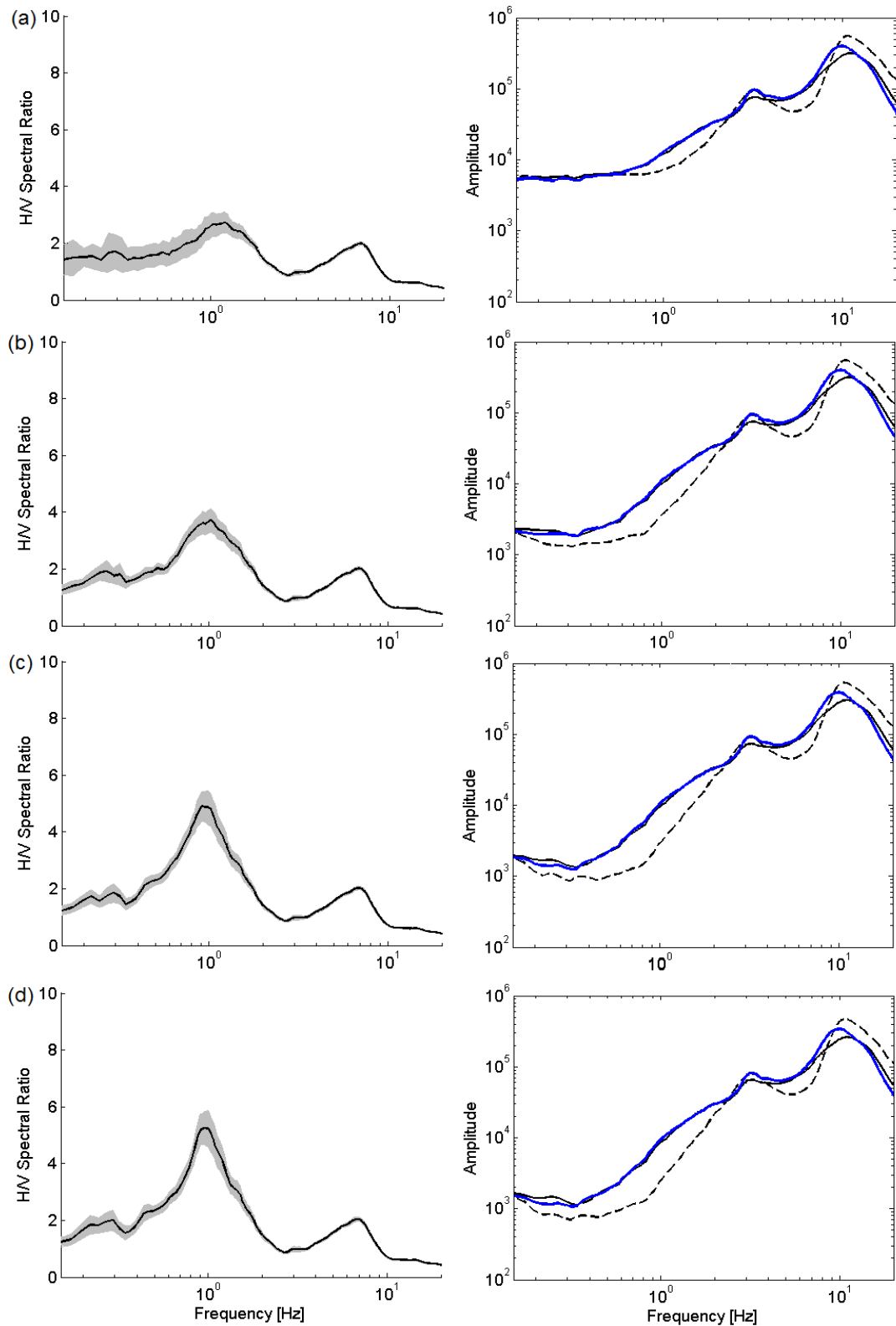
The result of applying the cosine taper family of windows with a length of 30 s is presented in Figure 2. As the taper ratio  $a$  is decreased, or equivalently, as the shape of the window becomes increasingly rectangular, the bandwidth of the Fourier transform decreases. Therefore, a decreased taper ratio will increase the amount of leakage. As can be seen in Figure 2 (a), spectra from the untapered time series show a significant overestimation in spectral energy for low frequencies (leakage), resulting in a much lower H/V amplitude. Below 0.5 Hz the spectra for all three components show the same level of amplitude, limiting the effective bandwidth of the spectra when no tapering is applied.

Conversely, as the taper ratio is increased or the window shape becomes more like that of a 'cosine bell', the bandwidth increases, reducing the amount of leakage and giving rise to an increase in the resolution of closely-spaced low-frequency spectral components. The amplitudes of the Fourier spectra are significantly reduced, resulting in a clear H/V peak and a decrease of the spectral components towards lower frequencies.

Therefore, if the real spectrum contains a strong low-frequency component, cutting the time series into short time windows without tapering may strongly distort the observed spectrum. However, one should keep in mind that too much tapering may seriously impair the statistical efficiency of the estimates. In general, it is difficult to give precise recommendations which tapers to use in any specific situation (see, e.g., the discussion in Bingham et al., 1967). Yet, as these authors could show, it is recommendable to reduce the leakage more efficiently by increasing the length  $N$  of the time window and at the same time decreasing the taper ratio  $a$ , such that the length ( $Na$ ) of the cosine half-bells is kept constant.

Usually, a value of  $a = 5\%$  for window lengths of 30 s or 60 s is good enough for frequencies down to 0.2 Hz, whereas a further increase of the taper ratio  $a$  up to  $a = 20\%$  does not imply further changes in the spectra (see Figure 2).

For further readings, we refer to Harris (1978), Oppenheim et al. (1999) and Park et al. (1987).



**Figure 2** Effect of different cosine tapering for the same data set using a window length of 30 s. **Left column:** Thick lines show medium H/V spectral ratio plus/minus one standard deviation using (a) no tapering or cosine tapering with (b)  $a = 1\%$ , (c)  $a = 5\%$ , (d)  $a = 20\%$ . **Right column:** The corresponding Fourier spectra for both horizontal (continuous black and blue lines) and vertical (dashed line) components.



## References

- Bingham, C., Godfrey, M. D., and Tukey, J. W. (1967). Modern techniques of power spectrum estimation. *IEEE Trans. Audio Electroacoustic*, AU-15, 55-66
- Harris, F. (1978). On the use of windows for harmonic analysis with the discrete Fourier transform. *Proc. Instr. Electr. Eng.*, **66**, 51-83
- Oppenheim, A. V., Schafer, R. W., and Buck, J. A. (1999). *Discrete time signal processing*. Prentice Hall International Editions, New York
- Park, J., Lindberg, C. R., and Vernon, F. L. (1987). Multitaper spectral analysis of high-frequency seismograms. *J. Geophys. Res.*, **92**, 12675-12684

# Chapter 15

## CTBTO: Goals, Networks, Data Analysis and Data Availability

(Version: January 2012; DOI: 10.2312/GFZ.NMSOP-2\_ch15)

John Coyne (USA)<sup>1)</sup>, Dmitry Bobrov (Russia)<sup>1)</sup>, Peter Bormann (Germany)<sup>2)</sup>,  
Emerenciana Duran (Philippines)<sup>1)</sup>, Patrick Grenard (France)<sup>1)</sup>,  
Georgios Haralabus (Greece)<sup>1)</sup>, Ivan Kitov (Russia)<sup>1)</sup> and Yuri Starovoit  
(Russia)<sup>1)</sup>

<sup>1)</sup> Comprehensive Nuclear Test Ban Treaty Organization, International Data Centre Division,  
Vienna, Austria; contact via Fax: +43 1 26030 5923 or E-Mail: Dmitry.Bobrov@ctbto.org

<sup>2)</sup> Formerly: GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany;  
E-mail: pb65@gmx.net

|                                            | page |
|--------------------------------------------|------|
| <b>15.1 Introduction</b>                   | 2    |
| 15.1.1 The Treaty                          | 2    |
| 15.1.2 Nuclear explosions                  | 3    |
| <b>15.2 CTBTO networks</b>                 | 5    |
| 15.2.1 Seismic stations                    | 6    |
| 15.2.1.1 Three-component station           | 6    |
| 15.2.1.2 Array station                     | 6    |
| 15.2.2 Hydroacoustic station               | 8    |
| 15.2.2.1 Hydrophone station                | 8    |
| 15.2.2.2 T-phase station                   | 9    |
| 15.2.3 Infrasound station                  | 11   |
| 15.2.4 Radionuclide particulate            | 12   |
| 15.2.5 Radionuclide noble gas              | 14   |
| 15.2.6 Radionuclide laboratories           | 15   |
| <b>15.3 Data acquisition</b>               | 16   |
| 15.3.1 Introduction                        | 16   |
| 15.3.2 Security concerns                   | 16   |
| 15.3.3 Continuous data stations            | 17   |
| 15.3.4 Segmented data stations             | 18   |
| 15.3.5 Radionuclide stations               | 19   |
| 15.3.6 Command and control                 | 19   |
| 15.3.7 Global Communication Infrastructure | 19   |
| <b>15.4 Waveform data processing</b>       | 20   |
| 15.4.1 Introduction                        | 20   |
| 15.4.2 Seismic processing                  | 21   |
| 15.4.2.1 DFX                               | 21   |
| <i>Data quality</i>                        | 21   |
| <i>Improve and estimate SNR</i>            | 22   |
| <i>Find detection</i>                      | 24   |

|                                                              |    |
|--------------------------------------------------------------|----|
| <i>Onset time refinement</i>                                 | 24 |
| <i>Measure amplitude and period and determine magnitudes</i> |    |
| <i>Mb, Ms, and ML</i>                                        | 25 |
| <i>Estimate azimuth and slowness</i>                         | 28 |
| 15.4.2.2 StaPro                                              | 29 |
| <i>Determine initial wave type</i>                           | 29 |
| <i>Group detection</i>                                       | 30 |
| <i>Assign initial phase name</i>                             | 31 |
| <i>Locate single station events</i>                          | 31 |
| 15.4.3 Event location                                        | 32 |
| 15.4.3.1 Input data                                          | 32 |
| <i>Use of time, azimuth, and slowness observations</i>       | 33 |
| <i>Establishing the initial location</i>                     | 33 |
| <i>Predicting travel-time, azimuth and slowness values</i>   | 34 |
| <i>Source-Specific Station Corrections (SSSC)</i>            | 34 |
| <i>Predicting seismic slowness and azimuth values</i>        | 34 |
| <i>Slowness-Azimuth Station Corrections (SASC)</i>           | 34 |
| <i>Hydroacoustic measurements</i>                            | 34 |
| <i>Hydroacoustic blockage</i>                                | 35 |
| <i>Infrasound measurements</i>                               | 35 |
| 15.4.3.2 Non-linear least squares inversion                  | 36 |
| 15.4.3.3 Evaluating the solution stability                   | 36 |
| <i>Stability</i>                                             | 36 |
| <i>Divergence test</i>                                       | 36 |
| <i>Convergence test</i>                                      | 37 |
| <i>Maximum iteration test</i>                                | 37 |
| <i>Updating hypocenters</i>                                  | 37 |
| <i>Estimating errors</i>                                     | 37 |
| 15.4.4 Event definition criteria                             | 37 |
| <b>15.5 Data availability</b>                                | 38 |
| <b>Acknowledgments</b>                                       | 38 |
| <b>Disclaimer</b>                                            | 38 |
| <b>Recommended overview readings</b>                         | 38 |
| <b>References</b>                                            | 39 |

## 15.1 Introduction

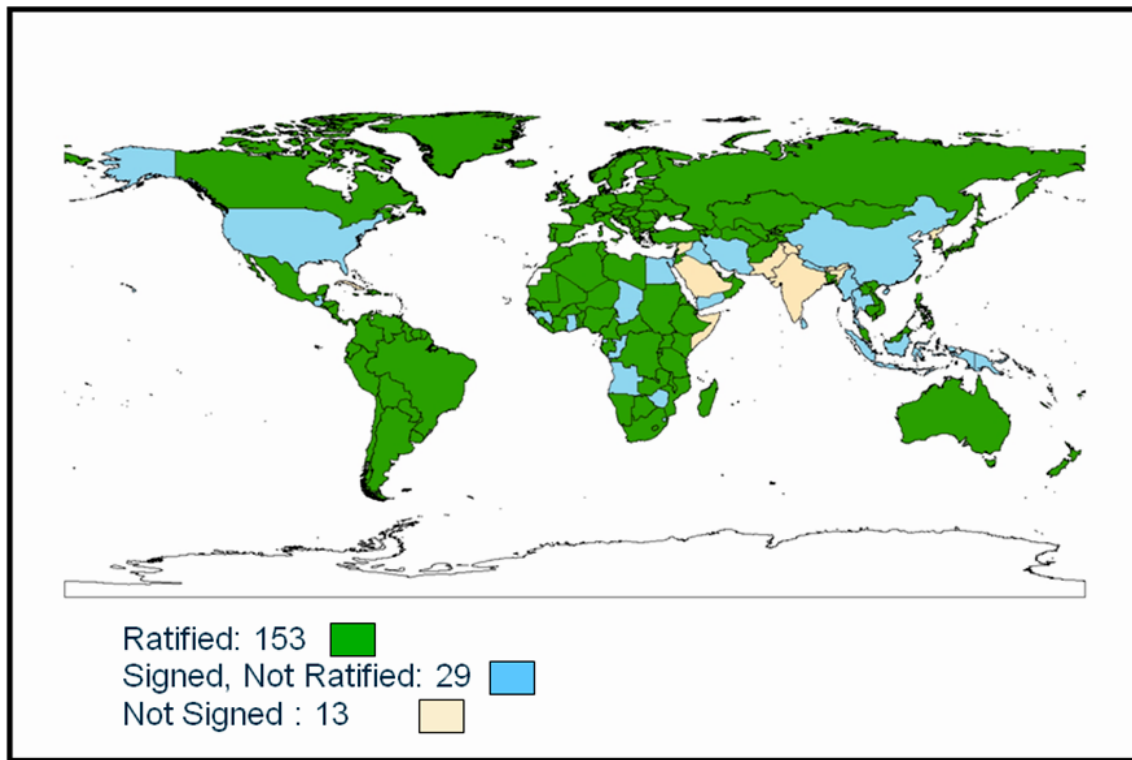
### 15.1.1 The Treaty

The Comprehensive Nuclear-Test-Ban Treaty (CTBT) was opened for signature on 15 September 1996. The CTBT prohibits nuclear test explosions in all environments: underground, in the oceans, and in the atmosphere. As of 1 April 2011, the CTBT has been signed by 182 States, and has been ratified by 153 States. The CTBT will enter into force once it has been ratified by 40 States specified in Annex 2 of the treaty, the so called “Annex 2 States”. As of 1 April 2011, 35 of the Annex 2 States have ratified the CTBT.

The Comprehensive Nuclear-Test-Ban Treaty Organization Preparatory Technical Secretariat (CTBTO/PTS) was established in 1997 to prepare the systems for monitoring compliance with the CTBT once the Treaty enters into force. During this preparatory phase a great

amount of work is being done to establish the International Monitoring System (IMS), the Global Communication Infrastructure (GCI), the International Data Centre (IDC) in Vienna, Austria and the On-Site-Inspection (OSI) capabilities.

Fig. 15.1 shows a map with the countries who have as of April 2011 ratified, not ratified or not yet signed the Treaty.



**Fig. 15.1** Ratification status of the CTBT as of 1 April 2011.

### 15.1.2 Nuclear explosions

A nuclear explosion unleashes a tremendous amount of energy. Part of this energy can travel through the Earth, atmosphere, and oceans as seismic, infrasound, and hydroacoustic waves, respectively. The amount of energy travelling through each of these media is a function of where the explosion takes place, i.e., whether the explosion is underground, in the atmosphere, or underwater. The energy can also transfer from one media to another, depending on the site of the explosion. For example, an explosion on a boat will generate both infrasound and hydroacoustic signals, and the hydroacoustic signals can be converted into seismic signals.

A nuclear explosion also generates tremendous amounts of radioactive particles and gases, known as radionuclides. Observing these nuclides can provide unequivocal evidence that a nuclear explosion has occurred, since many of the radionuclides that can be detected and identified are fission products and can only arise from nuclear fission. Moreover, it is possible to distinguish between fission products from a nuclear explosion and those arising from atmospheric releases from civil nuclear power and reprocessing plants. The primary role of radionuclide monitoring is to provide unambiguous evidence of a nuclear explosion through

the detection and identification of fission products. These fission products can be sampled from ambient air as particulates or noble gases.

In addition to nuclear explosions, there are many other signals, which can be generated by other natural or man-made sources in the environments of interest. These include earthquakes, other non-nuclear explosions (e.g., mining explosions, quarries, accidental explosions), volcanic activity, airgun surveys, lightening, microbaroms, and biological sources (e.g., whales, elephants, etc.). Fortunately, not all of these sources generate signals similar to those generated by nuclear explosions.

Signals which are not of interest for CTBT purposes include repetitive sources (e.g., air gun surveys), very low energy events (e.g., lightening), signals outside the frequency range of interest (e.g., whale noise), signals outside the velocity range of interest, quickly moving objects (e.g., aircraft), and long, continuous signals (e.g., microbaroms). These distinguishing characteristics can be used to identify these signals as being noise during automatic processing, and are not considered further.

While signals from some sources can be discarded as noise, there is a wide range of sources which generate signals which must be processed, analyzed, and reviewed. The location and time of these sources are subsequently reported by the IDC in its bulletins. Potential sources in the bulletins include earthquakes, non-nuclear explosions, volcanoes, and meteorites. These sources may have similar characteristics of a nuclear explosion in terms of size, impulsive energy release, or frequency content.

Due to the nature of nuclear explosions and the environments in which they may take place, the IMS network was designed to have seismic, infrasound, hydroacoustic, and radionuclide sensors. The notional concept of monitoring the earth for a 1 kiloton nuclear explosion was used during the negotiations of the Treaty for illustrative purposes, and for comparisons between various proposed designs. This led to the design of the network in terms of number of stations, passband, and sensitivity. The provision for an onsite inspection following a suspicious event and its associated timelines as well as security concerns contributed to the timeliness requirements concerning IMS data acquisition and the issuance of IDC products.

While a nuclear weapon can be detonated in any environment, a clandestine test is most likely to take place underground. This is due to the fact that a nuclear explosion in the atmosphere would generate significant amounts of infrasound energy and copious quantities of radionuclides which would clearly indicate that a nuclear explosion took place. An underwater test would generate clear hydroacoustic signals as well as a release of radionuclides which would indicate a nuclear explosion.

Due to these reasons, environmental concerns, as well as to the Partial Test Ban Treaty (which prohibits nuclear weapon tests in the atmosphere, in outer space and under water), all nuclear explosions since 1980 have been underground. Since underground nuclear explosions generate significant seismic energy, seismic monitoring is particularly well suited for locating an underground nuclear explosion and determining its magnitude. The conversion of seismic magnitude to explosive yield is a complicated process, which relies on many different factors including geology and hydrology (e.g., Ringdahl, et al., 1992; Richards and Wu, 2011).

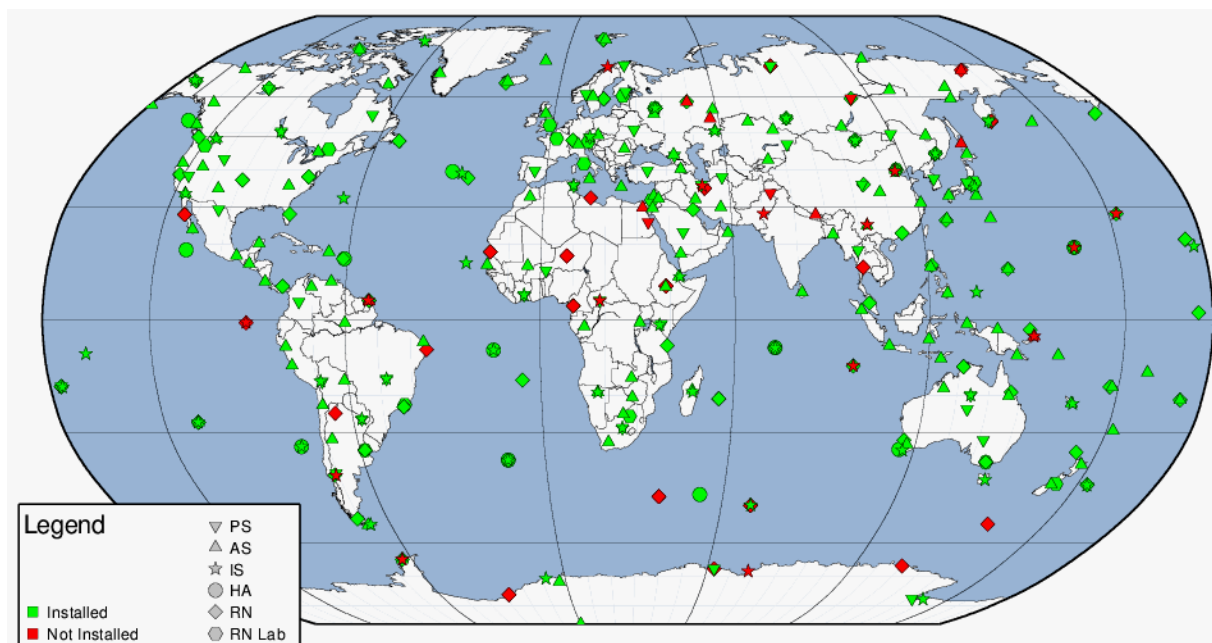
A number of techniques have been developed to discriminate between underground nuclear explosions and earthquakes (Bowers and Selby, 2009). Currently, the most effective

technique used at the IDC is the mb/Ms screening criterion, which relies on the fact that for an explosion and an earthquake with the same mb magnitude, the explosion will have a smaller Ms magnitude than the earthquake (see section 11.2.5.2 in Chapter 11, and Fig. 2 in Richards and Wu, 2011). Other criteria used for screening at the IDC consider event depth, the offshore location of an event, and regional seismic characteristics. Each of these criteria relies heavily upon seismic observations.

## 15.2 CTBTO networks

The International Monitoring System (IMS) is comprised of monitoring facilities as specified in the CTBT. The monitoring facilities are explicitly listed in the CTBT, and consist of 50 primary seismic stations and 120 auxiliary seismic stations, 11 hydroacoustic stations (6 hydrophone stations and 5 T-phase stations), 60 infrasound stations, and 80 radionuclide stations (80 stations capable of monitoring particulate radionuclides in the atmosphere, where 40 of those stations will be capable of monitoring for noble gases at the time the Treaty enters into force), and 16 radionuclide laboratories, which provide support activities.

The technical and operational requirements of each type of facility are specified in the appropriate IMS Operational Manual. Each new facility is certified and added to the IMS network once it is built and passes the certification process, in which it is demonstrated that it meets all technical specifications, including requirements for data authentication and transmission through the Global Communications Infrastructure (GCI) link to the IDC.



**Fig. 15.2** The complete IMS Network. The legend shows Primary Seismic (PS), Auxiliary Seismic (AS), Infrasound (IS), Hydroacoustic (HA), Radionuclide Particulate (RN (P)), Radionuclide Noble Gas (RN (NG)), and Radionuclide Laboratory (RN LAB) facilities.

### 15.2.1 Seismic station

A seismic station has three basic parts; a seismometer to measure the ground motion, a recording system which records the data digitally with an accurate time stamp, and a communication system interface.

Within the primary and auxiliary seismic networks, there are two types of seismic stations, three-component (3C) stations and array stations. The primary seismic network is mostly composed of arrays (30 arrays out of 50 stations), whereas the auxiliary seismic network is mostly composed of 3 C stations.

As of 1 April 2011 42 primary seismic and 99 auxiliary seismic stations have been certified. The specifications for primary and auxiliary seismic stations are summarized in Tab. 15.1.

#### 15.2.1.1 Three-component station

A three-component seismic station requires recording of broadband ground motion in three orthogonal directions. This type of recording can be done using either a single three-component broadband seismometer that covers the combined long-period and short-period frequency ranges, or with separate long-period and short-period seismometers (see DS 5.1).

#### 15.2.1.2 Array station

An IMS seismic array station consists of multiple short-period seismometers and three-component broadband instruments. Some existing short-period arrays also have associated multiple long-period seismometers. Both short-period and long-period seismometers are typically deployed in a geometrical configuration designed to allow ground motion signals to be combined to enhance the signal-to-noise ratio (see Chapter 9). New arrays should contain at least 9 vertical short-period elements and at least one three-component broadband element. The broadband element can use a single broadband instrument or one three-component short-period instrument with one three-component long-period instrument (see Tab. 15.1).

A teleseismic array is a seismic array designed for optimum detection and slowness estimation of seismic phases from sources at distances greater than 3000 kilometres. The distances between the array sensors are typically between 1000 and 3000 metres, and the total aperture is up to 60 kilometres.

A regional array is a seismic array designed for optimum detection and slowness estimation of seismic phases from sources at distances less than 3000 kilometres. The distances between array sensors are typically between 100 and 3000 metres, and the total aperture is up to 5 kilometres.

Primary seismic stations send continuous near-real-time data to the International Data Centre. Auxiliary seismic stations provide data upon request from the International Data Centre. In order to provide information on their operational state, in addition to ground motion data, seismic stations transmit state-of-health information such as:

- Clock status, to indicate whether the clock is synchronized to Coordinated Universal Time.

- Calibration status, to indicate whether a calibration signal is on or off.
- Vault/borehole status, to indicate whether the lid or door of the equipment vault or borehole is closed or open as a data surety measure

**Tab. 15.1** Specifications for primary and auxiliary seismic stations (minimum requirements; from CTBT/PC/II/Add.2 16 May 1997, p.43).

| CHARACTERISTICS                                           | MINIMUM REQUIREMENTS                                                                                                   |
|-----------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------|
| <i>Sensor type</i>                                        | Seismometer                                                                                                            |
| <i>Station type</i>                                       | Three-component or array                                                                                               |
| <i>Position (with respect to ground level)</i>            | Borehole or vault                                                                                                      |
| <i>Three-Component Passband</i> <sup>1</sup>              | Short-period: 0.5 to 16 Hz plus Long-period: 0.02 to 1 Hz or Broadband: 0.02 to 16 Hz                                  |
| <i>Sensor response</i>                                    | Flat to velocity or acceleration over the passband                                                                     |
| <i>Array Passband</i>                                     | (Short-period: 0.5 to 16 Hz<br>Long-period: 0.02 to 1 Hz) <sup>2</sup>                                                 |
| <i>Number of sensors for new arrays</i> <sup>3</sup>      | 9 short-period (one component) plus 1 short-period (three component) plus 1 long-period (three component) <sup>4</sup> |
| <i>Seismometer noise</i>                                  | ≤ 10 dB below minimum-earth noise at the site over the passband                                                        |
| <i>Calibration</i>                                        | Within 5% in amplitude and 5 degrees in phase over the passband                                                        |
| <i>Sampling rate</i>                                      | ≥ 40 samples per second <sup>5</sup><br>Long-period: ≥ 4 samples per second                                            |
| <i>System Noise</i>                                       | ≤ 10 dB below the noise of the seismometer over the passband                                                           |
| <i>Resolution</i>                                         | 18 dB below the minimum local seismic noise                                                                            |
| <i>Dynamic range</i>                                      | ≥ 120 dB                                                                                                               |
| <i>Absolute timing accuracy</i>                           | < 10 ms                                                                                                                |
| <i>Relative timing accuracy</i>                           | < 1 ms between array elements                                                                                          |
| <i>Operation temperature</i>                              | -10 °C to + 45 °C <sup>6</sup>                                                                                         |
| <i>State of health</i>                                    | Status to be transmitted to the International Data Centre: clock, calibration, vault and/or borehole status, telemetry |
| <i>Delay in transmission to International Data Centre</i> | < 5 minutes                                                                                                            |
| <i>Data frame length</i>                                  | Short-period: < 10 seconds; long-period: < 30 seconds                                                                  |

<sup>1</sup> For existing Global Telemetered Seismic Network stations, upgrading needs further consideration.

<sup>2</sup> For one-component element of teleseismic arrays, the upper limit is 8 Hz.

<sup>3</sup> In case of noisy sites or when increased capability is required, number of sensors could be increased.

<sup>4</sup> Can be achieved by a single broadband instrument.

<sup>5</sup> This applies to three-component and regional arrays. For existing teleseismic arrays, 40 samples per second are necessary for three-component but 20 samples per second are suitable for other sensors.

<sup>6</sup> Temperature range to be adapted for some specific sites.



|                                                                  |                                                                                                                         |
|------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------|
| <i>Buffer at station or at National Data Centre</i> <sup>1</sup> | >7 days                                                                                                                 |
| <i>Data availability</i>                                         | > 98%                                                                                                                   |
| <i>Timely data availability</i>                                  | > 97%                                                                                                                   |
| <i>Mission-capable arrays</i>                                    | >80% of the elements should be operational                                                                              |
| <i>Precision on station location</i>                             | <100 m absolute for stations (World Geodetic System 84)<br>< 1 m relative for arrays<br>elevation above sea level <20 m |
| <i>Seismometer orientation</i>                                   | < 3 degrees                                                                                                             |
| <i>Data format</i>                                               | Group of Scientific Experts (GSE) format                                                                                |
| <i>Data transmission</i>                                         | primary : continuous<br>auxiliary : segmented                                                                           |

## 15.2.2 Hydroacoustic station

In order to detect the acoustic energy ducted through the SOFAR channel a network composed of 11 hydroacoustic stations is being implemented. Six stations will be equipped with hydrophones. The other five hydroacoustic stations are located on steep-sloped islands and will make use of seismic sensors to detect the waterborne energy which is converted to a seismic wave at the boundary of the island. This type of propagation has long been known to the seismic community as T-phase propagation. T-phase stations are not as effective in detecting and identifying hydroacoustic signals from explosions, but they are considerably cheaper than hydrophone stations. The mixture of hydrophone and T-phase stations selected for the IMS hydroacoustic network was chosen as a cost-effective compromise. As of 1 April 2011 10 hydroacoustic stations have been certified.

An International Monitoring System hydroacoustic station can consist of either ocean-deployed hydrophone sensors and data acquisition systems, referred to herein as a *Hydrophone Station*, or one or more island-deployed seismometer sensors and data acquisition systems, referred to herein as a *T-phase Station*. There are 6 hydrophone stations and 5 T-phase stations.

### 15.2.2.1 Hydrophone station

A hydrophone station consists of an underwater segment and a data acquisition segment. When the hydrophone station is located on an island, two distinct cables and hydrophone sensors are deployed nominally off opposite shores. The hydrophone sensor elements are placed at or near the axis of the Sound Fixing and Ranging (SOFAR) channel, using a subsurface float and an ocean-bottom anchor. Each cable has three sensors with electronics for digitizing the signal. In order to provide the station with some directional capabilities, the three hydrophones are placed in a triangular configuration and each sensor is separated horizontally by a distance of approximately 2 kilometers. A single cable is used to bring the signals from the hydrophone sensors to shore. Hydrophone locations are selected considering

---

<sup>1</sup> Procedures for buffering to ensure minimum loss of data and single point failure should be addressed in the International Monitoring System Operational Manual.

the best combination of these factors: local optimum of low noise, wide viewing azimuth, and ease of cable installation.

All hydrophone stations, except for Cape Leeuwin, are located on relatively small islands. They generally consist of two undersea trunk cables, each with three hydrophones sensors. To avoid bathymetric blockage, by the island, cables and sensors are deployed on opposite shores of the island. Each hydrophone sensor has an independent wet-end digitizer. The digital signals are transmitted to the shore facility via a non-repeated fiber optic cable for processing and transmission to the IDC in Vienna. The (minimum requirement) specifications for hydrophone stations are summarized in Tab. 15.2.

**Tab. 15.2** Specifications for IMS Hydrophone Stations (minimum requirements; from CTBT/PC/II/Add.2 16 May 1997, p.46)

| CHARACTERISTICS                                           | MINIMUM REQUIREMENTS                                                                                 |
|-----------------------------------------------------------|------------------------------------------------------------------------------------------------------|
| <i>Sensor type</i>                                        | Hydrophone with wet-end digitiser                                                                    |
| <i>Passband</i>                                           | 1 to 100 Hz                                                                                          |
| <i>Sensor Response</i>                                    | Flat to pressure over the passband                                                                   |
| <i>Number of sensors</i>                                  | 1 operational sensor with 2 back-up sensors per cable                                                |
| <i>Sensors location</i>                                   | In the Sound Fixing and Ranging channel                                                              |
| <i>Location precision</i>                                 | ≤500 meters                                                                                          |
| <i>Number of cables</i>                                   | 2 at a site when necessary to prevent local blockage                                                 |
| <i>System noise</i>                                       | ≤10 dB below Urick's deep ocean low noise curve                                                      |
| <i>Calibration</i>                                        | Within 1 dB, no phase requirements                                                                   |
| <i>Sampling rate</i>                                      | ≥240 samples per second                                                                              |
| <i>Timing accuracy</i>                                    | ≤10 ms                                                                                               |
| <i>Delay in transmission to the Technical Secretariat</i> | ≤5 minutes                                                                                           |
| <i>State of health</i>                                    | Status to be transmitted to the International Data Centre: hydrophone, clock, calibration, telemetry |
| <i>Data availability</i>                                  | ≥ 98%                                                                                                |
| <i>Timely data availability</i>                           | ≥ 97%                                                                                                |
| <i>Sensitivity</i>                                        | ≤60 dB per μPa (1-Hz band)<br>≤81 dB per μPa (wide band)                                             |
| <i>Dynamic range</i>                                      | 120 dB                                                                                               |
| <i>Data transmission</i>                                  | Continuous                                                                                           |
| <i>Data format</i>                                        | GSE format                                                                                           |
| <i>Data frame length</i>                                  | ≤10 seconds                                                                                          |
| <i>Buffer at dry end</i>                                  | ≥7 days                                                                                              |
| <i>Mean time between failures for wet-end equipment</i>   | 20 years (to be confirmed)                                                                           |

### 15.2.2.2 T-phase station

A T-phase station consists of one or more seismometers and one or more data acquisition systems. The T-phase station is sited near the shore of small islands with steep bathymetry in order to detect seismic waves generated by the coupling of waterborne energy along the flanks of the island. Each T-phase station location is chosen to provide as large an azimuthal monitoring coverage as possible while minimizing the background seismic noise level. Up to three T-phase seismometer systems may be required at different island locations to provide full azimuthal monitoring coverage. For specifications of T-phase stations see Tab. 15.3.

**Tab. 15. 3** Specifications for IMS T-phase stations (minimum requirements; from CTBT/PC/II/Add.2 16 May 1997, p.45)

| CHARACTERISTICS                                                  | MINIMUM REQUIREMENTS                                                                                                   |
|------------------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------|
| <i>Sensor type</i>                                               | Seismometer                                                                                                            |
| <i>Station type</i>                                              | Minimum of one vertical component                                                                                      |
| <i>Position (with respect to ground level)</i>                   | Borehole or vault                                                                                                      |
| <i>Passband</i>                                                  | 0.5 to 20 Hz                                                                                                           |
| <i>Sensor response</i>                                           | Flat to velocity or acceleration over the pass band                                                                    |
| <i>Seismometer noise</i>                                         | $\leq 10$ dB below minimum-earth noise at the site over the passband                                                   |
| <i>Calibration</i>                                               | Within 5% in amplitude and 5 degrees in phase over the passband                                                        |
| <i>Sampling rate</i>                                             | $\geq 50$ samples per second                                                                                           |
| <i>Resolution</i>                                                | 18 dB below the minimum local seismic noise                                                                            |
| <i>System Noise</i>                                              | $\leq 10$ dB below the noise of the seismometer over the passband                                                      |
| <i>Dynamic range</i>                                             | $\geq 120$ dB                                                                                                          |
| <i>Absolute timing accuracy</i>                                  | $\leq 10$ ms                                                                                                           |
| <i>Operation temperature</i> <sup>1</sup>                        | -10 °C to + 45 °C                                                                                                      |
| <i>State of health</i>                                           | Status to be transmitted to the International Data Centre: clock, calibration, vault and/or borehole status, telemetry |
| <i>Delay in transmission to Technical Secretariat</i>            | $\leq 5$ minutes                                                                                                       |
| <i>Data frame length</i>                                         | $\leq 10$ seconds                                                                                                      |
| <i>Buffer at station or at National Data Centre</i> <sup>2</sup> | $\geq 7$ days                                                                                                          |
| <i>Data availability</i>                                         | $\geq 98\%$                                                                                                            |
| <i>Timely data availability</i>                                  | $\geq 97\%$                                                                                                            |

<sup>1</sup> Temperature range to be adapted for some specific sites

<sup>2</sup> Procedures for buffering to ensure minimum loss of data and single point failure should be addressed in the International Monitoring System Operational Manual.

|                                      |                                                                                                      |
|--------------------------------------|------------------------------------------------------------------------------------------------------|
| <i>Precision on station location</i> | $\leq 100$ m absolute for stations (World Geodetic System 84); Elevation above sea level $\leq 20$ m |
| <i>Seismometer orientation</i>       | $\leq 3$ degrees                                                                                     |
| <i>Data format</i>                   | Group of Scientific Experts format                                                                   |
| <i>Data transmission</i>             | Continuous                                                                                           |

### 15.2.3 Infrasound station

An infrasound station consists of an array of 4 up to 15 elements; typical designs have 4 to 8 elements. Each array element has three basic parts: 1) a microbarometer connected to a wind noise reduction system to measure the infrasound signals, 2) a recording system to record digital data with an accurate time stamp, and 3) a communication system interface.

A typical infrasound station is designed for infrasound signals detection along with slowness and azimuth estimation in the frequency regime of 0.02 to 4 Hz. The arrays are usually shaped either, as irregular triangles with an internal element or an inner sub-array, or as pentagons with a triangular inner sub-array. The aperture of the main arrays (outer) varies between 1 and 3 km while the sub-arrays (inner) usually have an aperture of 100 to 300 m.

The total number of the arrays in the IMS infrasound network is 60. As of 1 April 2011, 43 stations are installed. All the stations in the IMS infrasound network are new and are or will be constructed under the IMS program. All the infrasound stations send continuous near-real-time data to the International Data Centre, in Vienna.

In order to provide information on their operational status, in addition to micropressure data, infrasound stations transmit also state-of-health information, such as:

- Clock status, to indicate whether the clock is synchronized to Coordinated Universal Time.
- Vault status, to indicate whether the lid or door of the equipment vault is closed or open as a data surety measure

The minimum requirements for the installation and certification of an infrasound station are summarized in Tab. 15.4.

**Tab. 15.4** Specifications for IMS Infrasound Stations (minimum requirements; from CTBT/PC/II/Add.2 16 May 1997, p. 47)

| CHARACTERISTICS                  | MINIMUM REQUIREMENTS                           |
|----------------------------------|------------------------------------------------|
| <i>Sensor type</i>               | Microbarograph                                 |
| <i>Number of sensors</i>         | 4-element array <sup>1</sup>                   |
| <i>Geometry</i>                  | Triangle with a component at the centre        |
| <i>Spacing</i>                   | Triangle basis: 1 to 3 km <sup>2</sup>         |
| <i>Station location accuracy</i> | $\leq 100$ m                                   |
| <i>Relative sensor location</i>  | $\leq 1$ m                                     |
| <i>Measured parameter</i>        | Absolute <sup>3</sup> or differential pressure |
| <i>Passband</i>                  | 0.02 to 4 Hz                                   |

|                                                     |                                                  |
|-----------------------------------------------------|--------------------------------------------------|
| <i>Sensor response</i>                              | Flat to pressure over the passband               |
| <i>Sensor noise</i>                                 | ≤18 dB below minimum acoustic noise <sup>4</sup> |
| <i>Calibration</i>                                  | <5% in absolute amplitude <sup>5</sup>           |
| <i>State of health</i>                              | Status data transmitted to the IDC               |
| <i>Sampling rate</i>                                | ≥10 samples per second                           |
| <i>Resolution</i>                                   | ≥1 count per 1 mPa                               |
| <i>Dynamic range</i>                                | ≥108 dB                                          |
| <i>Timing accuracy</i>                              | ≤1 msec                                          |
| <i>Standard temperature range</i>                   | -10 °C to +45 °C <sup>6</sup>                    |
| <i>Buffer at station or at National Data Centre</i> | ≥7days                                           |
| <i>Data format</i>                                  | Group of Scientific Experts format               |
| <i>Data frame length</i>                            | ≤30 seconds                                      |
| <i>Data transmission</i>                            | Continuous                                       |
| <i>Data availability</i>                            | ≥98%                                             |
| <i>Timely data availability</i>                     | ≥97%                                             |
| <i>Mission capable array</i>                        | ≥3 elements operational                          |
| <i>Acoustic filtering</i>                           | Noise reduction pipes (site dependent)           |
| <i>Auxiliary data</i>                               | Meteorological data <sup>7</sup>                 |

<sup>1</sup> In case of noisy sites or when increased capability is required, number of components could be increased.

<sup>2</sup> 3 km is the recommended spacing

<sup>3</sup> Used for daily state of health.

<sup>4</sup> Minimum noise level at 1 Hz : ~ 5 mPa

<sup>5</sup> Periodicity: once per year (minimum).

<sup>6</sup> Temperature range to be adapted for some specific sites.

<sup>7</sup> Once per minute (minimum).

#### 15.2.4 Radionuclide particulate station

Among the monitoring technologies in the IMS network, radionuclide monitoring is the only technique that provides the forensic or confirmatory evidence that a detected explosion is nuclear in nature. . Eighty radionuclide stations are being established in the IMS for sampling and radionuclide analysis of airborne particulates. Forty of these stations will also have the capability for sampling and analysis of noble gases. The radionuclide network was designed to achieve a detection capability of not less than 90% within approximately 14 days (including 3 days reporting time) for a 1 kt nuclear explosion in the atmosphere or from venting by an underground or underwater detonation.

Following a nuclear explosion, the radioactive materials released consist mainly of fission products and neutron activation products (from the interaction of the neutrons released during fission with elements in the device and surrounding materials). For particulate samples, there are 84 fission and neutron activation products which are regarded as relevant to CTBT. From their relative abundance or mutual ratios in the sample, it will be possible to determine whether they are from a nuclear explosion or other possible sources of radioactivity such as nuclear power or reprocessing plants.

In a radionuclide particulate station, at least 12,000 m<sup>3</sup> of air is pumped daily through a filter with high collection efficiency in order to produce a large volume of air sample for high detection sensitivity. This sample is then analyzed using high-resolution gamma-ray spectrometry since the fission products and neutron activation products are gamma-emitters. The data is directly transmitted to the IDC thru the GCI for processing, analysis and review.

The main components of a radionuclide particulate station are:

- High-volume air- sampler – for collection of large volume of airborne particulates or aerosols;
- Filter material - for trapping particulates with high efficiency; after sampling, this is compressed or cut into a geometry best suited for gamma spectrometry;
- Detection equipment – high-purity germanium crystal with good resolution, high efficiency and encased inside lead shielding to reduce background radioactivity;
- Multi-channel analyzer, computer system, station operation software - for production of spectral raw data for transmission to the IDC for analysis and operation of the system;
- State-of-health (SOH) sensors - status of station monitors, i.e., air flow rate, detector temperature, indoor temperature and humidity, filter position monitor, power supply status, lead shield status, that could be used to interpret the measured radionuclide data and provide an indication of normal and secure operation;
- Meteorological sensors - meteorological data monitors, i.e., precipitation, temperature, wind speed, wind direction;
- Very small aperture antenna (VSAT) - for transmission of data to IDC via satellite;
- Uninterruptible power supply and auxiliary generator - for power stability and alternate source of electrical power.

There are two types of radionuclide particulate stations: (1) manual station which requires an operator to perform the daily operations such as filter change and preparation, and (2) automatic station. The two types of automatic systems in the IMS network are the Radionuclide Aerosol Sampler/Analyzer (RASA) and Automatic Radionuclide Air Monitoring Equipment (ARAME). The minimum technical requirements for certification of a particulate station are listed in Tab. 15.5.

**Tab. 15.5** Specifications for IMS Radionuclide Particulate Stations (minimum requirements; from CTBT/PC/II/Add.2 16 May 1997, p. 48)

| CHARACTERISTICS        | MINIMUM REQUIREMENTS               |
|------------------------|------------------------------------|
| <i>System</i>          | Manual or automated                |
| <i>Airflow</i>         | 500 m <sup>3</sup> h <sup>-1</sup> |
| <i>Collection time</i> | 24 h                               |
| <i>Decay time</i>      | ≤ 24 h                             |

|                                                         |                                                                                                                     |
|---------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------|
| <i>Measurement time</i>                                 | $\geq 20$ h                                                                                                         |
| <i>Time before reporting</i>                            | $\leq 72$ h                                                                                                         |
| <i>Reporting frequency</i>                              | Daily                                                                                                               |
| <i>Filter</i>                                           | Adequate composition for compaction, dissolution and analysis                                                       |
| <i>Particulate collection efficiency</i>                | For filter : $\geq 80\%$ at $\varnothing = 0.2 \mu\text{m}$<br>Global $\geq 60\%$ at $\varnothing = 10 \mu\text{m}$ |
| <i>Measurement mode</i>                                 | HPGe high resolution gamma spectrometry                                                                             |
| <i>HPGe relative efficiency</i>                         | $\geq 40\%$                                                                                                         |
| <i>HPGe resolution</i>                                  | $< 2.5$ keV at 1332 keV                                                                                             |
| <i>Base line sensitivity</i>                            | 10 to 30 $\mu\text{Bq m}^{-3}$ for $^{140}\text{Ba}$                                                                |
| <i>Calibration range</i>                                | 88 to 1836 keV                                                                                                      |
| <i>Data format for gamma spectra and auxiliary data</i> | RMS (Radionuclide Monitoring System) format                                                                         |
| <i>State of health</i>                                  | Status data transmitted to IDC                                                                                      |
| <i>Communication</i>                                    | Two-way                                                                                                             |
| <i>Auxiliary data</i>                                   | Meteorological data<br>Flow rate measurement every 10 minutes                                                       |
| <i>Data availability</i>                                | $\geq 95\%$                                                                                                         |
| <i>Down time</i>                                        | $\leq 7$ consecutive days<br>$\leq 15$ days annually                                                                |

As of 1 April 2011, of the 80 radionuclide stations which comprise the IMS radionuclide particulate network, 60 have been certified as meeting the requirements above.

### 15.2.5 Radionuclide noble gas station

For underground or underwater nuclear explosions, the radioactive materials with greater probability of being vented and released into the atmosphere are noble gases rather than particulates. Noble gases are chemically inert and will not be attached to its surrounding environment; thus these can leak from an underground cavity through vents or cracks or be transported from underwater to the surface. Once in the atmosphere these remain gaseous, travel long distances and unlike particulates, are not washed out by precipitation. . Among the noble gases likely to be released into the atmosphere following a nuclear explosion, radioactive xenon isotopes (radioxenons) are presently the most suitable for monitoring purposes. They are produced with a relatively high yield in nuclear explosions and have suitable half-lives. By examining the ratios of xenon isotopes in a sample, it is possible to distinguish whether the radioxenons detected are from a nuclear explosion or other civilian applications such as nuclear reactor releases. The 4 radioxenons monitored by the noble gas stations are  $^{131\text{m}}\text{Xe}$ ,  $^{133\text{m}}\text{Xe}$ ,  $^{133}\text{Xe}$  and  $^{135}\text{Xe}$ , which are beta (or conversion electron) and gamma emitters. The technology for sampling and analysis of noble gas under field conditions is less mature and more complex than that for particulates. Thus, the noble gas

systems have been tested and evaluated under the International Noble Gas Experiment (INGE) conducted by the Technical Secretariat.

In a noble gas station, the air sample collected undergoes processing to separate and concentrate xenon from other constituents. The activity of the purified/concentrated radionuclides is then measured by high-resolution gamma-ray spectrometry or beta-gamma coincidence counting. The volume of air analyzed is determined from the volume of non-radioactive xenon (stable xenon), taking into account the natural abundance of xenon in air (0.087 ppmV at STP). Stable xenon is measured using gas chromatography or thermal conductivity detector.

The three types of noble gas systems in the IMS network, all of which operate automatically are: (1) SPALAX (Système de Prélèvement Atmosphérique en Ligne avec l'Analyse du Xénon) (2) SAUNA (Swedish Automatic Unit for Noble Gas Acquisition) and (3) ARIX (Analyzer of Radioactive Isotopes of Xenon). The SPALAX uses high-resolution gamma-ray spectrometry while the SAUNA and ARIX use beta-gamma coincidence counting.

The minimum technical requirements for certification of noble gas stations are listed below in Tab. 15.6.

**Tab. 15.6** Specifications for IMS Radionuclide Noble Gas Stations (minimum requirements; CTBT/PC/II/Add.2 16 May 1997, p. 49)

| CHARACTERISTICS                                      | MINIMUM REQUIREMENTS                                                         |
|------------------------------------------------------|------------------------------------------------------------------------------|
| <i>Air flow</i>                                      | 0.4 m <sup>3</sup> /h                                                        |
| <i>Total volume of sample</i>                        | 10 m <sup>3</sup>                                                            |
| <i>Collection time</i>                               | ≤24 h                                                                        |
| <i>Measurement time</i>                              | ≤24 h                                                                        |
| <i>Time before reporting</i>                         | ≤48 h                                                                        |
| <i>Reporting frequency</i>                           | Daily                                                                        |
| <i>Isotopes measured</i>                             | <sup>131m</sup> Xe, <sup>133m</sup> Xe, <sup>133</sup> Xe, <sup>135</sup> Xe |
| <i>Measurement mode</i>                              | Beta–gamma coincidence<br>or<br>high resolution gamma spectrometry           |
| <i>Minimum Detectable Concentration</i> <sup>1</sup> | 1 mBq/m <sup>3</sup> for <sup>133</sup> Xe                                   |
| <i>State of health</i>                               | Status data transmitted to IDC                                               |
| <i>Communication</i>                                 | Two way                                                                      |
| <i>Data availability</i>                             | 95%                                                                          |
| <i>Downtime</i>                                      | ≤7 consecutive days<br>≤15 days annually                                     |

As of 1 April 2011, there are 27 noble gas stations installed out of the 40. Certification started in 2010, with 3 noble gas stations certified as meeting the abovementioned requirements.

<sup>1</sup> Smallest concentration of radioactivity in a sample that can be detected with a 5% probability of erroneously detecting radioactivity, when in fact none was present (Type I error) and also, a 5% probability of not detecting radioactivity, when in fact it is present (Type II error).



## **15.2.6 Radionuclide laboratories**

In accordance with the CTBT, the network of radionuclide monitoring stations is supported by 16 radionuclide laboratories which perform re-analysis of samples from stations. The purpose of re-analysis of samples from stations by a laboratory is: (1) to corroborate the results of the routine analysis of a sample from an IMS station, in particular to confirm the presence of fission products and/or activation products; (2) to provide more accurate and precise measurements; and (3) to clarify the presence or absence of fission products and/or activation products in the case of a suspect or irregular analytical result from a particular station (CTBT/WGB/TL-11, 17-18 /Rev. 4).

These laboratories have a quality system to include sample handling, preparation, measurement, data analysis and reporting. Their software enables data transmission to the IDC in a standardised format. The Technical Secretariat organizes annual proficiency test or intercomparison exercises as part of a quality assurance program. These laboratories are certified by the Technical Secretariat according to rigorous management and technical requirements contained in PTS/INF.96 Rev. 7. As of 1 April 2011, ten of the 16 radionuclide laboratories are certified.

## **15.3 Data acquisition**

### **15.3.1 Introduction**

The stations in the IMS network send data to the International Data Centre (IDC) according to the Formats and Protocols specified in the corresponding IMS Operational Manual. Stations in the primary seismic, hydroacoustic, and infrasound networks send data in Continuous Data (CD) format. As the name implies, data from these stations are sent continuously to the IDC.

Stations in the auxiliary seismic network send data to the IDC in response to a data request received from the IDC. Each data request is formulated based on IDC data processing results from stations in the primary seismic, hydroacoustic, and infrasound networks. Consequently, only segments of data from auxiliary seismic stations are sent to the IDC. Radionuclide stations send data to the IDC on a regular daily schedule as the data become available at the station.

All IMS data received at the IDC are parsed and are accessible through the IDC relational database management system (RDBMS). After the data are stored in the RDBMS, they are available for automatic processing.

All stations in the IMS network can be securely managed using command and control messages. The communication between IMS stations and the IDC, and between the IDC and users is done over the Global Communications Infrastructure (GCI).

### **15.3.2 Security concerns**

The security concerns of the IMS data are addressed in the following ways. Data collected at IMS stations are signed with a private key, and the resulting signature from that process is

transmitted along with the data. The recipient of the data can then use the data, the signature, and the public key from the station to authenticate the data. If the data passes this authentication process, it means that the data were signed by the private key at the station, and that the data have not been modified.

The private key is stored in a hardware device, and with tamper resistant features. Any attempt to tamper with the private key is reported.

The private and public keys are issued and maintained using a Public Key Infrastructure (Xenitellis, 2000), which is operated and maintained by CTBTO. This process of authentication and verification is also used for command and control messages, which are elaborated in section 3.6.

### **15.3.3 Continuous data stations**

Primary seismic, hydroacoustic, and infrasound stations send data to the IDC in Continuous Data (CD) format. At present there are two versions of CD format, namely CD-1.0 format and CD-1.1 format. Both are currently in use, but all new stations employ CD-1.1 format.

The data completeness and timeliness requirements for continuous data stations are quite stringent. These stations are required to send at least 98% of all expected data to the IDC, and 97% of the data should be received at the IDC within 5 minutes after the data have been recorded at the station. Consequently, after any type of outage of sending data, it is very important to restore the real time data flow to the IDC in order to meet the timeliness requirement. Consequently, the CD format specifies a Last In First Out (LIFO) sequence of sending data after an extended outage. This means that the newly recorded data are sent first, which can be interspersed with older data, progressing from younger data to older data.

CD protocol is based on socket level communication. The data sender initiates the data flow by sending a Connection Request Frame to the receiver's well known port and IP address. The receiver responds by sending a Port Assignment Frame to the sender, which specifies the IP address and port where the sender should begin sending data. In case of CD-1.1 format, there is additional exchange of Option Request Frames at this stage.

The sender then begins to send the data to the receiver at the specified IP address and port. In case of CD-1.0, this consists of a Data Format Frame (DFF), which specifies the format of the subsequent Data Frames (DF). All DFs will adhere to the format specified in the DFF, and can only be interpreted with the DFF information. If there is a change of format, for example if a channel is added or removed from the DF, then a new DFF, preceded by an Alert Frame, must be sent by the sender. The data flow can be terminated by either party by sending the appropriate Alert Frame.

The fact that the data format information is only sent once in CD-1.0 format leads to an efficient use of communication bandwidth. However, it also means the Data Frames can only be interpreted with information from the corresponding Data Format Frame. In order to address this shortcoming, and to add additional functionality, CD-1.1 format was created.

In the case of CD-1.1 format, the Option Request Frames are exchanged after the Port Assignment Frame is sent to the sender, as mentioned above. The sender will then begin to

send Data Frames (DF). In CD-1.1, each DF fully describes the data, so there is no need for Data Format Frames, as is required for CD-1.0 format. The CD-1.1 Data Frames also contain a sequence number, which can be used to record which frames have been sent and received. The receiver and sender exchange Acknack Frames to synchronize this information. Every frame in CD-1.1 format can also be signed, to ensure integrity of the data.

CD-1.1 format also include the capability for Command and Control Frames, which can be used to manage IMS stations. More information concerning Command and Control can be found in Section 3.6.

In both CD-1.0 and CD-1.1 the Data Frame contains Channel Sub-Frames (CSF), where each CSF contains the data from a particular data channel or sensor. For example, a typical three component station will have three CSFs, with one CFS for each of the three channels BHZ, BHN, and BHE. The waveform data in each CSF is signed, and the resulting signature is also stored in the CSF. In the case of CD-1.1, the entire data frame will also be signed, and the signature of the entire frame will be sent in the frame trailer. The components of a CD-1.1 Data Frame are summarized in Fig. 15.3.

|                                                |                  |
|------------------------------------------------|------------------|
| Frame Header<br>and<br>Channel Subframe Header |                  |
| Channel Subframe 1                             | <i>signature</i> |
| Channel Subframe 2                             | <i>signature</i> |
| ...                                            |                  |
| Channel Subframe n                             | <i>signature</i> |
| Frame Trailer                                  | <i>signature</i> |

**Fig. 15.3** Components of a CD-1.1 Data Frame. Note that each channel subframe is signed, as well as the frame itself. The data frame signature is found in the frame trailer.

### 15.3.4 Segmented data stations

IMS auxiliary seismic stations send data to the IDC in response to a waveform data request from the IDC. The data request and the response adhere to IMS 2.0 Formats and Protocol. The request specifies the start and end time, as well as the station and channels which should be returned to the requestor.

The data request is based on a preliminary event location made using data from primary stations, i.e., stations which send data in continuous mode. Once an event is detected by the primary network, a data request is sent to all auxiliary stations within 30 degrees of this preliminary location plus auxiliary stations which have a high probability of detecting the

event. In this way, the auxiliary data assist in refining the location by providing additional regional observations.

The messages exchanged in IMS 2.0 format are sent via e-mail. The original concept of using email for waveform data exchange was pioneered by Urs Kradolfer (Kradolfer, 1996). Prior to IMS 2.0, there were a number of earlier revisions of the format, including GSE 2.0, GSE 2.1, and IMS 1.0 (see also Chapter 10 and IS 10.1 and 10.2). While additional data types and features have been added with each revision, the data request and response messages have remained relatively stable.

Since IMS 2.0 data are typically sent via email, the waveform data are converted into ASCII representation in the email message. There are a few different supported waveform data types. One data type is integer format, named “INT”, where the waveform digital counts are reported as integer values. The second format is named “CM6”, where the digital waveform counts are processed with an algorithm to compress the data and transmit the data in ASCII format. A third option allows waveform data to be sent along with corresponding data signatures. This format is named “CSF”, and is formed by converting a CD-1.1 Channel Sub-Frame (see Section 3.3), into ASCII format using base 64 encoding.

### **15.3.5 Radionuclide stations**

Radionuclide stations send data to the IDC on a regular daily schedule as the data become available at the station. The data are sent in IMS 2.0 format via e-mail. Particulate systems produce one radionuclide spectrum per day which is sent to the IDC. Noble Gas systems produce one or two radionuclide spectrum per day, depending on the type of measurement system. In addition, radionuclide stations send state of health information on a regular basis, ranging from 2 to 6 hours, depending on the type of system. Radionuclide stations also send various spectra types for calibration purposes, including blank spectra, detector background spectra, and calibration source spectra.

### **15.3.6 Command and control**

Command and Control messages are used to securely command IMS stations. The process is typically initiated by sending a signed command message from the IDC to the station. At the station the command is checked for authenticity, the requested action is carried out, and a command response message is returned to the IDC. The commands are used for seismic station calibration. Command messages are sent in IMS 2.0 format via email. There are also provisions in CD-1.1 format to support the exchange of command and control frames.

### **15.3.7 Global Communication Infrastructure**

Data are sent from the IMS stations to the IDC over the Global Communication Infrastructure (GCI). The GCI is also used to send IMS data and IDC products from the IDC to authorized users. The topology between stations and the IDC is typically done in one of two ways; one option is basic topology, and the other is the Independent Sub Network (ISN).

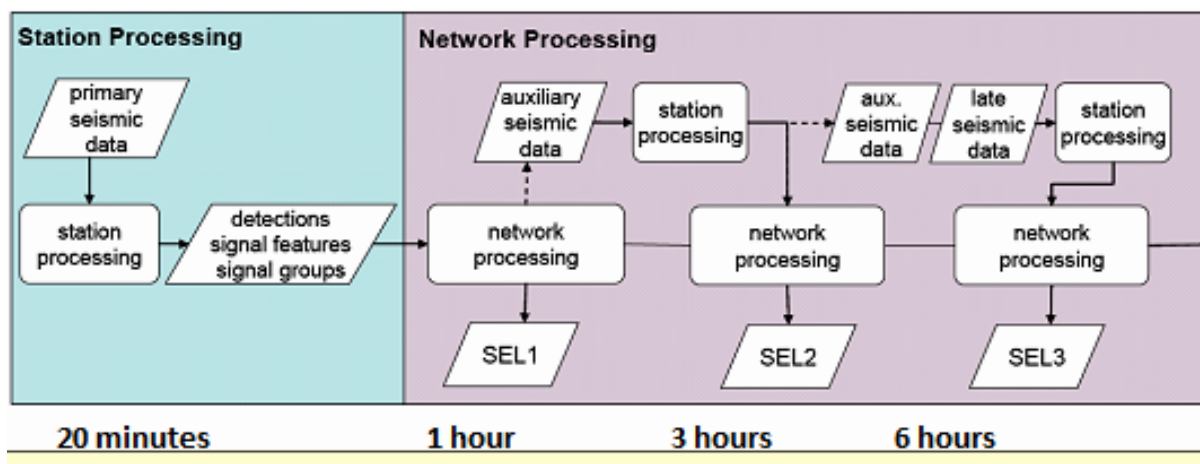
In the case of basic topology, data are sent from the station over a Very Small Aperture Terminal (VSAT) through a satellite link to a satellite ground station, which in turn is connected to the IDC via a terrestrial link. In the case of an ISN, a country may take the responsibility to transport data from several stations on its territory to a central location. The data at that central location are then sent onward to the IDC, typically over a terrestrial connection.

The complete GCI is a large operation which includes 215 VSAT connections and 6 terrestrial connections. The GCI link availability requirement is 99 %, which is currently being routinely achieved.

## 15.4 Waveform data processing

### 17,4.1 Introduction

Waveform data are automatically processed once they have arrived at the IDC. Waveform station processing for continuous data stations is done in fixed time intervals (Fig. 15.4). The duration of those time intervals is 10 minutes for primary seismic and hydroacoustic stations, and 30 minute intervals for infrasound stations. Each interval is processed after 95 % of data used for station processing from that station has arrived in the RDBMS. Data from auxiliary seismic stations are processed once the requested data segments have been parsed into the RDBMS and a sufficient amount of requested data has been received.



**Fig. 15.4** Overview of station processing and network processing. Note that network processing is run three times, to accommodate data from auxiliary stations and late arriving data. Network processing results in the Standard Event Bulletins (SEL) 1, 2, and 3, which are automatically generated 1, 3, and 6 hours after real time.

Waveform station processing performs quality control checks, band-pass filtering and beamforming (see Chapters 4 and 9) to improve the signal-to-noise ratio and detect signals in the data interval. Feature extraction is then performed for each detection and the resulting detection time (see also Chapter 16), amplitude, period, azimuth and slowness are written to the RDBMS. The final stages of station processing include signal grouping, initial phase

identification, and location of single station events. The arrivals resulting from station processing are the input in network processing, where the arrivals are combined to form events which are published as automatic bulletins, named SEL1, SEL2, and SEL3 where SEL is an abbreviation for Standard Event List.

After the final automatic bulletin is produced, the data are ready for review by the IDC analysts. The analysts correct mistakes in the automatic bulletin, refine the results, and add events which were missed. The result of this work is published as the Reviewed Event Bulletin (REB). Currently, during provisional operations, the target timeline for publishing the REB is within 10 days of real time. After Entry Into Force (EIF) of the Treaty, the target timeline is reduced to 48 hours.

## 15.4.2 Seismic station processing

Broadly speaking, there are three types of IMS seismic stations; three component stations, array stations (which typically vary in aperture up to 20 kilometers), and one large array, which has an aperture of more than 60 kilometers. Different processing techniques are applied to these three station types, as described below.

Station processing is performed by a series of two applications, DFX (**D**etection and **F**eature **E**xtraction) and StaPro (**S**tation **P**rocessing). DFX is used to detect signals and extract their features. The output of DFX includes detection time, amplitude, period, signal to noise ratio (SNR), azimuth, azimuth uncertainty, slowness, and slowness uncertainty. The StaPro application recognizes noise arrivals, assigns initial phase names, and locates single station events. The output of StaPro includes the initial phase name and single station event locations.

### 15.4.2.1 DFX

DFX performs the following processing steps:

1. Data Quality Control (DQC) – removes bad channels, repairs data gaps and spikes;
2. Improve and Estimate SNR – performs filtering and beaming to improve SNR for all detection beams;
3. Find detection – use STA/LTA detector (see IS 8.1), best beam and trigger time;
4. Refine time – use Akaike Information Criteria (AIC) to refine onset time;
5. Measure amplitude and period;
6. Estimate azimuth and slowness – perform frequency-wavenumber (f-k) (cf. Chapter 9) or polarization analysis (see Fig. 2.6 of Chapter 2) to estimate slowness, azimuth, and associated error estimates;
7. In the case of array data, steps 4 and 5 are repeated, based on the results from step 6.

#### ***Data Quality Control***

Data Quality Control (DQC) is the initial step in data processing. Problematic data are identified as spikes, continuous zeros, repeated amplitude values, gaps, and channels with anomalous amplitude values. Spikes are recognized by the difference in amplitude relative to the surrounding data values. Repeated amplitude values are recognized by the number of consecutive samples with the same amplitude value (usually between 4 and 10). In some cases, problematic data can be repaired through interpolation, or detrending. If the data cannot

be repaired, then a mask is created indicating where the data have quality problems and should be excluded from processing. If a large percentage of the data (usually more than 33%) on a given channel are masked, then the channel is excluded from processing for the entire processing interval. Recall that processing intervals are typically 10 minutes long for seismic data.

### ***Improve and estimate SNR***

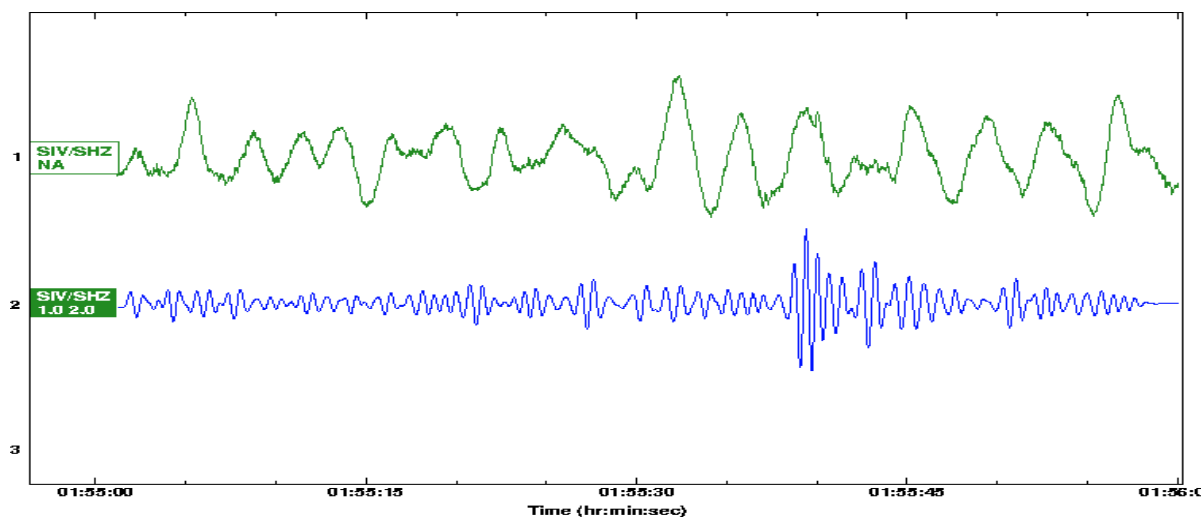
The SNR is improved at three component stations by filtering the vertical channel in various filter bands, and incoherent beams are made from the two horizontal channels. An example beam recipe for a three component station is shown in Fig. 15.5 and the filtered vertical channel at the station SIV is shown in Fig. 15.6.

An array station consists of sensors which are spatially distributed. When a signal traverses an array, it is recorded by individual sensors at different times. Time delays relative to the reference beam point can be calculated by using the relative distance to the reference point, and an assumed slowness vector (Fig. 15.7). Based on the coherence of the signal and the incoherence of noise, beaming delay-corrected signals from array elements can significantly improve the SNR of a signal.

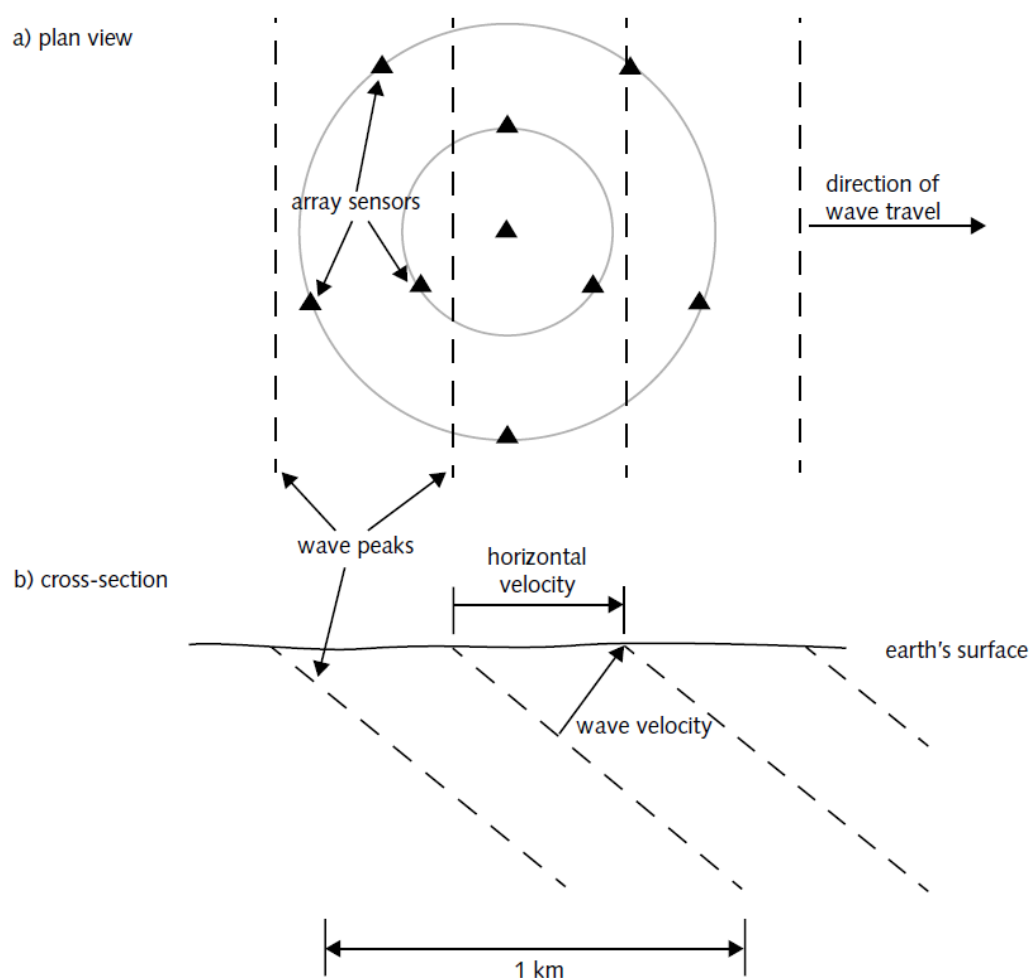
### **Typical beam recepies for 3-C stations:**

| name  | type | rot | std | snr  | azi | slow  | phase | flo  | fhi  | ford | zp | ftype | group      |
|-------|------|-----|-----|------|-----|-------|-------|------|------|------|----|-------|------------|
| Z0515 | coh  | no  | 0   | 3.50 | 0.0 | 0.000 | -     | 0.50 | 1.50 | 3    | 0  | BP    | vertical   |
| Z1020 | coh  | no  | 0   | 3.50 | 0.0 | 0.000 | -     | 1.00 | 2.00 | 3    | 0  | BP    | vertical   |
| Z1530 | coh  | no  | 0   | 4.00 | 0.0 | 0.000 | -     | 1.50 | 3.00 | 3    | 0  | BP    | vertical   |
| Z2040 | coh  | no  | 0   | 4.00 | 0.0 | 0.000 | -     | 2.00 | 4.00 | 3    | 0  | BP    | vertical   |
| Z3060 | coh  | no  | 0   | 5.00 | 0.0 | 0.000 | -     | 3.00 | 6.00 | 3    | 0  | BP    | vertical   |
| Z4080 | coh  | no  | 0   | 6.00 | 0.0 | 0.000 | -     | 4.00 | 8.00 | 3    | 0  | BP    | vertical   |
| H0515 | inc  | no  | 0   | 5.50 | 0.0 | 0.000 | -     | 0.50 | 1.50 | 3    | 0  | BP    | horizontal |
| H1020 | inc  | no  | 0   | 5.50 | 0.0 | 0.000 | -     | 1.00 | 2.00 | 3    | 0  | BP    | horizontal |
| H1530 | inc  | no  | 0   | 5.50 | 0.0 | 0.000 | -     | 1.50 | 3.00 | 3    | 0  | BP    | horizontal |
| H2040 | inc  | no  | 0   | 5.50 | 0.0 | 0.000 | -     | 2.00 | 4.00 | 3    | 0  | BP    | horizontal |
| H4080 | inc  | no  | 0   | 5.50 | 0.0 | 0.000 | -     | 4.00 | 8.00 | 3    | 0  | BP    | horizontal |

**Fig. 15.5** Example of a beam recipe for a three component station. The column headings are: name – channel name for the beam; type – beam type, coh – coherent or inc – incoherent; std – specifies whether or not this is a standard beam (1 if yes); snr – threshold snr for declaring a detection; azi – azimuth to use for beamforming at an array; slow – slowness to use for beamforming at an array; phase – phase to use for beamforming when making origin beams; flo – low cut value of the applied Butterworth filter; fhi – high cut value of the applied butterworth filter; ford – specifies the order of the applied Butterworth filter; zp – specifies if the applied Butterworth filter should be zero phase (1 if yes); ftype - filter of the applied Butterworth filter (BP – band pass; HP – high pass; LP – low pass); group – specifies which channels from the station should contribute to forming the beam.



**Fig. 15.6** Filtered (1.0 to 2.0 Hertz) vertical channel from the station SIV, in San Ignacio, Bolivia for arrival at 01:55:38.9 for event on 09-Oct-2006 resulted in a good signal-to-noise ratio (lower trace), in comparison with the more broadband velocity raw data (upper trace).



**Fig. 15.7** Illustration of a seismic plane wave passing through an array.



For each IMS array, there are several beams which are created, where each beam specifies a specific azimuth and slowness to steer the beam, specific filter band, and a list of which elements to use to form the beam. A beam recipe for an array station uses the same format as for 3-component stations, as was shown in Fig. 15.5.

### **Find detection**

An STA/LTA detector is used at the IDC for detecting signals in seismic data. STA (short-term-average) is a time-average of some function of data (for example, amplitude, power, or energy) over a short period of time relative to the LTA (long-term-average) (see IS 8.1). The STA/LTA detector for seismic and hydroacoustic data uses a running average for the STA and a recursive average for the LTA. The LTA “lags” the STA in time by half the width of the STA window. The STA and LTA are defined as:

$$\text{STA:} \quad \text{STA}(k) = \text{STA}(k-1) + \frac{1}{S} [ |x(k + \frac{S}{2})| - |x(k-1 - \frac{S}{2})| ]$$

$$\text{LTA:} \quad \text{LTA}(k) = (1 - \frac{1}{L})\text{LTA}(k-1) + \frac{1}{L} \text{STA}(k - S)$$

The initial value of the STA is:

$$\text{STA}(\frac{S}{2}) = \frac{1}{S} \sum_{S=0}^{S-1} |x(S)|$$

where:  $x(n)$ : time series

$S$ : the number of samples in the STA window

$L$ : the effective number of samples in the LTA window.

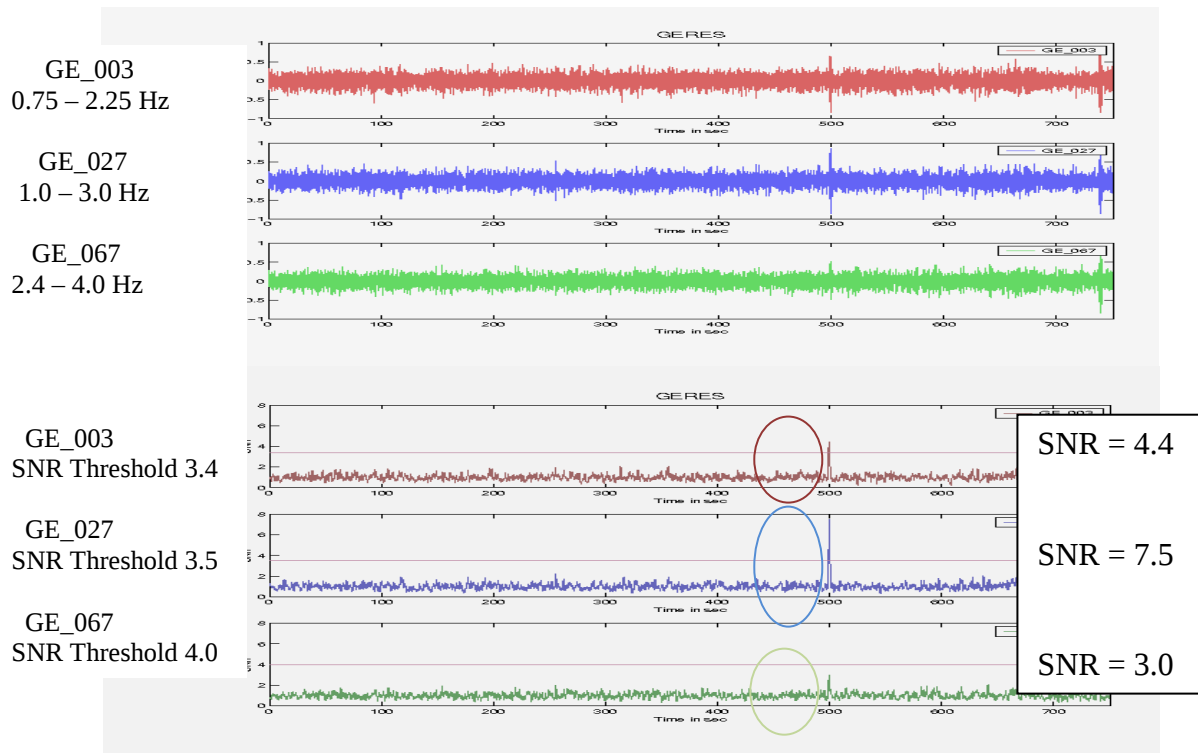
The typical STA length is one second (1.5 seconds for the large array NOA), and 60 seconds for the LTA length.

The STA/LTA threshold (SNR) is specified in the associated beam recipe. When the SNR threshold is exceeded, there is a potential detection. When the SNR thresholds are exceeded for many beams at nearly the same time, the beam with the largest SNR is identified as the best beam. This beam and related parameters are saved for subsequent station processing. An example of identifying the best beam is shown in Fig. 15.8.

### **Onset time refinement**

Once a detection is declared, the trigger time is refined using the Akaike Information Criteria (AIC) (Akaike, 1974). The AIC is applied with a signal window of four seconds with a three second lead time, and a noise window of three seconds which starts six seconds prior to the initial trigger time. The maximum adjustment time is up to two seconds compared to the initial trigger time. The onset time picking error is constrained in the range from 0.685 s to 1.72 s for automatic processing and from 0.12 s to 1.07 s for interactive processing.

For arrays, the AIC is applied twice, i.e., once for the detecting beam after the trigger time is declared, and again for the f-k beam after f-k analysis is performed.



**Fig. 15.8** The best beam for this detection at GERES is the beam with the highest SNR. Top three panels show three beams of GERES, while bottom panels show the SNR traces for these three beams. The beam named GE\_027 has an SNR of 7.5, which is the highest.

***Measure amplitude and period and determine magnitudes mb, Ms and ML***

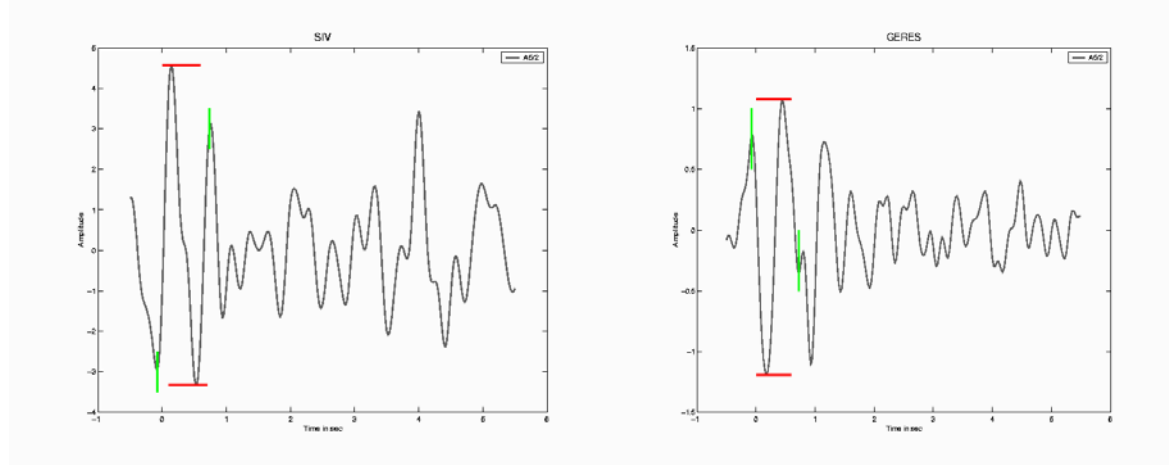
There are two amplitude types measured by DFX during seismic station processing (Tab. 15.7) which are used for calculating magnitude values. The A5/2 amplitude type is used for calculating the IDC mb magnitude during automatic processing, while the SBSNR amplitude is used to calculate the ML magnitude. After interactive review, the IDC mb magnitude is calculated using A5/2 amplitude type measured from the origin beam.

**Tab. 15.7** Amplitudes computed by DFX during station processing.

| Amp Type | Beam Type  | Filter Band  | Time Window       | Magnitude |
|----------|------------|--------------|-------------------|-----------|
| A5/2     | Coherent   | 0.8 – 4.5 Hz | 6.0 (0.5 + 5.5) s | mb        |
| SBSNR    | Incoherent | 2.0 – 4.0 Hz | 4.0 s             | ML        |

The A5/2 amplitude used for IDC mb magnitude is measured as half the maximum peak-to-trough (or trough-to-peak) amplitude on the vertical channel at three component stations using the fixed filter band 0.8 – 4.5 Hz, while at array stations the A5/2 amplitude is measured on the f-k beam using all vertical sensors at the array using the same filter band and the azimuth and slowness based on the origin for which the mb magnitude is being calculated. The time window used for finding the maximum amplitude is 0.5 s before the first P onset and 5.5 s after the first P onset. Thus, the name A5/2 refers to half peak-to-peak amplitude in a 6 second time window. The period of the A5/2 measurement is based on the three half periods

around the observed maximum amplitude (Fig. 15.9). The A5/2 amplitude measurement is made based on the maximum amplitude found in the time window specified in Tab. 15.7.



**Fig. 15.9** The mb amplitude is measured on filtered vertical channel for arrival at SIV (3-C) (left) and on a coherent f-k beam for all vertical channels at GERES (array) (right). The final period is based on the three half periods between the two green lines. The peak-to-trough amplitude is measured between the interpolated peaks indicated by red lines in the diagrams above.

In contrast to standard mb, which is calculated by calibrating the measured P-wave amplitude, depending on epicentral distance  $\Delta$  and source depth  $h$ , with the Gutenberg and Richter (1956)  $Q(\Delta, h)$  values (see Fig. 1a in DS 3.1), the IDC amplitudes A5/2 are calibrated with the Veith and Clawson (1972) calibration function for 1 Hz vertical component P-waves, which is given for peak-to-trough amplitudes (see Figure 2 in DS 3.1).

The specifics of the IDC measurement procedure for mb has been optimally tuned to the mandate of the CTBTO. This is to assure, amongst others, the best possible detection of and discrimination between underground nuclear explosions (UNE) and natural earthquakes. It should be noted, however, that on average mb(IDC) values are systematically too small for stronger earthquakes when comparing mb(IDC) with mb resulting from NEIC or the new IASPEI (2005 and 2011) standard procedure for mb (Bormann et al., 2007 and 2009). These difference grows with magnitude. While it is negligible for mb (NEIC) < 4 it amounts already to almost -0.5 magnitude units for events with mb(NEIC) between 5.0 and 5.9. For the great 2004 Mw9.3 Sumatra-Andaman earthquake it even reached extreme values between -1.5 and -1.8 m.u., respectively (mb(IDC) = 5.7, mb(NEIC) = 7.2 and mb(IASPEI 2005) = 7.5). The reasons for the more pronounced magnitude saturation of mb(IDC) are discussed in detail in Chapter 3 as well as in Bormann et al. (2007) e.g., the filter band and length of the time window used for calculating amplitude and period.

The SBSNR amplitude for three component stations is measured on the vertical channel using the filter band 2.0 – 4.0 Hz, while at array stations the SBSNR amplitude is measured on a beam using all vertical channels with no steering with the same filter band. No period is measured for the SBSNR amplitude (as for classical ML measurements; see Chapter 3, Section 3.2.4), and the amplitude is calculated as:

$$amp = \sqrt{(MAX(STA))^2 - LTA^2}. \quad (15.1)$$

In contrast to amplitude measurement for mb, based on the maximum P-wave amplitude within the first 5.5 seconds after the first onset of P, the largest short-term-average MAX(STA) is taken over the whole (2-4Hz)-filtered wave train of the considered local event. The SBSNR amplitudes are measured in the distance range  $2^\circ < \Delta < 20^\circ$  and when the difference between estimated depth - depth error is  $< 40$  km.

The attenuation correction for ML is calculated from the formula:

$$a + b * r + c * \log_{10}(r) \quad (15.2)$$

where  $r$  is the epicentral distance (in km) and the coefficients  $a$ ,  $b$ ,  $c$  have been tailored for each station that contributes to the REB to maximize agreement between ML and mb. Each station has its own  $a$ ,  $b$ , and  $c$  values, and the values of these coefficients may change from time to time as part of tuning work to make more consistent magnitudes. The REB ML magnitude is obtained from:

$$ML = \log_{10}(amp/per) + a + b * r + c * \log_{10}(r) \quad (15.3)$$

where  $amp$  is the short-term average amplitude as it appears in the REB in nm (0-peak). It has been transformed from a short-term average value, corrected for long-term noise and measured in a 2-4 Hz bandpass. The period  $per$  in the formula above is always 1/3 sec (0.33 in the REB) for ML, as the amplitude is measured from a bandpass filtered channel (between 2-4 Hz) with a center frequency of 3 Hz. Note that for stations with instrument calibration periods different from 1 sec, the instrument calibration period will enter the formula.

Thus, the IDC ML, based on rms amplitudes being measured on a narrow-band filtered vertical component trace and on coefficients in the ML formula being deliberately tailored to assure best possible agreement between local ML and teleseismic mb, differs from the IASPEI standard ML as outlined in Chapter 3, section 3.2.7.

As mentioned in 15.1.2, the most effective technique used at the IDC for discriminating between natural earthquakes and underground nuclear explosions is the mb/Ms screening criterion. Therefore, additionally to the two body-wave magnitudes, the IDC determines for individual stations also the surface-wave magnitude  $Ms(sta)$ . It is based, as the classical NEIC and now standard  $Ms(20)$  procedure on reading the maximum vertical component surface-wave amplitude at periods between 18 s and 22 s. However, the IDC  $Ms$  values are not calibrated with the IASPEI standard Prague-Moscow calibration formula for surface-wave amplitude readings, but rather with the modified formula (18) in Rezapour and Pearce (1998). The latter avoids distance-dependent biases in  $Ms$  when applying the IASPEI standard formula to exclusively  $20 \pm 2$  s surface waves readings. The reason for these distance-dependent biases in  $Ms_{20}$  are that the IASPEI standard formula used for calibration was derived on the basis of  $(A/T)$  max readings that were not restricted to periods around 20 s but allowing for periods in a much wider range between 2 s and  $< 30$  s (see Vanek et al., 1962; Karnik et al., 1962). Indeed, observations of surface-wave  $(A/T)$  max from globally distributed earthquakes confirm that for earthquakes with magnitudes  $< 7$  and/or recorded at distances  $< 50^\circ$   $(A/T)$  max is measured mostly at periods well below 18 s, down to about 3 s (see Bormann et al., 2009). The  $Ms$  formula used at the IDC reads:

$$Ms(sta) = \log_{10}(amp/per)_{max} + B(\Delta) \quad (15.4)$$

where  $M_s(\text{sta})$  is the  $M_s$  Magnitude for a given station,  $amp$  and  $per$  are from the **amplitude** table ( $amp\text{type} = \text{ALR}/2$ , or  $\text{ANL}/2$ ).  $\text{ALR}/2$  identifies Rayleigh wave signal measurements, and  $\text{ANL}/2$  identifies noise measurements.  $amp/per$  is the maximum amplitude/period ratio within the allowed period range of 18 to 22 s (Stevens and McLaughlin, 2000).  $B(\Delta)$  is identical with the Rezapour and Pearce (1998) calibration function (which is a scaled inverse attenuation formula) for 20 s surface waves. When  $amp$  is measured in nm instead of  $\mu\text{m}$ , then this calibration function reads:

$$B(\Delta) = 1/3 \log_{10}(\Delta) + 1/2 \log_{10}(\sin \Delta) + 0.0046\Delta + 2.370, \quad (15.5)$$

where the term 0.0046 is  $\gamma$  times  $\log_{10}(e)$  ( $= 0.4343$ ), and  $\gamma$ , the attenuation coefficient.  $\gamma$  was determined empirically by Rezapour and Pearce (1998) to be 0.0105 using a very large data set.

### **Estimate azimuth and slowness**

Azimuth and slowness are measured in DFX using different techniques for three component stations (polarization analysis), arrays (f-k analysis), and for the large array (beampacking technique) as shown in Tab. 15.8.

**Tab. 15.8** Azimuth and slowness estimation methods for different station types. The filter band used for f-k analysis at arrays is dependent on the band of the best beam.

| Station type | Method       | Time Window     | Filter band  | Channel    |
|--------------|--------------|-----------------|--------------|------------|
| 3-C          | polarization | 5.5 sec         | 1.0 – 4.0 Hz | Z/N/E      |
| Array        | f-k analysis | 3.15 – 10.8 sec | ~beam*       | Z channels |
| Larray       | beampacking  | 6.0 sec         | 1.2 – 3.2 Hz | Z channels |

\*the filter band depends on the band of the best beam, i.e., the beam with the highest SNR

In polarization analysis for three component stations, polarization analysis is done over a series of overlapping windows near the detection. In each of the overlapping windows a covariance matrix is estimated for the three components of motion and then the eigenvalues ( $\lambda_1, \lambda_2, \lambda_3$ ), and eigenvectors ( $\mathbf{u}_1, \mathbf{u}_2, \mathbf{u}_3$ ) are solved for the covariance matrix. The signal rectilinearity is a measure of the linearity of the particle motion, and is calculated as:

$$rect = 1 - \frac{\lambda_3 + \lambda_2}{2\lambda_1} \quad (15.6)$$

In the overlapping window with the maximum rectilinearity for the short-period filtered P wave the incident angle  $inang$  is estimated as:

$$inang = \frac{\arccos(|u_{11}|)}{90} \quad (15.7)$$

where  $u_{11}$  is the direction cosine of the eigenvector  $\mathbf{u}_1$  associated with the largest eigenvalue.

The observed azimuth,  $seazp$ , of the signal (assuming that the signal is a P-phase) is the azimuth of the eigenvector ( $\mathbf{u}_1$ ) associated with the smallest eigenvalue:

$$seazp = \text{atan2}\left(\frac{u_{13}}{u_{12}}\right) \quad (15.8)$$

The final slowness  $s$  during polarization analysis is calculated as:

$$slowness = polar\_alpha \cdot \sin\left(\frac{inang}{2}\right) \quad (15.9)$$

where  $polar\_alpha = A/(2 \cdot B^2)$ , and  $A = P$ -wave velocity at the station site (km/sec), and  $B = S$ -wave velocity at the station site.

In f-k analysis for seismic arrays, spectra are computed from the vertical channels in a station-dependent time window. For each slowness vector, the f-k power spectrum is calculated as:

$$P(s_n, s_e) = \frac{\sum_{f=f_1}^{f_2} \left| \sum_{i=1}^J F_i(f) \cdot e^{2\pi\sqrt{-1}f(s_n \cdot dnorth + s_e \cdot deast)} \right|^2}{J \cdot \sum_{f=f_1}^{f_2} \left\{ \sum_{i=1}^J (F(i))^2 \right\}} \quad (15.10)$$

where  $deast_i$  and  $dnorth_i$  are the east-west and north-south coordinates, respectively, of the  $i$ th sensor array element relative to the reference station.

The slowness vector with the maximum f-k power is selected for estimating the azimuth and slowness values.

The beampacking technique, also called time domain f-k analysis is used for estimating azimuth and slowness for the large array NOA. This scheme is based on beamforming over a predefined grid of slowness points and measuring the power. The relative power is the signal power of the beam for the peak slowness divided by the average channel power in the same time window.

#### 15.4.2.2 StaPro

StaPro performs the following processing steps:

1. Determine initial wave type – categorize seismic detections as one of four initial wave types: noise (N), teleseismic ( $\Delta > 20^\circ$ ), regional P or regional S ( $\Delta < 20^\circ$ )
2. Group detections – place signals into groups in which each member has similar characteristics suggesting that they were generated by the same event.
3. Assign initial phase name – regional phase identification is done by examining all arrivals from the same group. Phase names for teleseismic data are limited to a simplified phase naming convention used by StaPro.
4. Locate single station events – compute a single-station event location and estimate local magnitude for association groups.

##### **Determine initial wave type**

The initial wave types at seismic arrays are mainly dependent on the velocity estimates from DFX. Tab. 15.9 shows the velocity ranges for determining the initial wave types in StaPro. There are several noise screening functions used at the IDC to recognize noise arrivals.

**Tab. 15.9** Relationship between initial wave types and velocity ranges at seismic arrays.

\*Note that the exact velocity ranges are station dependent.

| Initial wave type | Velocity range*  |
|-------------------|------------------|
| P (teleseismic)   | $\geq 11.0$ km/s |
| P (regional)      | 5.7 – 11.0 km/s  |
| S (regional)      | 2.9 – 5.7 km/s   |
| N(oise)           | $\leq 2.9$ km/s  |

The initial wave types at three component seismic stations are determined using default rules or neural networks. The default rules are shown in Tab. 15.10.

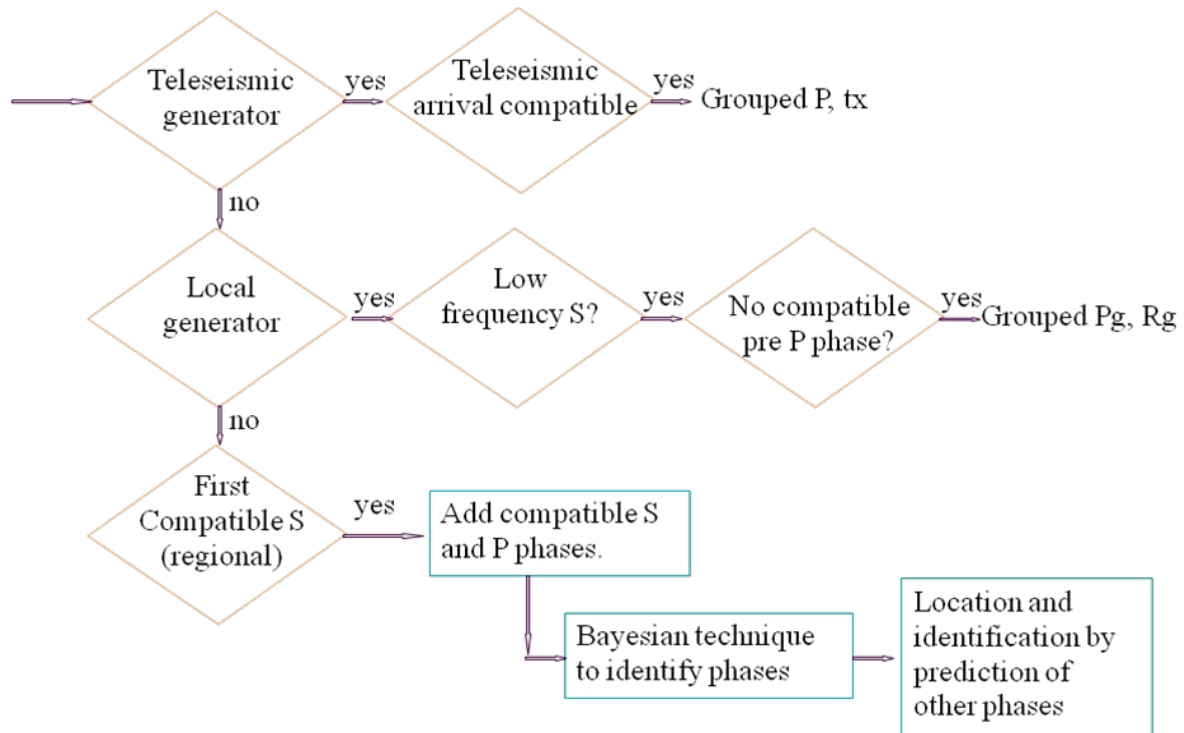
**Tab. 15.10** Default rules for determining initial wave types at 3-C stations. Rect = rectilinearity, hvrat = horizontal to vertical ratio, and freq = central frequency of the detecting channel.

| Signal Type   | Rule                                                                             |
|---------------|----------------------------------------------------------------------------------|
| Teleseismic P | $\text{rect} > 0.7$ and $\text{hvrat} \leq 1.0$ and $\text{freq} < 3$ Hz         |
| Regional P    | $\text{rect} \geq 0.7$ and $\text{hvrat} \leq 1.0$ and $\text{freq} \geq 3.0$ Hz |
| Regional S    | $\text{rect} < 0.7$ or $\text{hvrat} > 1.0$                                      |

Alternatively, neural networks can be applied to determine the initial wave types for three component stations. The neural networks use the signal period, the polarization features, and the context of the given arrival (the relative time difference and the number of arrivals in a window centered on the arrival of interest) as inputs and produce a probability for each of the four classes as output. Three neural networks are used: the first distinguishes between noise and signals, the second one between teleseismic and regional phases, and the last one for regional P and S.

### **Group detections**

Signals are grouped together by StaPro based on the compatibility of the signal type, azimuth, associated azimuth error, amplitude, and relative time difference. The logic flow of the grouping process is shown in Fig. 15.10.



**Fig. 15.10** Logic of the phase grouping algorithm in StaPro, used to group together phases from the same event.

#### ***Assign initial phase name***

The initial phase names are assigned based on the initial wave type and the grouping process. The relationship between initial wave type and initial phase name is shown in Table 11.

**Table 15.11** Initial phase names based on initial wave types.

| Initial Wave Type | Initial Phase Name |
|-------------------|--------------------|
| P (teleseismic)   | P, tx              |
| P (regional)      | Pn, Pg, Px         |
| S                 | Sn, Lg, Rg, Sx     |

#### ***Locate single station events***

Compatible phases which are grouped together can be used to form single station events. During the location process, time, azimuth and slowness values can be defining when locating the event. Each observation which is defining (time, azimuth or slowness) is assigned a certain weight, as shown in Tab. 15.12. Single station events must have a phase weight for 2.6 or more for the single station event to be saved in the automatic bulletin.

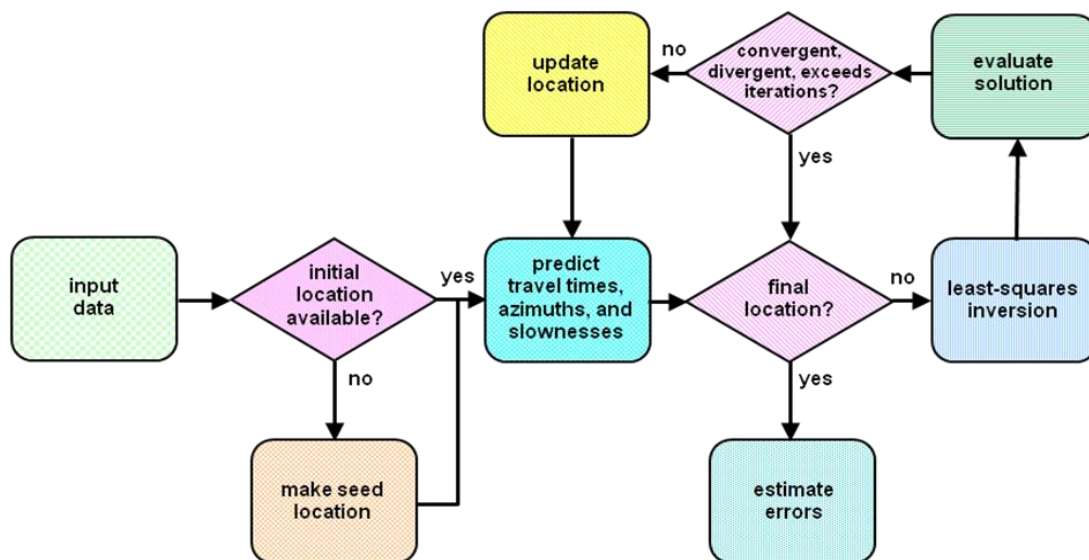


**Tab. 15.12** Seismic phase weights used in StaPro for forming a single station event at a primary seismic station.

| Phase type | Station | Time defining | Azimuth defining | Slowness defining |
|------------|---------|---------------|------------------|-------------------|
| Primary    | Array   | 1.0           | 0.5              | 0.5               |
| Secondary  | Array   | 0.7           | 0.5              | 0.5               |
| Primary    | 3-C     | 1.0           | 0.25             | 0.25              |
| Secondary  | 3-C     | 0.7           | 0.25             | 0.25              |

### 15.4.3 Event location

The determination of event hypocentral locations is a core functional element of many automatic and interactive applications within the IDC system. Hypocenters are determined via an iterative non-linear least-squares inversion. A flow chart of this process is shown in Fig 15.11.



**Fig. 15.11** Flow chart of the location process used at the IDC.

#### 15.4.3.1 Input data

Phases used in the location process are known as **defining** phases. A defining phase may have one or more of three attributes that are used in the location process: **time**, **azimuth** and **slowness**. Time-defining phases are the most common.

For seismic data azimuth and slowness defining phases are generally more useful from array stations than from 3-C stations because the standard deviation of the errors tends to be much smaller for array stations.

All infrasonic stations of the IMS are arrays, so for infrasonic data azimuth and slowness information are available for all stations. The hydroacoustic data from the IMS are processed as groups of hydrophones, so azimuth and slowness is available for these data. However, the slowness at hydroacoustic stations is treated as a constant.

### ***Use of time, azimuth, and slowness observations***

Arrival time information is often the best information to use in event determination, but it is often unavailable in sufficient quantity. In such instances, azimuth and slowness information, as obtained from f-k or 3-C analysis, can be added to obtain more reliable hypocenters with smaller uncertainties. In particular, azimuth/slowness estimates are critical when a location is required at a single station.

Pre-determined uncertainty estimates associated with arrival time information are typically better than that of azimuth/slowness information, these latter data can actually prove counterproductive to the goal of accurate event location.

In practice, the IDC analysts exclude azimuth and slowness estimates from seismic data when more than **six defining arrival times of primary stations** are available to the hypocentral location process.

### ***Establishing the initial location***

A reasonable seed (starting) location increases the likelihood of finding the global minimum. The initial locations in preferential order are:

- 1) a previously determined hypocenter;
- 2) a location based on an exhaustive grid search as determined by the GA (Global Association) subsystem;
- 3) an internal “best guess” algorithm built into the location library.

In the rare situation when a best guess seed location must be made, the strategy is to use the combination of arrival time, azimuth, and slowness information. The seed algorithm uses a list of **location scenarios** to find a starting location:

- Use the time defining S–P time at the closest seismic station and the best-determined defining azimuth from a seismic station to compute a seismic epicentral location.
- Use various combinations of defining S–P times and defining P-wave arrival times from seismic stations to compute a seismic epicentral location.
- Use the two best defining azimuths available from two different stations to define two great circles whose intersections are candidates for the seed location. Choose the location that is closest to the two stations.
- Compute a seismic epicentral location by using the best P-wave azimuth (for direction) and slowness (for distance) from a single seismic station.
- Put the epicentral location near the closest station. Use azimuth information, if available.

### ***Predicting travel-time, azimuth and slowness values***

Travel-time, slowness, and azimuth of seismic phases can all be used for locating seismic events. The IASPEI-91 travel time tables contain travel-time parameterized by distance and depth. The modeling error associated with a given phase-dependent travel-time table is also explicitly specified. Travel-time corrections include:

- Ellipticity correction
- Elevation correction
- Source-specific station correction (SSSC)

### ***Source-Specific Station Correction (SSSC)***

Source-specific station corrections to the travel time of a seismic phase attempt to correct for the travel-time effects of lateral heterogeneity. SSSC tables are specified by a two-dimensional parameterization along latitudinal and longitudinal nodes. By introducing SSSC information the phase-dependent modeling errors assumed within the one-dimensional travel-time tables should be reduced.

### ***Predicting seismic slownesses and azimuth values***

The slowness model is based on the derivative of travel time with respect to distance. Azimuth modeling is based on the exact azimuth between the station and the hypothesized event location.

Azimuth and slowness data from seismic stations are most useful for locating small events, especially when their data importances are high. Unfortunately, significant biases that can map into large event mislocations exist within these data. These biases, coupled with relatively large modeling and measurement errors, also produce very large error ellipses. The goal of the slowness-azimuth station corrections (SASC) is to remove these systematic biases from these azimuth-slowness measurements

### ***Slowness-Azimuth Station Corrections (SASC)***

The slowness/azimuth station corrections and related modeling errors are represented on a binned polar slowness/azimuth grid (Bondar, et al, 1999). A constant correction and modeling error is defined over the entire bin. A background correction and modeling error can be defined for those bins without a specified SASC. The background correction is parameterized as a vector slowness correction to adjust for large-scale structural features near the station. In addition, rotational coefficients are specified. These are to be applied to each slowness vector. The total background correction is thus an affine transformation of the slowness vector with a constant component and a rotational component.

### ***Hydroacoustic measurements***

Hydroacoustic stations with hydro-channel groups can be used to estimate azimuth (typically under the assumption of fixed slowness). Travel-times and, if available, azimuths of hydroacoustic phases and knowledge of all possible source locations for those phases (expressed as blockage grids for each hydroacoustic station) are used for locating events detected by hydroacoustic network.

The hydroacoustic travel-time model is complicated by the dynamic nature of the oceans through which the hydroacoustic phases propagate. Unlike the seismic travel-time predictions for which corrections are applied to a standard one-dimensional model, seasonally varying travel-time models are used to predict the propagation times for hydroacoustic phases. One-dimensional models are used only if the seasonal models are unavailable.

Because of the slow propagation velocity in the ocean and spatial and seasonal variations in temperature that induce velocity variations, actual travel times may depart substantially from constant-velocity travel times. To account for these spatio-temporal variations, the system provides for the use of seasonally varying, azimuth, and range-dependent travel-time tables specifically calculated for each hydroacoustic station. The travel times and their modeling uncertainty are specified on a polar grid centered at the station.

### ***Hydroacoustic blockage***

Hydroacoustic phases propagate long distances with limited attenuation in the water column. The phases are converted at the water-land interface of sizable islands or continents, but the converted waves do not propagate efficiently and are effectively blocked by the island or continental mass.

Blockage is modeled for each hydroacoustic station as the maximum distance of propagation along a set of azimuths. The blockage subsystem is not used as an integral part of the location algorithm. Rather, it is invoked directly by several applications (ARS and GA for instance) to check if a path is topographically blocked. ARS is the software used by IDC analysts to review waveform data. The Global Association software is used during automatic processing to associate phases and form events.

### ***Infrasound measurements***

All of the infrasonic stations of the IMS are arrays and thus the data from these stations yield travel-time, azimuth, and slowness values.

Infrasound does not propagate with a constant velocity in a straight line from its source to the receiver. Rather, there appear to be two basic mechanisms for transporting low frequency acoustic energy in the atmosphere. The first is via reflections between the Earth's surface and the Stratosphere at an altitude of around 50 km. The second is via reflections between the Earth's surface and the thermosphere at an altitude of around 120 km. This 'ducting' can be responsible for the horizontal transmission of infrasound over many thousands of kilometers.

Currently at the IDC the initial infrasonic phase is used during location, and is designated with the phase name I. The propagation model shown in Tab. 15.13 is used, and a large modeling error is attributed to that value. Travel times are specified as an extension of the IASPEI-91 travel time curves.

**Tab. 15.13** Infrasound wave propagation model used at the IDC.

| <b>Infrasound wave type</b>            | <b>Distance Range (deg)</b> | <b>Velocity (km/s)</b> |
|----------------------------------------|-----------------------------|------------------------|
| Direct arrival and Tropospheric regime | 0 to 1.2                    | 0.330                  |
| Stratospheric/Thermospheric regime     | 1.2 to 20                   | 0.295                  |
| Stratospheric regime                   | 20 - 60                     | 0.303                  |

### 15.4.3.2 Nonlinear least-squares inversion

Hypocentral inversion or event location is performed as a nonlinear least-squares inversion of travel time, azimuth and slowness measurements from stations at local, regional and teleseismic distances.

Input data:  $N$  observations (travel times, slownesses, azimuths);

initial guess of event location and origin time ( $lat_{org}$ ,  $lon_{org}$ ,  $depth_{org}$ ,  $t_{org}$ ).

Goal: Minimize the sum of squares of  $N$  non-linear functions (weighted residuals) that depend on a set of parameters ( $lat_{org}$ ,  $lon_{org}$ ,  $depth_{org}$ ,  $t_{org}$ ).

$$\begin{bmatrix} res_1^{tt}/\sigma_1^{tt} = t_1^{obs} - t_{org} - t_1^{pred}(lat_{org}, lon_{org}, depth_{org}, lat_{sta1}, lon_{sta1}) \\ res_1^{sl}/\sigma_1^{sl} = sl_1^{obs} - sl_1^{pred}(lat_{org}, lon_{org}, depth_{org}, lat_{sta1}, lon_{sta1}) \\ res_1^{az}/\sigma_1^{az} = az_1^{obs} - az_1^{pred}(lat_{org}, lon_{org}, lat_{sta1}, lon_{sta1}) \\ res_2^{tt}/\sigma_2^{tt} = t_2^{obs} - t_{org} - t_2^{pred}(lat_{org}, lon_{org}, depth_{org}, lat_{sta2}, lon_{sta2}) \\ \dots \\ res_N^{tt}/\sigma_N^{tt} = t_N^{obs} - t_{org} - t_N^{pred}(lat_{org}, lon_{org}, depth_{org}, lat_{staN}, lon_{staN}) \end{bmatrix}$$

where  $\sigma = \sqrt{\sigma_{model\_err}^2 + \sigma_{measure\_err}^2}$

### 15.4.3.3 Evaluating the solution

Events are located through an iterative procedure. The procedure is continued, while stable, until convergence is achieved, divergence is recognized, or the maximum number of iterations has been exceeded.

#### **Stability**

If the location adjustment at a given iteration is extremely large, then the revised location may be suspect; the stability of the solution is questionable. If the adjustment has been identified as being very large, then the first step is to scale down the adjustment. In the early iterations this means not permitting the location to move more than 3000 km in a single step. For latter iterations this step is reduced to 1500 km.

#### **Divergence tests**

Local divergence is associated with a single iteration where the location seems to be getting worse because the current adjustment vector is larger than that encountered during the previous iteration. Global divergence is established when local divergence of sufficient magnitude is identified over two consecutive iterations. A single local divergent iteration is harmless and permits the iterative process to continue. If two consecutive locally divergent steps are identified, the last more than 10% worse than that two iterations prior, then global divergence is declared, and the iterative location process is terminated with an appropriate error message.

### ***Convergence tests***

If the total spatial adjustment vector in all dimensions is less than 0.5 km coupled with an origin time change of less than 0.5 s, then the solution has converged.

### ***Maximum iteration test***

If convergence has not been achieved, iteration is terminated when the maximum number of iterations is reached. The maximum number of iterations for location is set to 60 for analyst review. A minimum of four iterations are always performed.

The iteration number is also used to control the conditions of the inversion. The depth element of hypocentral parameters is constrained (fixed) during the first two iterations.

### ***Updating hypocenters***

After all of the stability and convergence tests have been completed the hypocenter is updated based on the newly-determined adjustment solution vector. When convergence or divergence has been declared, the prediction process has to be invoked one last time so the data residuals and network statistical measures coincide with the most recently adjusted hypocenter.

### ***Estimating errors***

Estimating errors includes calculating the model covariance matrix, data resolution matrix, the resultant 90th percentile error ellipse, depth error, and origin time error.

## **15.4.4 Event definition criteria**

The IMS network records a variety of signals from events throughout the world. A criterion is applied in order to decide which events to include in the automatic and reviewed bulletins produced by the IDC. Considering the mission of the organization, to look for evidence of potential Treaty violations, the concept of the event definition criteria was introduced.

Each event recorded by the network is composed of defining phases used in the location process. Each of these defining phases is assigned a weight, and the sum of these weights is the weight of the event (Tab. 15.14). Note that defining phases at auxiliary seismic stations do not contribute to the weight of an event, but auxiliary seismic stations are used to refine the event solution. Events must have a weight of 3.55 or above to appear in the SEL1, SEL2, or SEL3 bulletins, and an event must have a weight of 4.6 or above to appear in the REB (reviewed event bulletin). This essentially means that an event must be observed by at least two primary stations to appear in the automatic bulletins (SEL1, SEL2, SEL3), and must be observed by at least three primary stations to appear in the REB. Primary stations in this case refer to primary seismic stations, hydroacoustic stations, and infrasound stations.

**Tab. 15.14** Phase weights used for event definition criteria. A primary seismic type phase is the first P-type phase (from the list P, Pn, Pg, or PKP) and the secondary seismic type phases are all subsequent defining phases.

| Phase Type        | Station Type          | Arrival Time | Azimuth | Slowness |
|-------------------|-----------------------|--------------|---------|----------|
| Primary seismic   | Primary seismic array | 1.0          | 0.4     | 0.4      |
| Secondary seismic | Primary seismic array | 0.7          | 0.4     | 0.4      |
| Primary seismic   | Primary seismic 3-C   | 1.0          | 0.2     | 0.2      |
| Secondary seismic | Primary seismic 3-C   | 0.7          | 0.0     | 0.0      |
| H                 | Hydroacoustic         | 1.2          | 0.6     | 0.0      |
| I                 | Infrasonic            | 0.8          | 1.0     | 0.0      |

## 15.5 Data availability

The data from the IMS network and the bulletins from the IDC are available to authorized users in close to real time. Each country that has signed the CTBT may designate a National Data Centre (NDC). Staff at the NDC and other institutes may be designated as authorized users.

Additionally, the Review Event Bulletin (REB) is also provided to the International Seismologic Centre (ISC), who includes the REB as one of their contributed bulletins. The ISC bulletin is freely available from <http://www.isc.org>.

There is also a new CTBTO initiative known as the Virtual Data Exploitation Centre (VDEC), which aims to make data and products more widely available to the broader scientific community. More information about this project can be obtained by sending your inquiry to the email address [vdec@ctbto.org](mailto:vdec@ctbto.org).

Moreover, many products of the former Prototype International Data Center (PIDC) are available at the Defense Threat Reduction Agency (DTRA; <http://www.dtra.mil/Home.aspx>) Verification Data Base (<http://www.rdss.info>). This website permits access to and data retrieval from an extensive (>18 Terabytes) archive of hydroacoustic, infrasound and seismic waveform data covering 1994-present, and access to an archive of event bulletins from the PIDC (1995-Sept. 2003) as well as of the IDC with several years delay (as of 2011 IDC data 2000-2004). On this website one finds data availability statistics and infrasound station metadata from stations that are part of the IMS as well as from a few selected non-IMS sites. A research data base provides access also to waveforms, signal measurements and source information drawn from a wide variety of official and unofficial sources for nuclear explosions, GT locations and infrasound events. When available, also waveform data from these events can be retrieved via AutoDRM.

## Acknowledgments

The authors would like to acknowledge the text provided by Yan Jia, a former staff member of CTBTO. Additional input was received from Mark Prior and Pierrick Mialle. This document has been improved thanks to a thorough review by Svein Mykkeltveit.

## Disclaimer

The views expressed in this paper are those of the authors and do not necessarily reflect the views of the CTBTO Preparatory Commission.

## Recommended overview readings

Barrientos et al. (2001)  
Bowers and Selby (2009)  
Dahlmann and Israelson (1977)  
Dahlmann, Mykkeltveit, and Haak (2009)  
Husebye, and Dainty (1996)  
Richards (2002)  
Richards and Wu (2011)

## References

- Akaike, H. (1974). A new look at the statistical model identification. *IEEE Transactions on Automatic Control*, **19** (6), 716–723.
- Barrientos, S., Haslinger, F., and IMS Seismic Section (2001). Seismological monitoring of the Comprehensive Nuclear-Test-Ban Treaty. *Carl Hanser Verlag, München, Kerntechnik*, **66**, 3, 82-89.
- Bondár, I., North, R. G., and Beall, G. (1999). Teleseismic slowness-azimuth station corrections for the International Monitoring System Seismic Network. *Bull. Seism. Soc. Am.*, **89**(4), 989 - 1003.
- Bormann, P., Liu, R., Ren, X., Gutdeutsch, F., Kaiser, D., and Castellaro, S. (2007). Chinese National Network magnitudes, their relation to NEIC magnitudes, and recommendations for new IASPEI magnitude standards. *Bull. Seim. Soc. Am.*, **97**, 114-127; doi: 10.1785/0120060078
- Bormann, P., Liu, R., Xu, Z., Ren, K., Zhang, L., Wendt S. (2009). First application of the new IASPEI teleseismic magnitude standards to data of the China National Seismographic Network, *Bull. Seism. Soc. Am.* **99** (3), 1868-1891; doi: 10.1785/0120080010
- Bowers, D., and Selby, N. D. (2009). Forensic seismology and the Comprehensive Nuclear-Test-Ban Treaty. *Annual Reviews of Earth and Planetary Sciences*, **37**, 209-236.
- CTBT/PC/II/Add.2 16 (May 1997). Report of Working Group B to the Second Session of the Preparatory Commission for the Comprehensive Nuclear-Test-Ban Treaty Organization, Vienna.



- CTBT/WGB/TL-11,17/15/Rev.4 (2009). Operational manual for seismological monitoring and the international exchange of seismological data (draft).
- CTBT/WGB/TL-11,17/16/Rev.4 (2009). Operational manual for hydroacoustic monitoring and the international exchange of hydroacoustic data (draft).
- CTBT/WGB/TL-11,17/17/Rev.4 (2009). Operational manual for infrasound monitoring and the international exchange of infrasound data (draft).
- CTBT/WGB/TL-11, 17-18 /Rev. 4 (2010). Operational manual for radionuclide monitoring and the international exchange of radionuclide data (draft).
- CTBT/PTS/INF.96/Rev. 7 (2005). Certification of radionuclide laboratories.
- CTBTO Noble Gas Monitoring Handbook (2003), 105 pp.
- Dahlman, O., and Israelson, H. (1977). Monitoring underground nuclear explosions. *Elsevier Scientific Publishing Company*, Amsterdam-Oxford-New York, 440 pp.
- Dahlman, O., Mykkeltveit, S., and Haak, H. (2009). Nuclear test ban – converting political visions to reality. Springer Verlag, xviii+277 pp., ISBN 978-1-4020-6883-6, doi: 10.1007/978-1-4020-6885-0.
- Gutenberg, B., and Richter, C. F. (1956). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Husebye, E. S., and Dainty, A. M. (Eds.) (1996). Monitoring a comprehensive nuclear test ban treaty. Dordrecht: *Kluwer Academic. NATO ASI series E*, Vol. 303, 247-293.
- IASPEI (2005). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf)
- IASPEI (2011). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_WG-Recommendations\\_20110909.pdf](http://www.iaspei.org/commissions/CSOI/Summary_WG-Recommendations_20110909.pdf)
- Karnik, V., Kondorskaya, N. V., Riznichenko, Yu. V., Savarensky, Ye. F., Soloviev, S. L., Shebalin, N. V., Vanek, J., and Zatopek, A. (1962). Standardisation of the earthquake magnitude scales. *Studia Geophysica et Geodaetica*, **6**, 41-48.
- Kradolfer, U. (1996). AutoDRM - The first five years. *Seism. Res. Lett.*, **67**, 4, 30-33.
- Rezapour, M., and Pearce, R. G. (1998). Bias in surface-wave magnitude  $M_S$  due to inadequate distance corrections. *Bull. Seism. Soc. Am.*, **88**, 1, 43-61.
- Richards, P. G. (2002). Seismological methods of monitoring compliance with the Comprehensive Nuclear Test-Ban Treaty. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam*, 369-382.
- Richards, P., and Wu, Z. (2011). Seismic monitoring of nuclear explosions. In: Gupta, H. (Ed.) (2011), *Encyclopedia of Solid Earth Geophysics*. Springer, Dordrecht, Vol. 1144-1156.
- Ringdahl, F., Marshall, P. D., Alewine, R. W. (1992). Seismic yield determination of Soviet underground nuclear explosion at the Shagan river test site. *Geophys. J. Int.*, **109**, 160-179.
- Stevens, J. L. and K. L. McLaughlin (2000). Optimization of surface wave identification and measurement. *Pure and Applied Geophysics*, **158**, 8, 1547 – 1582.

- Vaněk, J., Zapotek, A., Karnik, V., Kondorskaya, N.V., Riznichenko, Yu.V., Savarensky, E.F., Solov'yov, S.L., and Shebalin, N.V. (1962). Standardization of magnitude scales. *Izvestiya Akad. SSSR., Ser. Geofiz.*, **2**, 153-158.
- Veith, K. F., and Clawson, G. E. (1972). Magnitude from short-period P-wave data. *Bull. Seism. Soc. Am.*, **62**, 435-452.
- Xenitellis, S. (2000). The Open-source PKI Book. <http://ospkibook.sourceforge.net>

# Chapter 16

## Automated Event and Phase Identification

(Version August 2011; DOI: 10.2312/GFZ.NMSOP-2\_ch16)

Ludger Küperkoch <sup>1)</sup>, Thomas Meier <sup>2)</sup>, and Tobias Diehl <sup>3)</sup>

- <sup>1)</sup> Ruhr-Universität Bochum, Institut für Geologie, Mineralogie und Geophysik, Universitätsstr. 150, 44780 Bochum  
Now at: BESTEC GmbH, Oskar-von-Miller-Str. 2, 76829 Landau, E-mail:  
[kueperkoch@bestec-for-nature.com](mailto:kueperkoch@bestec-for-nature.com)
- <sup>2)</sup> Universität Kiel, Institut für Geophysik, Otto-Hahn-Platz 1, 24118 Kiel, Fax: +49 431 880-4432,  
E-mail: [meier@geophysik.uni-kiel](mailto:meier@geophysik.uni-kiel)
- <sup>3)</sup> Swiss Seismological Service, ETH Zurich, Sonneggstrasse 5, 8092 Zürich, Switzerland, E-mail:  
[tobias.diehl@sed.ethz.ch](mailto:tobias.diehl@sed.ethz.ch)

|                                                                                                                               | page |
|-------------------------------------------------------------------------------------------------------------------------------|------|
| <b>16.1 Introduction</b>                                                                                                      | 1    |
| 16.1.1 General remarks                                                                                                        | 2    |
| 16.1.2 Historical overview                                                                                                    | 4    |
| <b>16.2 Phase and event detection</b>                                                                                         | 5    |
| <b>16.3 Automated P-onset determination</b>                                                                                   | 7    |
| 16.3.1 The Allen picker                                                                                                       | 7    |
| 16.3.2 The Baer and Kradolfer picker                                                                                          | 14   |
| 16.3.3 P-onset determination using Higher Order Statistic                                                                     | 17   |
| 16.3.4 The AR-AIC picker                                                                                                      | 23   |
| 16.3.5 Discussion of presented P-picking algorithms                                                                           | 30   |
| <b>16.4 Automated S-onset determination</b>                                                                                   | 30   |
| <b>16.5 Automated quality assessment, phase identification, and outlier detection</b>                                         | 37   |
| 16.5.1 Quality assessment of phase times                                                                                      | 38   |
| 16.5.2 Phase identification                                                                                                   | 39   |
| 16.5.3 Post-picking and outlier detection                                                                                     | 43   |
| <b>16.6 Practical considerations: Implementation, calibration, and pitfalls of automatic detection and picking procedures</b> | 44   |
| 16.6.1 Calibration and test of automatic pickers                                                                              | 45   |
| 16.6.2 Pre-processing: wave filters                                                                                           | 46   |
| 16.6.3 Pre-processing: waveform quality                                                                                       | 48   |
| <b>Appendix</b>                                                                                                               | 48   |
| <b>Acknowledgments</b>                                                                                                        | 48   |
| <b>References</b>                                                                                                             | 48   |

## Introduction

In Chapters 2 and 11 the complexity of seismic waveforms is explained in detail. Much experience is needed to identify the seismic phases correctly and to pick onset times accurately and consistently. The majority of our current knowledge of the Earth's seismicity

is based on manual picking of onset times. In addition, manual picks have been used for the investigation of the structure of the Earth. However, for early warning purposes and for the processing of large data sets automatic picking is needed. Because of the variability of seismic waveforms, the occurrence of different seismic phases and the presence of noise, automatic picking remains still a challenge. Yet, in the light of the rapidly growing amount of digital waveform data produced by permanent and temporary networks, the issue of automatic event detection and phase picking becomes more and more important. Accordingly, the number of published algorithms has increased significantly during the past years. This makes it difficult for potential users to keep track of the different approaches and developments, even more, since most algorithms have been developed for a certain data set or a particular problem, such as early warning, real-time location, tomography, etc. Accordingly, only few algorithms were widely established in the community of observational seismology. Generalization of essential principles and criteria as well as large-scale comparative benchmark tests of different existing algorithms are still extremely rare. This makes it difficult for interested developers to choose the most appropriate approach, especially when aiming at multi-task high-quality performance, and for users to select the most suitable algorithms among the many available techniques for their specific problem. With this Chapter we intend, therefore, to provide a broader introduction into the problem of automated event detection and phase identification, to outline the underlying theory and methodological approaches, to illustrate results produced by different procedures, to assess the performance of available algorithms, to sketch the likely development in the near future and to highlight essential practical considerations for the application of automatic picking procedures.

### **16.1.1 General Remarks**

With the advent of digital real-time acquisition systems automated real-time event detection and location became feasible. Beginning around 1975, High Gain Long Period (HGLP) stations, including both digital and optical recordings, were deployed by the Seismic Research Observatories (SRO). WWSSN-stations were digitally upgraded (DWWSN), which constitute the Global Digital Seismic Network (GDSN) (e.g. Lay and Wallace, 1995). At the same time, rapid advances in computer technology enabled sophisticated analysis of increasing amount of seismic data. Nowadays, several digital, global seismic networks are in operation, monitoring continuously the global seismicity and providing rapid locations of earthquakes within minutes. Automated rapid and robust detection and location of earthquakes using real-time data of regional and global networks is essential for earthquake early warning systems. Usually, only automated first onset picks are used. A large number of automated P picks is required to ensure the robustness of the automated location. Fast automated locations are replaced after some time by more accurate manual locations that use also picks of later phases.

The need for automatic picking algorithms arises also from the increasing amount of digital data sets produced by modern passive seismic networks. Even local, temporary networks, as needed for reservoir characterization during stimulation experiments at enhanced geothermal systems (EGS) or hydrocarbon reservoirs, monitor ten-thousands of seismic events at high sampling rates. For tomographic studies, which usually require data merged from several seismological networks, highly accurate and consistent phase arrival times are needed. Automated post-processing algorithms have been developed to determine and select consistently high-quality picks for tomographic purposes whose quality might be even better than those of manual phase readings. The examples given above are typical applications of

automated seismic event detection and automated arrival time estimation. Automated processing schemes show the following potentials:

- fast and near-real time data processing;
- consistent arrival time estimation;
- processing of large data sets;
- if implemented, consistent onset quality estimation and phase identification.

The algorithms to choose depend on the specific application. Of course, the more precise and powerful the automated onset determination is, the more computationally expensive the applied algorithms are. For rapid earthquake location estimates or phase identification simple STA/LTA detections may be sufficient. E.g., Earle and Shearer (1994) use STA/LTA ratios taken from a smoothed envelope function, determined using a Hilbert transformation of the seismogram, to detect first and later arrivals. These algorithms are usually referred to as *phase detectors* in contrast to more accurate arrival time estimation algorithms, which are usually referred to as *phase pickers* (Baer and Kradolfer, 1987) and needed for precise location of earthquakes or tomographic studies. There is a considerable amount of automated picking algorithms available. Usually, these algorithms are optimized for certain requirements. Comparative studies and calibration tests are not yet common practice. While the STA/LTA detector is described in detail in Information Sheet IS 8.1, we here review the most widespread automatic picking algorithms and analyze their properties. An introduction to automatic phase identification is given.

Also the accuracy of manual readings is limited due to sampling rate, noise, or the occurrence of emergent onsets. These uncertainties must be added to the theoretical time resolution of modern, GPS-based broadband seismic networks, which allow precisions of P-onset readings within 0.2 s (e.g.; iP phases in teleseismic events; Leonard, 2000). Further problems are the inconsistency of manual picks, phase misidentification and incomplete documentation of applied filters. Douglas et al. (1997) compared P-wave readings of explosion and earthquake recordings. He showed that errors of manual picks are in the order of 0.1 seconds for explosions and about 0.5 seconds for teleseismic earthquakes with a magnitude range from 4.6 to 6.1. Using data of the Montana Bureau of Mines and Geology (MBMG) network, Zeiler and Velasco (2009) investigated manual picks of local and regional first arrivals. They found that the pick error is 0.1 seconds for phase measurements with high signal-to-noise ratios. Furthermore, the manual readings were biased towards late picks. Interestingly, in a second study using data collected by the International Seismological Center (ISC), they found that picks by different institutions may be inconsistent though individual institutions pick consistently. The standard deviation from the average arrival time determined from readings of several institutions for one single event varied between 0.2-0.6 seconds. For location purposes, the pick uncertainty should be significantly lower than the RMS-travel time residuals due to wave propagation effects not accounted for. From this point of view it is obvious that automatic phase picking has the potential to improve the consistency of arrival time estimations if it is supplemented by thorough quality estimations.

In the entire processing scheme from automated event detection to automated event location, usually a phase detector is applied to all available recordings first. Then the consistency of these detections is checked in order to detect an event. After a seismic event has been recognized, the more precise phase picker is applied to the time series to obtain arrival time estimations for a robust earthquake location. Following this basic processing scheme, section

16.2 gives an introduction to event detection based on phase detectors applied to a network of stations. In section 16.3 the following established P-phase picking algorithms are described: (1) the Allen-picker, (2) the Baer- and Kradolfer-picker, (3) picking based on Higher Order Statistics (HOS) and (4) Autoregressive-Akaike-Information-Criterion-picker (AR-AIC). Picking of later arrivals is briefly introduced in section 16.4. Automated algorithms for quality estimation and phase identification as well as outlier detection are discussed in section 16.5. Practical considerations on automatic picking, pre-processing and calibration are given in section 16.6.

The following section gives a brief overview on existing phase detection and picking algorithms. As the algorithms are presented in chronological order, this sub-section serves as a short historical overview on the development of automated picking algorithms.

### 16.1.2 Historical overview

One of the first mathematically based signal detectors was the one proposed by Freiburger (1963), who applied an approximate comparison of spectral densities for the detection of Gaussian signals in Gaussian noise. This method is suitable for detecting signals rather than measuring signal onset times. Stewart (1977) developed an automated procedure for P-phase detection, P-phase processing and coda processing for local seismic event analysis in central California. Using three moving windows for computing "moving-time noise averages" from the original seismic trace and its first difference, it is tested, whether the seismic station is operating within acceptable limits of noise or not. A P phase is detected, if the threshold exceeds 2.9 times the noise level. Goforth and Herrin (1981) developed an automatic seismic signal detector based on the Walsh transform, which is quite similar to the Fourier transform but computationally less expensive. Michael et al. (1982) used this approach to develop a real-time event detection and recording system for the MIT Seismic Network. Joswig (1987) proposed a pattern recognition technique using characteristic event features in spectrograms. However, the precision of these algorithms is limited.

A fundamental step towards automatic phase-onset determination was the algorithm proposed by Rex V. Allen (1978, 1982). He introduced the concept of the characteristic function (CF), resulting from non-linear transformations of the seismic trace to which a picker is applied. Allen's CF is based on short-term-average to long-term-average ratios (STA/LTA) calculated from an approximative squared envelope function of the seismogram. This picking algorithm is still frequently applied and used for automatic picking e.g. by the USGS Earthworm system (Johnson et al., 1995). Baer and Kradolfer (1987) developed an automatic phase picker by slightly changing Allen's envelope function and incorporating a dynamic signal threshold. This algorithm marks a milestone in automated phase picking and is still frequently used, e.g. in the Programmable Interactive Toolbox for Seismological Analysis (PITSA, Scherbaum and Johnson, 1992) and MannekenPix (Aldersons, 2004). Furthermore, in combination with a sophisticated quality assessment this algorithm yields high-quality onset times useful for tomographic studies (e.g. Di Stefano et al., 2006, Diehl et al., 2009b). Higher order statistics are proposed e.g. by Saragiotis et al. (2002) and Küperkoch et al. (2010) where skewness and kurtosis are calculated in sliding windows generating the CF. These algorithms provide very reliable P-arrival time estimates.

Beside these time and frequency domain approaches, model oriented algorithms became quite common, too. Autoregressive (AR) techniques are widely used. Based on the Akaike

Information Criterion (AIC), Takanami and Kitagawa (1988) developed a procedure for the fitting of a locally stationary autoregressive model to seismograms. They implemented this procedure in an on-line system and called it FUNIMAR (fast univariate case of minimum AIC method of AR model fitting). Leonard and Kennett (1999) propose an autoregressive method that detects increases in the AR-model order due to the higher complexity of signals compared to preceding noise. The standard autoregressive two-model Akaike Information Criterion (AR-AIC, e.g. Sleeman and van Eck, 1999) estimates the AR coefficients from predefined noise and signal windows.

Gentili and Michelini (2006) propose an artificial neural network approach for P- and S-phase onset time determination, called innovative model of neural network (IUANT2). They use variance, skewness, kurtosis and a combination of skewness and kurtosis and their time derivatives.

Many automatic phase-detection algorithms incorporate several approaches, using the different advantages of the applied methods (e.g. Zhang et al., 2003; Bai and Kennett, 2000). Nippres et al. (2010) applied STA/LTA picker, higher order statistics (Saragiotis et al. 2002) and damped predominant period  $T^{pd}$  (Hildyard et al., 2008; Hildyard and Rietbrock, 2010) picker to ANCORP data to estimate P- as well as S-arrival times.

Algorithms for the estimation of relative travel times instead of absolute ones have been proposed in order to improve the picking accuracy. Examples are multi-station and array approaches using cross-correlation methods (VanDecar and Crosson, 1990) or adaptive stacking techniques (e.g. Rawlinson and Kennett, 2004; Rowe et al., 2002). These methods require high waveform coherence at neighboring stations and high signal to noise ratios as is observed for example in the case of low-pass filtered teleseismic waveforms.

## 16.2 Phase and event detection

In order to obtain first rough P-phase arrival time estimates, a single-station detector - e.g. a simple STA/LTA trigger (see IS 8.1) - is applied to all available continuous data streams of a seismic network. An analog or digital version of this detector might be implemented directly at the stations. Then data in a short time interval following the detection are transmitted to the data center for further analysis. Nowadays, usually continuous data streams are transferred to the data centers where a phase detector is applied to the incoming real-time data. A theoretical justification for the STA/LTA trigger based on the logarithm of the likelihood ratio can be found in Basseville and Nikiforov (1993).

Event detectors are configured so that the number of false detections is minimized. On the other hand the detector has to be sensitive enough in order to detect also smaller events. Therefore, any phase detector yields a considerable number of false detections and not all P phases are detected. Hence, the consistency of the detections at different stations has to be checked before the detection of an event is declared. For local seismic networks it is sufficient that within a short time interval of a few seconds P phases are detected at a certain number of stations. The time of the phase detection  $\hat{T}_i$  at station  $i$  is interpreted as an estimation of the P- phase arrival time, which is, of course, afflicted with an error  $\varepsilon_i$ .  $\hat{T}_i$  may be written as:

$$\hat{T}_i = T_0 + t_i + \varepsilon_i,$$

where  $T_0$  is the source time and  $t_i$  is the travel time of a P wave to station  $i$ . The coincidence trigger detects an event, if for any combination of a minimum number of stations (typically three or four) the condition

$$|\hat{T}_i - \hat{T}_j| \leq \varepsilon$$

is met.  $\varepsilon$  is the maximum allowed difference between trigger times at neighboring stations. This coincidence trigger works satisfying for local networks, where the number of stations and the aperture of the network are not large. For regional and global networks this simple event detection algorithm has to be modified. Such a modified algorithm may be formulated as a grid search procedure. At every knot of a 3D grid a hypothetical hypocenter is assumed. The index  $k$  is introduced for a hypothetical hypocenter.  $T_{ki}$  denotes the expected P-phase arrival time at station  $i$ . The expected difference between arrival times at two stations for the assumed hypocenter  $k$  is

$$\Delta T_{kji} = T_{ki} - T_{kj} = t_{ki} - t_{kj},$$

where  $t_{ki}$  is the travel time from the hypocenter  $k$  to station  $i$ , that is in practise calculated using a reasonable background velocity model. The expected travel time difference  $\Delta T_{kji}$  may serve as a condition that an event is detected at the hypothetical hypocenter  $k$ . If for a certain number of detections  $\hat{T}_i$  occuring in the time interval  $[t - \Delta t, t]$  the condition

$$|\hat{T}_i - \hat{T}_j - \Delta T_{kij}| \leq \varepsilon$$

is met, an event at the hypocenter  $k$  is declared. Only trigger times at predefined subsets of stations in the vicinity of the hypocenter  $k$  need to be checked in order to detect an event at hypocenter  $k$ . This algorithm is fast and yields robust preliminary event locations even if the single-station phase detector produces a large number of false detections. This algorithm is implemented i.e. in the Earthworms phase associator “Binder” (Dietz, 2002).

A similar approach is proposed by Le Bras et al. (1994) in the widely used system Global Association (GA), where explicitly identified phases are assigned to synthetic earthquakes in overlapping circular grid cells with a complete global coverage. The performance of this algorithm relies on the correct phase identification, which is based on a combination of slowness and f-k analysis, polarization analysis, and frequency content.

Waveform correlation is used to identify seismic events. For finite-length time series from STA/LTA-triggered systems, Withers et al. (1998) propose the Local Waveform Correlation Event Detection System (LWCEDS). From observed waveforms envelopes are calculated using STA/LTA energy ratios and correlated with pre-calculated travel-time curves, which are transformed into processed time series by applying an envelope function to the arrival times for each distance bin. A single value for each station is summed, and the result is normalized by the number of stations, giving the final correlation value. After the summation over all given grid points and time intervals, respectively, the maximum sum is determined and an event is declared, if this maximum sum exceeds a certain threshold.



A grid search algorithm, which works without onset time detections, is proposed by Kao and Shan (2004). Within this Source-Scanning Algorithm (SSA) a brightness function is calculated by summing the absolute, normalized amplitudes observed at all stations at their predicted arrival times, i.e.

$$br(k, T) = \frac{1}{N} \sum_{i=1}^N |u_i(T + t_{ki})|$$

where  $u_i$  is the normalized seismogram at station  $i$ ,  $t_{ki}$  is the predicted traveltime from point  $k$  to station  $i$  of a particular phase with the largest observed amplitude (on regional scale S phase). The brightness is calculated of a point  $k$  at a specific time  $T$  and varies from 0 to 1. The brightness becomes 1 if all the largest amplitudes originate from a source at point  $k$  and time  $T$ . The spatial and temporal distribution of sources is identified by a systematic search throughout the model space and time for the maximum brightness. For each source, this scanning algorithm results into a center of maximum brightness. Waveform based association algorithms may also be applied to the detection and location of tremors. The SSA algorithm was successfully applied to locate non-volcanic tremors in the northern Cascadia subduction zone (Kao and Shan, 2004).

In the case of earthquake data, the event detection is usually followed by a more precise automatic picking of P and optionally also S phases that allow to improve the event location.

## 16.3 Automated P-onset determination

After an event has been identified, an interval of the time series containing the seismic signal is usually cut out for detailed processing. For a precise event location, a picking algorithm is applied to the data to estimate P- and S-phase arrival times, respectively. In this section we review the most frequently applied and wide spread P-picking algorithms. We describe the algorithms and show applications to synthetic as well as to real examples in order to demonstrate properties and the performance of the picking algorithms. All pickers are applied to the same local, regional and teleseismic event example waveforms for reasons of comparison. Furthermore, the behaviour of the corresponding CFs are tested on synthetic data with instantaneous changes in amplitude, frequency, and phase, respectively, as the arrival of a body wave may be indicated by one or several of those changes. The synthetic examples are designed to investigate the properties of the CF rather than to simulate a P-wave arrival.

### 16.3.1 The Allen picker

Allen (1978, 1982) introduced the concept of *characteristic function* CF, where the "character" of the seismic trace is specified. The CF is obtained by one or several non-linear transformations of the seismogram and should increase abruptly at the arrival time of a seismic wave. In addition to the calculation of the CF the next steps of a picking algorithm are the estimation of the arrival time from the CF and the quality estimation.

Let  $x_i$  be the time series under investigation with first difference  $\Delta$ , Allen defined the following envelope function  $E(t)$ :

$$E_i = x_i^2 + C_i \cdot \dot{x}_i^2, \quad (16.1)$$

where  $C_i$  is a weighting constant with

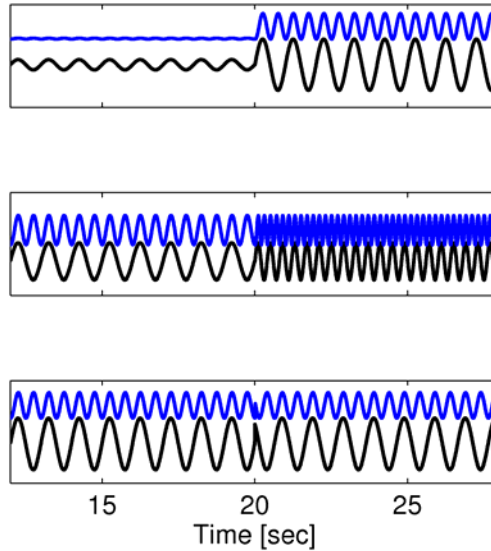
$$C_i = \frac{\sum_{j=1}^i |x_j|}{\sum_{j=1}^i |x_j - x_{j-1}|} \quad (16.2)$$

to control the relative contributions of amplitude and derivative. For a harmonic  $x_i$ , equation (16.2) reduces approximately to

$$C_i = \frac{1}{f_i dt}, \quad (16.3)$$

where  $f$  is the frequency and  $dt$  the sampling frequency. If  $C_i$  in eq. (16.1) was squared, this envelope function would be an effective, recursive approximation of the squared envelope. That means the envelope function eq. (16.1) represents a fast but rough approximation of the waveform envelope.

In the context of this article, we refer to the definition of the CF by Baer and Kradolfer (1987), who defined the CF as the time series, to which the picker is applied. Note, that Allen (1978) uses the term CF slightly differently. According to his notation the envelope function  $E(t)$  represents the CF, though the picker is applied not to this function but to the STA/LTA ratio calculated using the envelope function. In order to illustrate properties of  $E(t)$  it is calculated for synthetic data with sudden changes in amplitude, frequency, and phase, respectively. From Fig. 16.1 it becomes obvious that only changes in amplitude are detected by Allen's CF.



**Fig. 16. 1** Allen's approximated squared envelope function (blue) for synthetic data (black) with a change in amplitude (top), change in frequency (middle) and change in phase (bottom). envelope function. That means  $E(t)$  is sensitive to sudden changes in amplitude and frequency.

The following sophisticated algorithm is applied to  $E(t)$  in order to obtain an arrival time estimate and to check the reliability of the pick. At first an STA/LTA criterion is used to detect possible arrival times. Then the duration of the signal is considered in order to distinguish between noise and P-wave arrivals and to reduce the number of false alarms. In the following description of the Allen picker we refer to the Fig. 16.2 and 16.3:

A short-term average ( $\alpha$ ) and a long-term average ( $\beta \approx 100 \cdot \alpha[s]$ ) are calculated from the envelope function. If the STA/LTA ratio exceeds the reference level  $\gamma$ , the time is provisionally stored as a pick and the reference level  $\gamma$  is frozen (Fig. 16.2c). The picker now confirms or rejects the provisionally onset time by investigating the amplitudes of the raw seismogram. If low amplitudes (i.e. STA values exceed the continuation level  $\delta$ ) prevail, the picker assumes a short-term increase of noise and rejects the provisionally onset. If higher amplitudes prevail (i.e. STA values above the continuation level  $\delta$ ), the picker assumes a real seismic signal and estimates the length of this signal. To distinguish between short-term increases of noise and the seismic signal, the algorithm searches for the next zero crossing in the seismic trace. Reaching the next zero crossing, the picker starts counting the number  $M$  of observed peaks.  $M$  is incremented by 1 at each zero crossing. Using the number of observed peaks  $M$ , the continuation criterion  $\delta$  is determined by  $\delta = \beta(j) \cdot M$  and a "termination number"  $L = 3 + 3/M$  (Fig. 16.2d,e) is calculated.  $\delta$  and  $L$  are constantly increasing functions and serve as parameters to identify the length of the signal and hence to confirm the provisionally pick. In addition, also small amplitudes are counted: the short-term average  $\alpha$  is compared with the continuation criterion  $\delta$ . If  $\alpha$  exceeds  $\delta$ , a counter  $s$ , which Allen refers to as a "small count counter" and which counts the number of successive zero crossings occurring since  $\alpha$  drops below  $\delta$ , is reset. If  $\alpha$  does not exceed  $\delta$ ,  $s$  is incremented by 1. If  $s$  becomes larger than the termination number  $L$ , the event is supposed to be over, otherwise the processing goes on until  $s > L$ . The length of the time interval for which  $L > s$  serves as an estimate of the signal length (Fig. 16.2e). From Fig. 16.2d,e it is obvious, that the small count counter  $s$  increases significantly steeper than the termination number  $L$  as soon as the short-term average  $\alpha$  remains below the continuation criterion  $\delta$ . If the signal length exceeds a certain threshold  $t_{min}$ , the signal is supposed to be a seismic event, the pick is stored and optional post-processing starts. If the signal length is too short, the pick is removed,  $s$ ,  $L$ ,  $M$  and  $\delta$  are reset and a new reference level  $\gamma$  is calculated.

In order to account for automatic quality and error assessment, Allen introduced a weighting scheme, based on the seismogram and the corresponding CF. The estimated weights of the determined P onsets may serve as input for the location routine HYPOINVERSE (2002, 2003), where weight-0 onsets denote excellent or impulsive (100 % weight), weight-1 very good (75 % weight), weight-2 good (50 % weight) and weight-3 intermediate onsets (25 % weight). Weight-4 picks are not used for location. The information needed are

- $B$ , a measure of the noise level at the detection time;
- $A_0$ , the trace amplitude at the detection time;
- $D$ , the trace first difference at the detection time;
- $A_1, A_2, A_3$ , the first three amplitude peaks.

For a weight-0 P onset ("excellent"), the detection has to meet the following criteria (after Allen, 1978):

1.  $D > \sqrt{B}$ ,
2.  $A_1 > 450$ ,
3.  $A_1/\sqrt{B} > 4$ ,
4.  $A_2 > 6\sqrt{B}$  or  $A_3 > 6$ .

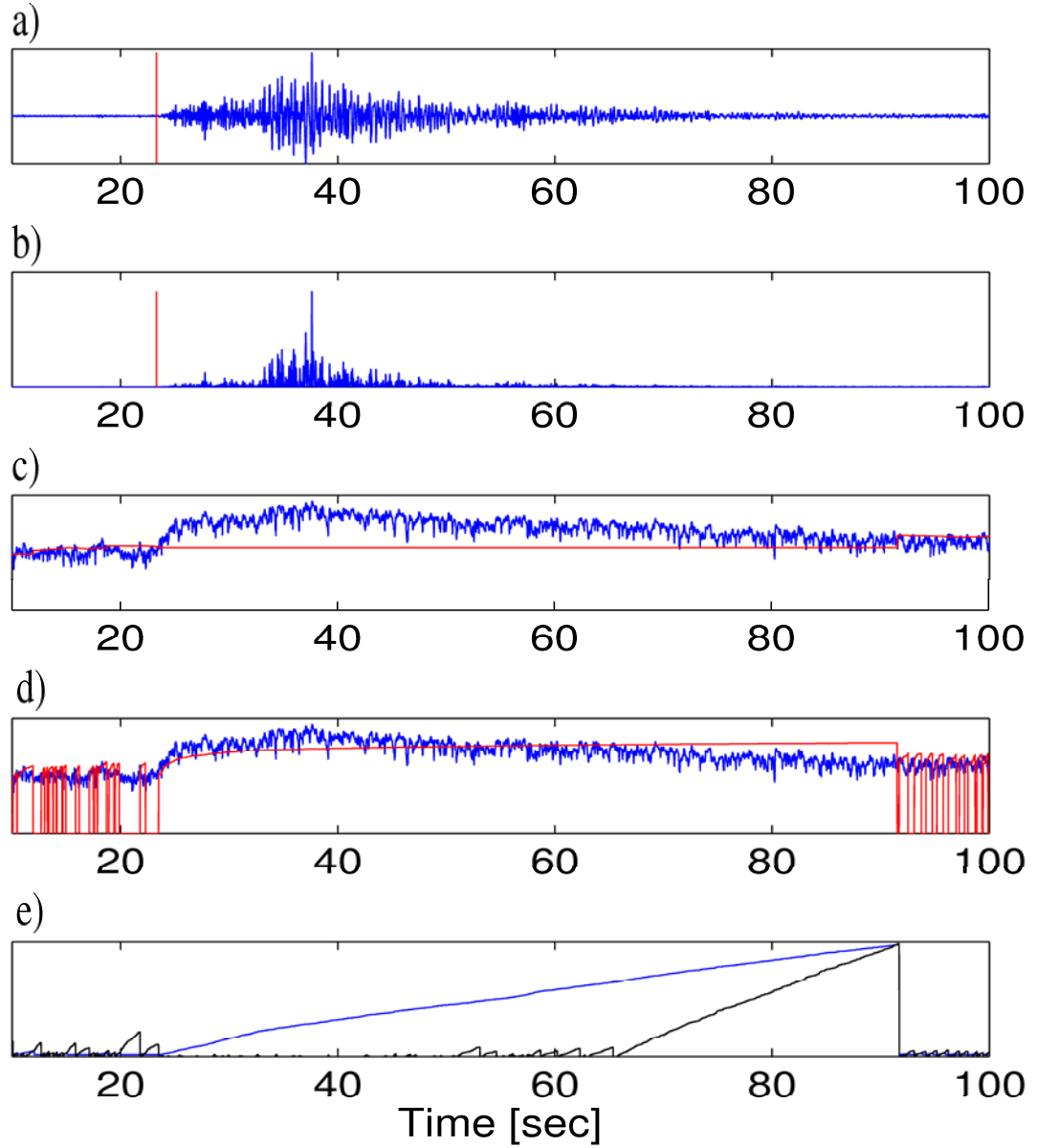
These criteria are successively to be relaxed to obtain lower weights 1,2 and 3.

**Tab. 16.1** Parameters to be adjusted for the Allen picker. The outer right columns represent the values of the parameters used for the example waveforms in Fig. 16.4.

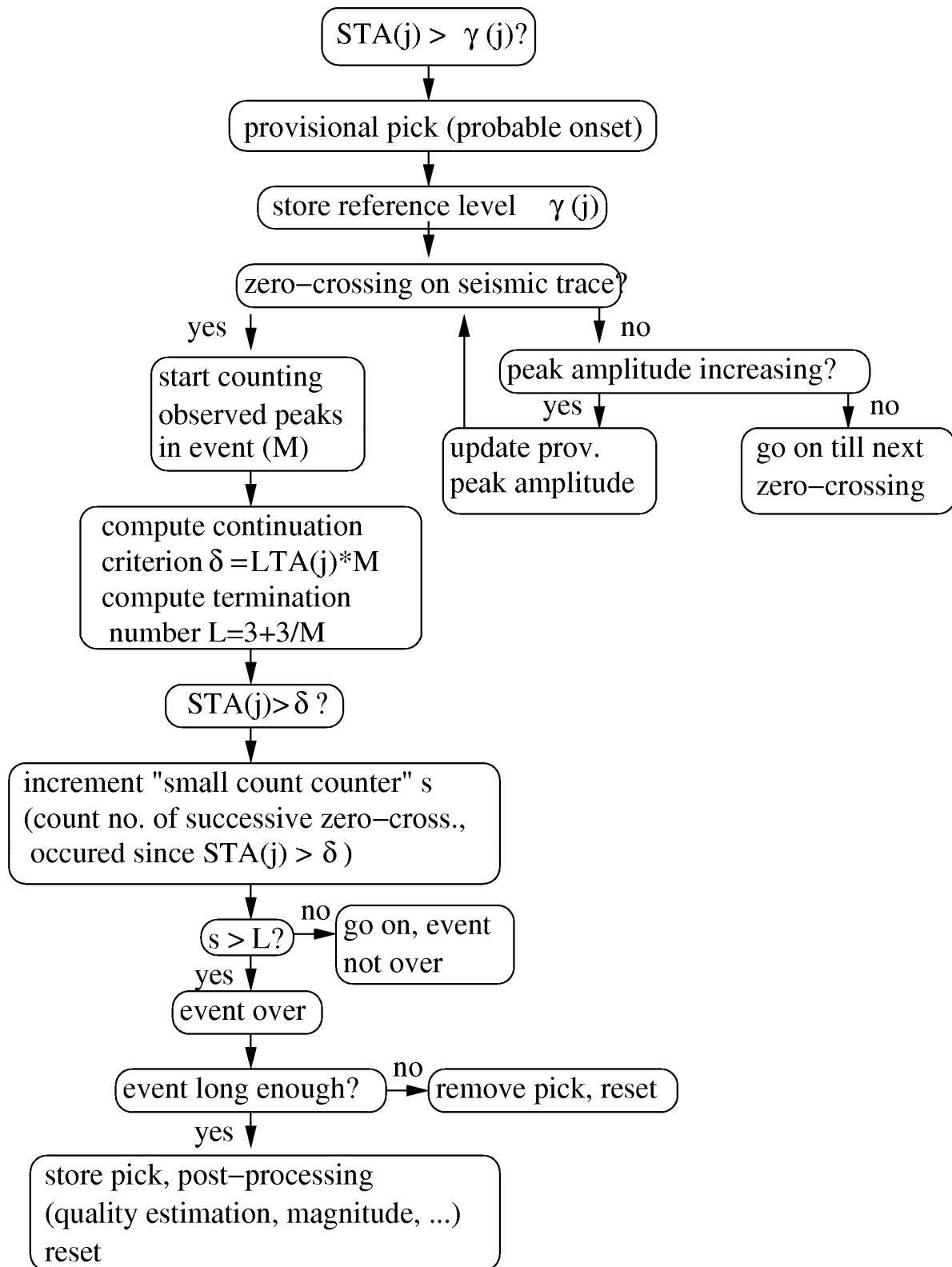
| Parameter | Remark                         | Values                  | Local | Regional | Tele |
|-----------|--------------------------------|-------------------------|-------|----------|------|
| $\alpha$  | short-term-average of CF [s]   | $0.01 < \alpha \leq 10$ | 0.1s  | 0.1s     | 10s  |
| $\beta$   | long-term-average of CF [s]    | $2 < \beta \leq 50$     | 5s    | 5s       | 50s  |
| C3        | weighting of STA values        | $0.2 < C3 \leq 0.8$     | 0.2   | 0.2      | 0.8  |
| C4        | weighting of LTA values        | $0.005 < C4 \leq 0.05$  | 0.005 | 0.005    | 0.05 |
| C5        | $LTA * C5 = \delta$            | $\sim 5$                | 2     | 2        | 3    |
| tmin      | minimum signal length required | 1.5 [s]                 | 3s    | 3s       | 40s  |

Fig. 16.4 shows applications of the Allen picker to waveforms of local, regional and teleseismic events. As shown by e.g. Küperkoch et al. (2010), this picking algorithm tends to pick somewhat early compared to an experienced analyst. Nevertheless, the Allen picker is a very robust and reliable algorithm. This sophisticated picking algorithm exploits informations provided by the CF as well as by the filtered seismogram. An advantage of the algorithm is that an automatic quality estimation of the P onset is implemented. The speed of the algorithm makes it also suitable for real-time picking of P phases.

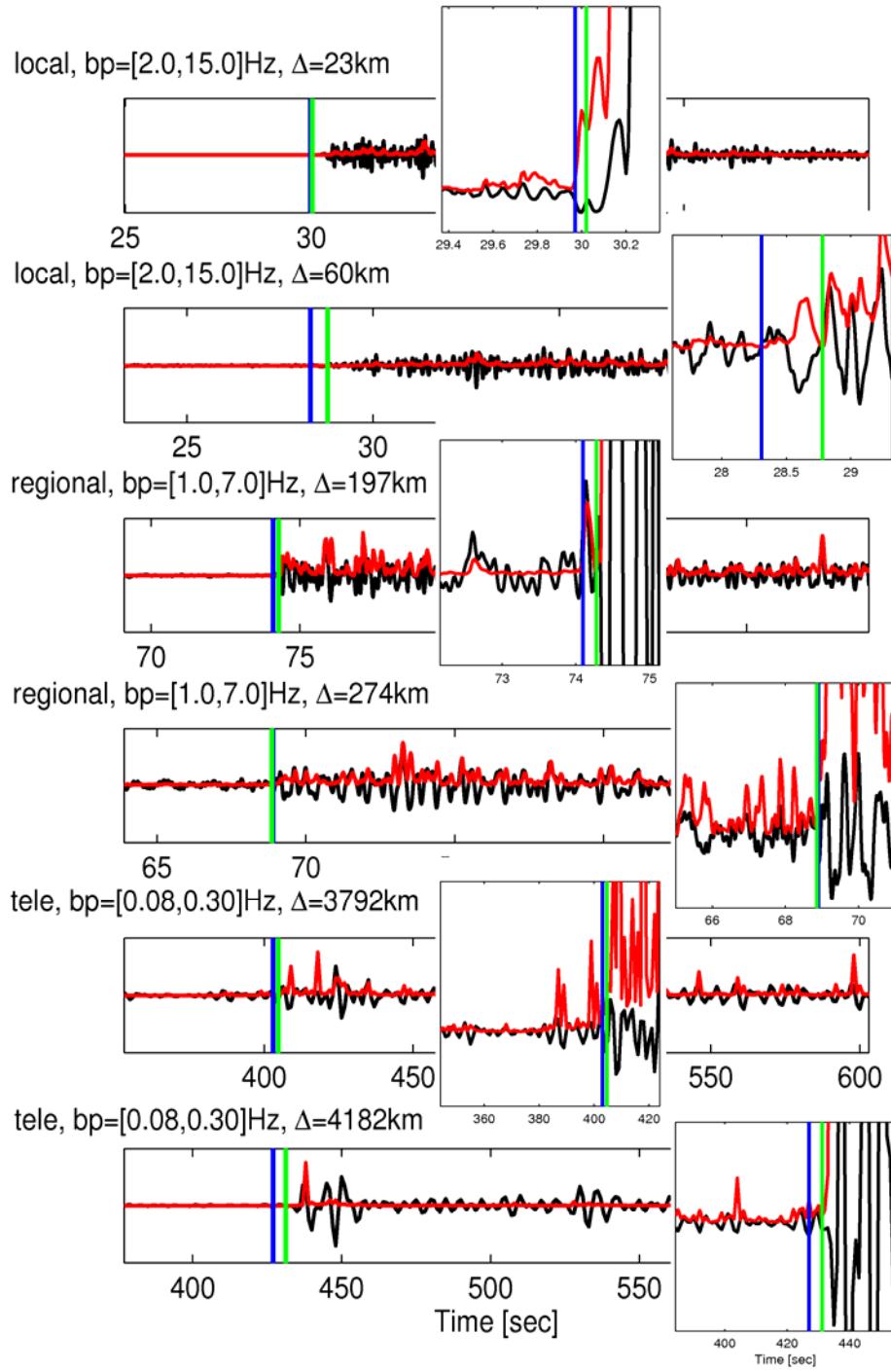
When applying this algorithm, the parameters listed in Tab. 16.1 have to be tuned. These parameters are, of course, different for local, regional and teleseismic events and hence depend on the sampling frequency and the applied filter. The outer right columns in Tab. 16.1 show the values used for the presented example waveforms in Fig. 16.4, while the parameter ranges given in column “Values” are taken from Allen (1978).



**Fig. 16.2** Visualization of parameters and thresholds needed for the Allen picker. a) Bandpass filtered (Butterworth 3<sup>rd</sup> order, 2-10 Hz) vertical component seismogram (blue) with automatically estimated P onset (red). b) Corresponding approximated, squared envelope function (blue) with estimated P onset (red). c) STA values of Allen's envelope function (blue) and corresponding reference levels  $\gamma = LTA \cdot C5$  (red). d) STA values of Allen's envelope function (blue) and corresponding continuation criterion  $\delta$  (red). e) Termination number  $L$  (blue) and number of observed zero crossings with drops below the continuation criterion  $\delta$  (black). When the termination number  $L$  and the number of observed zero crossings with  $STA < \delta$  intersect, the signal is supposed to be over. For details, see text and flow chart (Fig. 16.3).



**Fig. 16.3** Flow chart of Allen's picking algorithm. See text for details.



**Fig. 16. 4** Automatically derived P onsets (blue vertical lines) using Allen's algorithm applied to waveforms of local, regional and teleseismic events (black). Applied filtering and epicentral distances are given at top of each panel. Allen's CF (STA values of squared envelope function) is plotted in red. To the right zoomed in portions of waveforms and CFs in the vicinity of the P onset. The green vertical lines indicate the corresponding manual picks. The differences between manual P readings and automatically estimated onset times are less than 0.5 seconds for these local and regional waveform examples. For the teleseismic event waveform examples the differences are 1.7 and 4.2 seconds, respectively.

### 16.3.2 The Baer and Kradolfer picker

Another widely used picking algorithm is the one proposed by Baer and Kradolfer (1987). This algorithm is frequently applied, e.g. by PITSA ("Programmable Interactive Toolbox for Seismological Analysis", Scherbaum and Johnson, 1992) and the picking system MannekenPix (Aldersons, 2004).

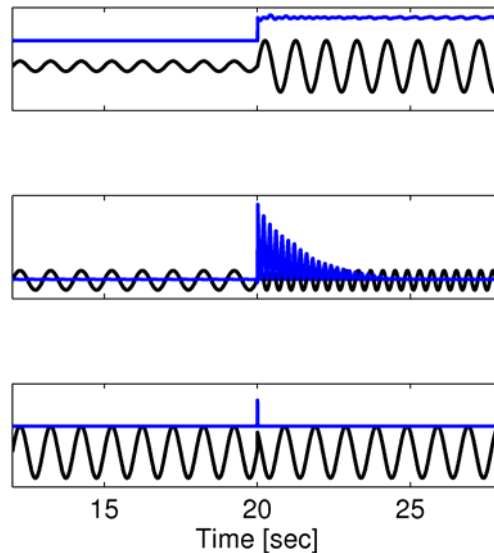
Baer and Kradolfer modified Allens' envelope function  $E(t)$  to

$$E_i^2 = x_i^2 + \frac{\sum_{j=1}^i x_j^2}{\sum_{j=1}^i 1}. \quad (16.4)$$

By squaring this envelope function and implementing the variance of  $E(t)$ , they obtain the following CF:

$$CF_i = \frac{E_i^4 - \overline{E_i^4}}{\sigma^2(E_i^4)} \quad (16.5)$$

where  $\overline{E_i^4}$  is the mean of  $E_i^4$  from  $j$ -th sample to  $i$ -th sample and  $\sigma^2(E_i^4)$  is the variance of  $E_i^4$  from sample  $j$ -th to sample  $i$ . As will be shown later, this CF is quite similar to the kurtosis, a parameter to quantify deviations from a Gaussian distribution. Tests on synthetic data using this CF are shown in Fig. 16.5. In contrast to Allen's squared envelope function this CF is sensitive to changes in amplitude, frequency as well as in phase.

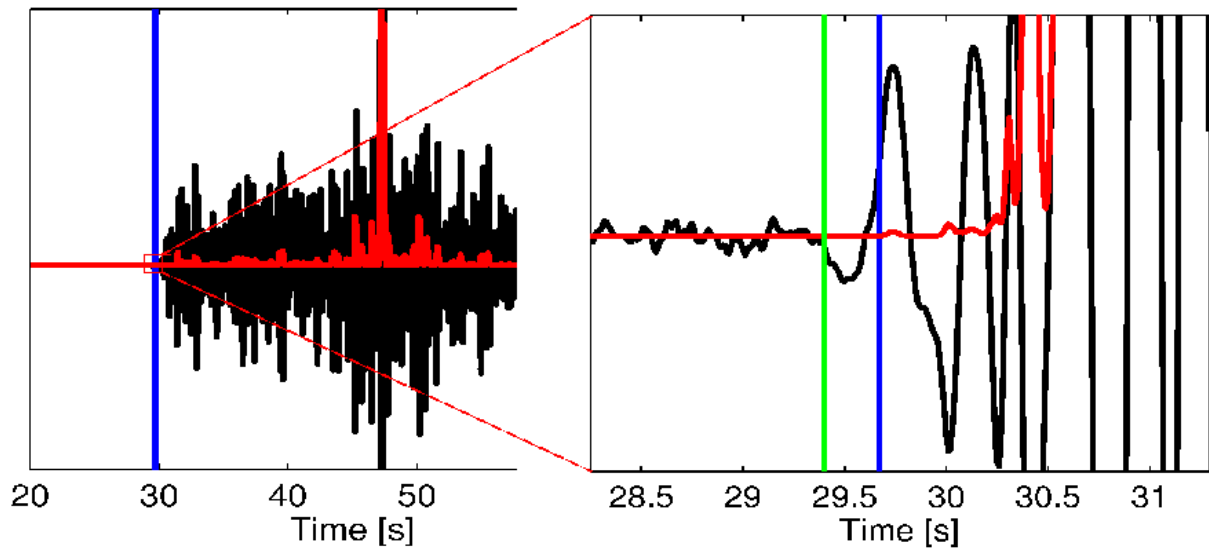


**Fig. 16.5** The Baer and Kradolfer CF (blue) for synthetic data (black) with change in amplitude (top), change in frequency (middle) and change in phase (bottom). The CF proposed by Baer and Kradolfer is sensitive to the three types of changes



A pick flag is set if  $CF_i$  exceeds a threshold  $\gamma \approx 10$ . In order to avoid detecting short-term increases caused by noise, a signal is only accepted if the CF does not drop below the signal threshold for times larger than the dominating period. The variance  $\sigma_i^2(E_i)$  is continuously updated, except when  $CF_i$  exceeds a second dynamic threshold  $\delta \approx 2 \cdot \gamma$ . If the CF decreases within a certain time “tup”, the provisional pick is cleared. However, due to the complexity of seismic signals, drops of the CF below the threshold  $\gamma$  for “tdown” seconds are allowed.

Fig. 16.6 shows an example of the proposed CF, calculated for the recording of a local event. In this specific case the derived onset is somewhat late compared to the manual P reading.

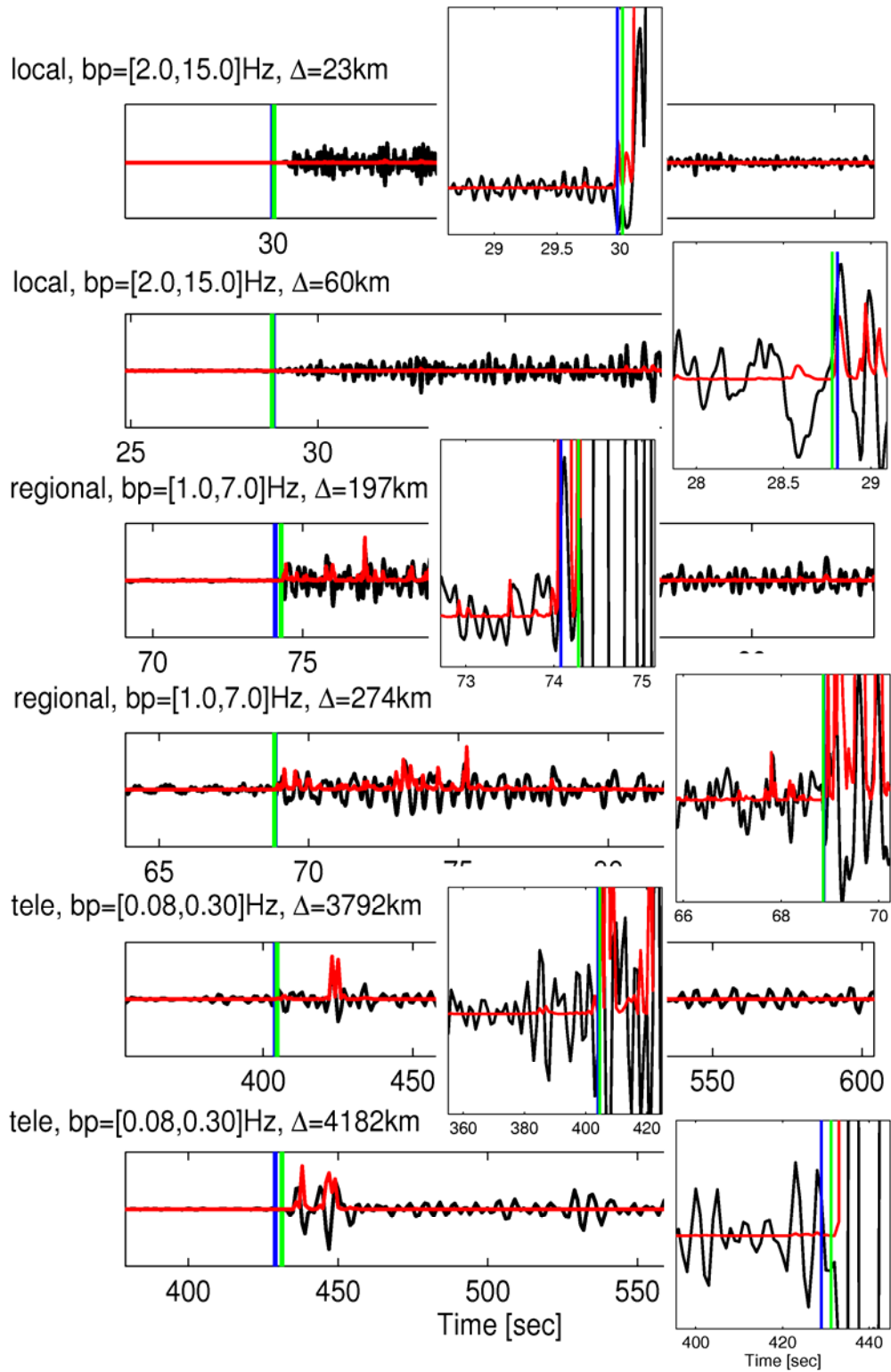


**Fig. 16. 6** Example of the Baer and Kradolfer CF (red) calculated for a local event waveform (black). The blue vertical line indicates the automatically derived P onset, the green vertical line the manual P reading.

Tab. 16.2 shows the parameters to be adjusted when applying the Baer and Kradolfer picker, with the right-hand side columns giving the respective values used for the example waveforms of a local, regional and teleseismic event in Fig. 16. 7.

**Tab. 16.2** Parameters to be adjusted when applying the Baer and Kradolfer picker with the respective case values of the parameters used for the example waveforms in Fig. 16.7.

| Parameter | Remark                                                                              | Values                     | Local | Regional | Tele |
|-----------|-------------------------------------------------------------------------------------|----------------------------|-------|----------|------|
| $\gamma$  | threshold                                                                           | 10                         | 10    | 5        | 2    |
| $\delta$  | threshold for updating $\sigma^2$                                                   | $2 \cdot \gamma$           | 20    | 10       | 10   |
| tup       | time [s] for CF to remain above threshold $\gamma$                                  | $>0.3$                     | 1.5   | 1.5      | 2    |
| tdown     | allowed time [s] for CF to drop below threshold $\gamma$ without clearing pick flag | mean of corner frequencies | 10    | 10       | 10   |



**Fig. 16.7** Examples of waveforms of local, regional and teleseismic events and of the Baer and Kradolfer CF. Applied filtering and epicentral distances are given at the top of each panel. The green vertical lines indicate the manual P picks, while the blue vertical lines indicate the automatically estimated P-onset times. The differences between manual and automatic picks are less than 0.2 seconds for these local and regional waveform examples. For the two teleseismic event waveform examples the differences are 0.7 and 2.8 seconds, respectively.

All automatically derived P onsets are in good agreement to the corresponding manual ones. However, it has been shown that this algorithm tends to be somewhat late compared to manual P readings (e.g. Sleeman and van Eck 1999; Küperkoch et al. 2010). Aldersons (2004) integrated the Baer and Kradolfer picker in the picking system MannekenPix and introduces a delay correction. The idea for the correction is to shift the automatic onset back in time as long as the CF decreases significantly towards earlier samples. The delay correction stops when  $CF_i - CF_{i-1}$  is smaller than 0.01, or when this condition cannot be met after moving back the onset by 3 samples. Aldersons states, that this simple correction usually provides “good to very acceptable results” for local earthquakes recorded by short-period instruments.

The Baer and Kradolfer picker is a fast and robust routine, which is also suitable for online detection. Baer and Kradolfer do not propose an automatic quality assessment. However, this algorithm yields high quality picks and was supplemented by the sophisticated quality assessment system MannekenPix (Aldersons, 2004; Di Stefano et al., 2006). Furthermore, the application of this algorithm is quite “user friendly” due to the low number of parameters to be set.

### 16.3.3 P-onset determination using Higher Order Statistics

When an earthquake signal arrives, the statistical properties of a seismogram change abruptly. Therefore, measurements of statistical properties in a moving window are suitable for the determination of a CF and subsequent estimation of arrival times. The statistical properties of the seismogram might be characterized by its distribution density function and by parameters like variance, skewness and kurtosis. The latter two are parameters of higher order statistics (HOS) and defined as follows (e.g. Hartung, 1991).

The expectation of a continuous distribution is given by

$$E[X] = \int_{-\infty}^{\infty} xp(x)dx \quad (16.6)$$

with the distribution function  $p(x)$  of the random variable  $X$ .

Using the expectation the statistic moment  $\alpha$  of order  $k$  of the random variable  $X$  is defined as:

$$\alpha_k = E[X^k] = \int_{-\infty}^{\infty} x^k p(x)dx. \quad (16.7)$$

By analogy the central statistic moment  $m$  of order  $k$  is defined as:

$$m_k = E[(X - E[X])^k], k > 1. \quad (16.8)$$

The second central moment is the variance, the lowest moment yielding informations about the variability of a random variable:

$$Var[X] := E[(X - E[X])^2] = m_2. \quad (16.9)$$

The variance defines the mean power of the alternating part of an ergodian process.

Using the third central moment the skewness becomes:

$$S = \frac{E[(X - E[X])^3]}{E[X - E[X]]^{3/2}} = \frac{m_3}{m_2^{3/2}} \quad (16.10)$$

S becomes zero if the distribution is symmetrical. It becomes negative (positive) if the distribution shows outlayers to the left (right) of the expectation value. The skewness provides information about positive or negative deviations of the distribution density function from the expectation value.

The kurtosis is defined using the fourth central moment:

$$K = \frac{E[(X - E[X])^4]}{E[X - E[X]]^{4/2}} = \frac{m_4}{m_2^2}. \quad (16.11)$$

K is 3 for normally distributed random variables.

An estimation of a central moment from a random sample  $x_j$ ,  $j = 1, K$ ,  $N$  is:

$$\hat{m}_k = \frac{1}{N} \sum_{j=1}^N (x_j - \bar{x})^k. \quad (16.12)$$

Estimations of the variance, skewness and kurtosis are hence given by:

$$\hat{\sigma}^2 = \hat{m}_2 = \frac{1}{N} \sum_{j=1}^N (x_j - \bar{x})^2 \quad (16.13)$$

$$\hat{S} = \frac{\hat{m}_3}{\hat{m}_2^{3/2}} \quad (16.14)$$

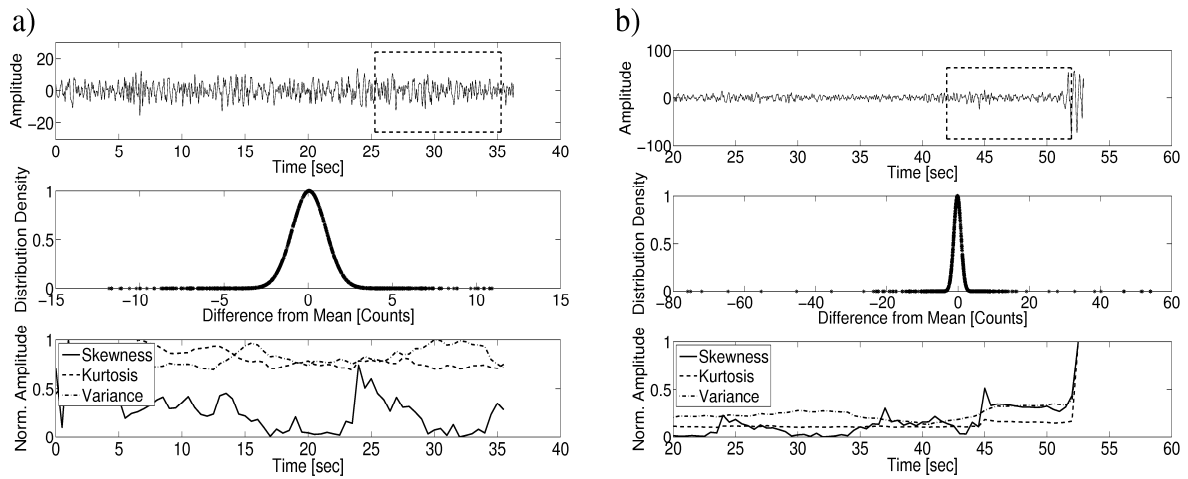
$$\hat{K} = \frac{\hat{m}_4}{\hat{m}_2^2} \quad (16.15)$$

Tab. 16.3 shows some example spot checks and their corresponding estimates of variance, skewness and kurtosis. Especially skewness and kurtosis show the potential to detect even small outliers. The first two examples show symmetrically distributed spot checks. The estimated skewness is zero. In the second example the variance increases, though no outlier distorts the distribution. The skewness remains zero and kurtosis increases only slightly. In the third example the outlier is clearly indicated by skewness and kurtosis, while the estimate of the variance remains the same as for the second example.

**Tab. 16.3** Examples of spot checks and corresponding values for variance, skewness and kurtosis.

| Spot Check              | $\hat{\sigma}^2$ | $\hat{S}$ | $\hat{K}$ |
|-------------------------|------------------|-----------|-----------|
| $[1, -1, 1, -1]$        | 1                | 0         | 1         |
| $[1, -1, 2, -2]$        | 2.5              | 0         | 1.36      |
| $[1, -1, 1, -\sqrt{7}]$ | 2.5              | -1.11     | 2.08      |

Fig. 16.8a shows the distribution density function of real background noise. Variance, skewness and kurtosis are calculated in a moving window with a length of 20 seconds. Skewness and kurtosis indicate an almost Gaussian distribution. As soon as the moving window reaches a signal onset (Fig. 16.8b), the shape of the distribution density function changes abruptly and deviates from a Gaussian distribution. Kurtosis and skewness increase strongly.



**Fig. 16.8** **a)** Background noise (top) and corresponding distribution density function (middle), calculated in a moving window of 20 seconds length (dashed box). Bottom: estimated variance, skewness and kurtosis (normalized).

**b)** The moving window reaches a signal onset. The distribution density function is no longer Gaussian shaped, variance, skewness and kurtosis increase strongly.

Longbottom et al. (1988) used higher order statistics (HOS) for the deconvolution of seismic data and called their algorithm a simplified minimum entropy deconvolution method. Saragiotis et al. (2002) suggest to estimate skewness and kurtosis in a moving window to pick P-wave arrival times. Gentili and Bragato (2006) and Gentili and Michelini (2006) used skewness, kurtosis and a combination of skewness and kurtosis and their time derivatives as input for a neural network trained to estimate P-arrival times. Groos and Ritter (2009) used higher order statistics to classify broadband urban noise (USN).

Using eqs. (16.14) and (16.15), skewness and kurtosis are estimated in a moving window. This yields the CF from which the arrival times of the P wave is determined. In order to make the calculation fast, a recursive procedure is suggested:

Let  $\{x(j)\}$  be a zero-mean, stationary process,  $T$  the length of the moving window and

$$N = T/dt + 1, \quad (16.16)$$

the number of samples of the moving window, with  $dt$  being the sampling interval. The actual value for the central moment of order  $k$  of the moving window is:

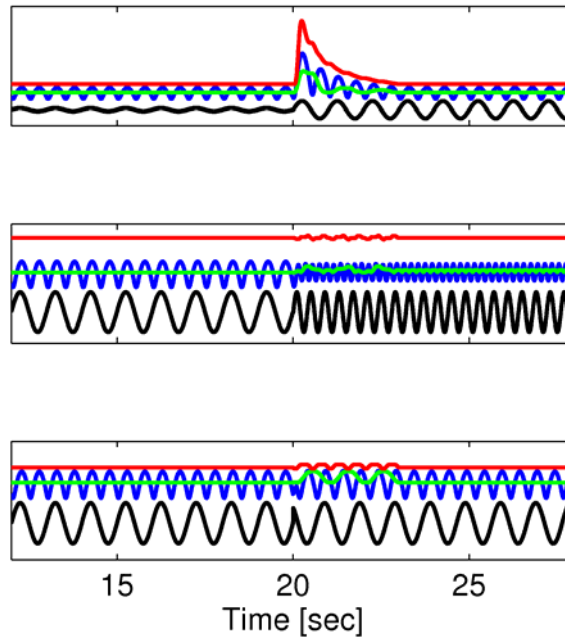
$$\hat{m}_k(j) = \frac{1}{N} \sum_{l=0}^{N-1} x_{j-l}^k. \quad (16.17)$$

Its estimate at sample  $j$  may be calculated from the previous estimate at sample  $j-1$ :

$$\hat{m}_k(j) = \hat{m}_k(j-1) - (x^k(j-N) + x^k(j)) / N. \quad (16.18)$$

Using eqs. (16.14) and (16.15) skewness and kurtosis are calculated from recursively estimated central moments. Computation times may be decreased by a factor of about 10.

In Fig. 16.9 CFs calculated using skewness and kurtosis are compared to the CF of an STA/LTA for synthetic data with a change in amplitude, frequency and phase, respectively. The test indicates the strong sensitivity of skewness and kurtosis for changes in amplitude. While changes in frequency are detected by skewness and kurtosis, there is only a marginal detectability of changes in phase.



**Fig. 16.9** Tests on synthetic data (black) with changes in amplitude (top), frequency (middle) and phase (bottom). Green: skewness, red: kurtosis, blue: STA/LTA.

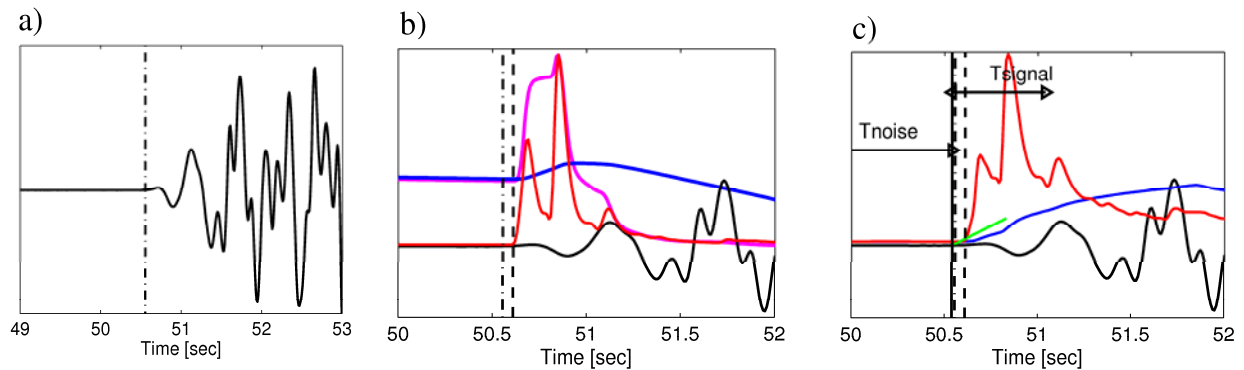
From the CF arrival times are determined. Küperkoch et al. (2010) propose a sophisticated, iterative algorithm, which is organized into four stages. A detailed description of the algorithm and the parameter settings is given in their paper. The algorithm might briefly be described as follows:

(1) A CF using skewness or kurtosis is calculated from a bandpass filtered waveform (2-10 Hz for local to regional events). In analogy to Maeda (1985) the Akaike Information Criterion (AIC, Akaike, 1971) is calculated, yielding a first approximate P onset, which is the minimum of the AIC function (Fig. 16.10b).

(2) Using this preliminary P onset, the CF is recalculated around the initial onset in a smaller window, considering a higher frequency content (e.g. 2-15 Hz). The picker searches then for a common local minimum of a smoothed and the unsmoothed CF as this indicates a P-wave onset. The search is carried out to the right and to the left of the initial P onset within a certain pick window. If local minima are found on both sides of the initial P onset, the lower common minimum with lower amplitude of the CF is assumed to coincide with the true P-arrival time (Fig. 16.10c).

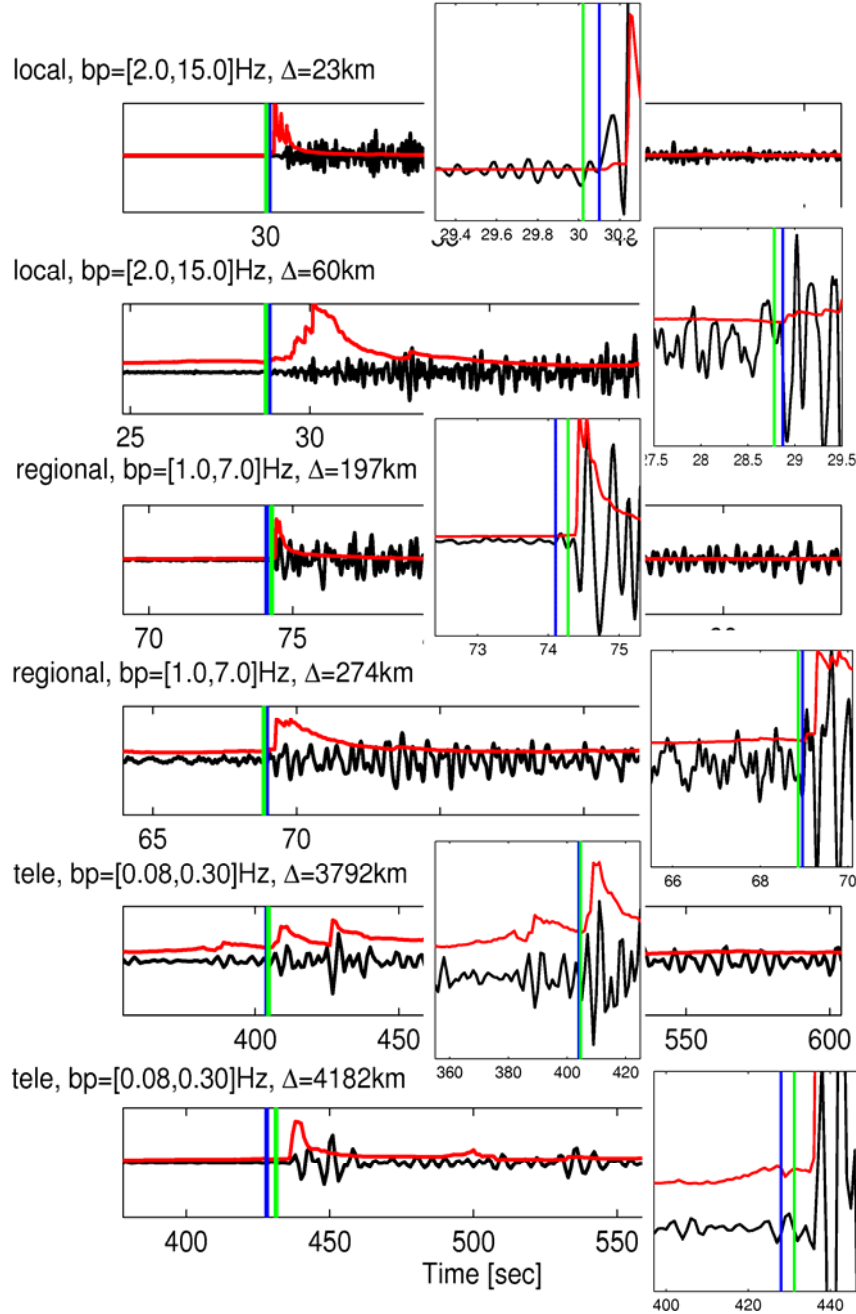
(3) For automatic quality assessment the slope of the CF right after the determined phase onset and the signal-to-noise ratio (SNR) serve as two quality criteria (Fig. 16.10c).

(4) Erroneous P onsets are found by checking the signal length and comparing the energy of the vertical component with the energy of the horizontal components to get rid of S picks, spuriously picked as P onsets. Furthermore, the consistency of automatically picked P onsets is checked. The difference between the picks should be lower than a certain threshold and the individual P picks should not have a strong influence on the estimate of the variance of the P picks. This is tested by a Jackknife procedure.



**Fig. 16.10** Automatic determination of a P onset in a 2-10 Hz bandpass filtered, local event waveform (a, black) using the iterative picking algorithm proposed by Küperkoch et al. (2010). The manual P reading is indicated by the dashdotted vertical line. (b) Zoomed in portion of the waveform (black), also showing the CF calculated using kurtosis (red), the unsmoothed AIC function (cyan) and the smoothed AIC function (blue). The initial P onset (dashed vertical line) is determined from the two AIC functions. (c) Zoomed in waveform (black), recalculated, unsmoothed CF (red) using kurtosis, calculated from 2-15 Hz bandpass filtered data, and recalculated, smoothed CF (blue). The green line indicates the slope fitted to the unsmoothed CF. The noise window  $T_{noise}$  and the signal window  $T_{signal}$  are used for SNR estimation.  $T_{noise}$  is 2 seconds.

Fig. 16.11 shows applications of the picking algorithm using kurtosis to local, regional and teleseismic events. For local and regional waveform examples the automatically derived P onsets show less than 0.2 seconds deviation from the manual picks. For the teleseismic waveform examples the differences are 0.7 and 3.2 seconds, respectively.



**Fig. 16.11** Examples for waveforms of local, regional and teleseismic events and the corresponding characteristic functions determined using kurtosis. Applied filtering and epicentral distances are given at the top of each panel. The green vertical lines indicate the manual P picks, the blue vertical lines the automatically derived P-onset times. While the differences between manual picks and automatically P-arrival time estimates are less than 0.2 seconds for the local and regional waveform examples, the differences for the teleseismic waveform examples are 0.7 and 3.2 seconds, respectively.



Skewness and kurtosis are very sensitive to changes in amplitude, while changes in frequency or phase are hardly detectable. The proposed picking algorithm yields very accurate P readings in combination with a reliable quality assessment. Due to the recursive calculation of higher order central moments, this approach is also suitable for onset detection. The length of the moving window for kurtosis and skewness calculation typically range between 2 and 20 seconds, depending on the dominating frequency from which the window should include not less than about approx. 1.5 oscillations.

Skewness and kurtosis are successfully applied by Saragiotis et al. (2002) to 44 seismic events. The corresponding P onsets are compared to manual derived P onsets as well to automatically derived P onsets using Allen's algorithm (see sub-section 16.3.1). The comparison yields better results for the HOS approach. Furthermore, a large scale comparison was performed by Küperkoch et al. (2010) using more than 3000 manually derived P onsets of local to regional events. They found for their data set excellent results when using kurtosis and outperformed the additionally applied picking algorithms proposed by Allen and Baer and Kradolfer (see sub-section 16.3.2).

### 16.3.4 The AR-AIC picker

A stochastic time series, in which the  $i$ -th sample is described as a linear combination of the  $p$  predecessors, is called an autoregressive process of order  $p$ , which can be abbreviated as AR( $p$ ). The mathematical representation is

$$x_m = \sum_{m=1}^p a_m x_{m-j} + \varepsilon_m, \quad (16.19)$$

with  $E[x]=0$ , where  $a_m$  are the coefficients or parameters of the AR process and  $\varepsilon_m$  white noise. Eq. (16.19) can be rewritten as

$$x_m = a_1 x_{m-1} + \dots + a_p x_{m-p} + \varepsilon_m. \quad (16.20)$$

When an earthquake signal arrives, the characteristics of a seismogram, such as variance and the spectrum, change abruptly. For the estimation of a phase arrival, it is assumed that each of the segments before and after the arrival of the seismic wave is stationary and might be expressed by an autoregressive model as follows (e.g. Kitagawa et al., 2001):

noise model (1<sup>st</sup>, pre-onset segment, for  $n=L, \dots, k$ ):

$$x_n^1 = \sum_{m=1}^M a_m^1 x_{n-m} + \varepsilon_n^1 \quad (16.21)$$

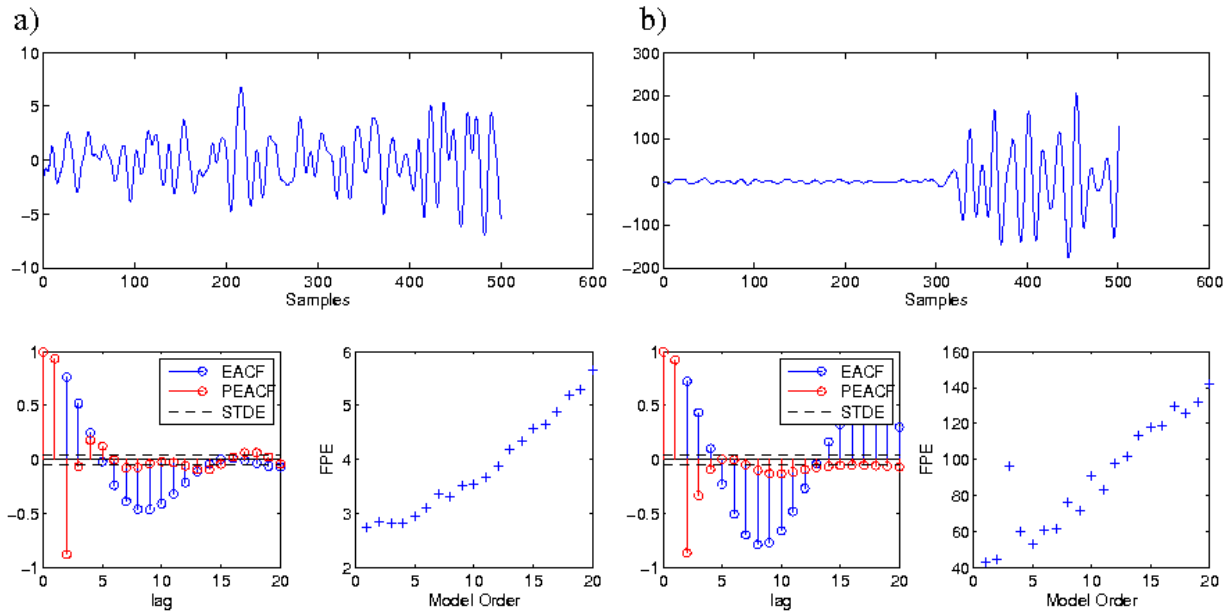
signal model (2<sup>nd</sup>, post-onset segment, for  $n=k+L, \dots, N$ ):

$$x_n^2 = \sum_{m=1}^L a_m^2 x_{n-m} + \varepsilon_n^2 \quad (16.22)$$

where  $k+1$  is the change point between the two segments (i.e. the phase arrival),  $a_m^1$  and  $a_m^2$  are the AR coefficients,  $\varepsilon_n^1$  and  $\varepsilon_n^2$  represent gaussian distributed noise in the two segments, and  $M$  and  $L$  are the orders of the AR processes of the noise and the signal model, respectively, which are all unknown parameters. In the following we briefly outline the estimation of these parameters and show an application of locally, stationary segments and its description with autoregressive processes.

Waveforms of seismic phases show usually a higher complexity than that of the preceding noise and should therefore be described by a larger number of AR parameters, i.e., by an AR model of higher order  $p$  (e.g. Leonard and Kennett, 1999). Leonard and Kennett (1999) obtain the order of the AR process by fitting AR power spectra to noise and signal power spectra. Another approach of estimating the order of an AR process (called model identification) is the use of the (empirical) partial autocorrelation function (PACF, e.g. Box et al., 1994). While the (empirical) autocorrelation function (ACF) gives the relation between the actual value of a time series and a later (earlier) value (delayed with time lag  $\tau$ ), not taking into account the influence of interjacent values, the PACF gives the direct correlation between lag- $k$  distant values by removing the influence of interjacent values. For an AR process of order  $p$  the PACF will be nonzero for time lags  $l$  less than or equal  $p$  and zero for time lags  $l > p$  and hence gives informations about the order of an AR process. Akaike (1970) proposed the final prediction error (FPE) to estimate the model order.

Fig. 16.12a) and 16.12b) show the pre-event noise window, the signal onset and the corresponding empirical ACF as well as the empirical PACF, from which the order of the process is estimated. The order of the time series incorporating the signal onset is much higher than that for the noise window. The PACF is assumed to be zero if the PACF is smaller than the standard S.E., which is  $S.E. = 1/\sqrt{N}$  (e.g. Box et al., 1994), where  $N$  is the number of observations. PACF and FPE yield the same model orders.



**Fig. 16.12** a) Pre-event portion of a seismogram, corresponding ACF and PACF for sample lags 0 to 20, and FPE as a function of model order. The order of the AR process can be obtained from the PACF. The last value of the PACF which is larger than the standard error (dashed line) occurs at sample lag 3. The order of the AR process is therefore 3. The FPE

function also gives a local minimum at model order 3. **b)** Pre-event portion, seismic phase onset, corresponding ACF and PACF for sample lags 0 to 20, and FPE as a function of the model order. The order of the AR process is estimated to be 5, as the last value of the PACF, which is larger than the standard error, is at sample lag 5, indicating a higher complexity of the seismic signal compared to the preceding noise. Also the FPE gives a local minimum at model order 5.

However, the estimation of the order of an AR process is very uncertain, and hence in most AR applications a fixed AR order is used, estimated with the introduced procedure or by trial and error (e.g. Leonard and Kennett, 1999).

The so called autoregressive-Akaike-Information-Criterion-picker (AR-AIC) proposed by Sleeman and van Eck (1999) is based on the work by Akaike (1971, 1974), Morita and Hamaguchi (1984) and Takanami and Kitagawa (1988). A longer time series is divided into two locally stationary segments each modeled by an AR process. The first segment represents noise, the second segment contains the signal. The picking algorithm can be described by five steps: (1) Bandpass filtering of the seismogram, (2) detection of a seismic phase using an STA/LTA detector, (3) estimation of two sets of AR parameters, namely for noise and signal, respectively, (4) calculation of two prediction errors using the AR parameters for noise and signal, respectively, (5) the minimum of the two-model AIC indicates the arrival time. For the following detailed description of the picker algorithm we assume that due to a first rough estimation of the P-wave arrival time the time series  $x_n = x_1, \dots, x_N$  can be divided into two subseries: the first represents noise, the second the seismic phase (signal). Thus, starting point for the following derivation are eqs. (16.21) and (16.22). Furthermore, we assume a fixed order  $M$  for both models, estimated by trial and error or the use of the PACF.

As we assume Gaussian distributions of  $\varepsilon_n^1$  and  $\varepsilon_n^2$ , we can write for the two corresponding likelihood functions:

$$l(\varepsilon_n^1) = \frac{1}{\sigma_1 \sqrt{2\pi}} e^{-\frac{(\varepsilon_n^1)^2}{2\sigma_1^2}} \quad (16.23)$$

$$l(\varepsilon_n^2) = \frac{1}{\sigma_2 \sqrt{2\pi}} e^{-\frac{(\varepsilon_n^2)^2}{2\sigma_2^2}} \quad (16.24)$$

The approximate likelihood of the locally stationary AR model in the intervals  $[M+1, k]$  and  $[k+1, N-M]$  is hence given by

$$L(x, M, a_m^i, \sigma_i^2 (i=1,2)) = \prod_{i=1}^2 \left( \frac{1}{\sigma_i^2 2\pi} \right)^{\frac{N_i}{2}} \exp \left( -\frac{1}{2\sigma_i^2} \sum_{n=p_i}^{q_i} \left( x_n - \sum_{m=1}^M a_m^i x_{n-m} \right)^2 \right) \quad (16.25)$$

with  $p_1 = M+1$ ,  $p_2 = k+1$ ,  $q_1 = k$ ,  $q_2 = N-M$ ,  $n_1 = k-m$  and  $n_2 = N-M-k$ . The condition for maximum likelihood estimates of the model parameters is

$$\frac{\partial \log(L(x, M, a_m^i, \sigma_i^2))}{\partial (a_m^i, \sigma_i^2)} = 0, \quad (16.26)$$

which yields

$$\sigma_{i,max}^2 = \frac{1}{n_i} \sum_{j=p_i}^{q_i} \left( x_j - \sum_{m=1}^M a_m^i x_{j-m} \right)^2. \quad (16.27)$$

Note, that the second sum is equivalent to the prediction error.

By substituting eq. (16.27) into eq. (16.25), the maximum of the logarithmic likelihood function  $\log(L(k, x, M, a_m^i, \sigma_i^2))$  for the two models as function of division point  $k$  becomes:

$$\log(L(k, x, M, a_m^i, \sigma_i^2)) = -\frac{1}{2}(k-M)\log(\sigma_{1,max}^2) - \frac{1}{2}(N-M-k)\log(\sigma_{2,max}^2) + C \quad (16.28)$$

where  $C$  is a constant. In the case of the locally stationary AR model, the AIC, which is a criterion for the selection of the best statistical model, is given by (Akaike, 1971):

$$AIC = -2(\text{maximum log likelihood}) + 2(\text{number of parameters}),$$

where the number of parameters is given by the model order. For the merging point  $k$ , which separates the two models, the AIC becomes:

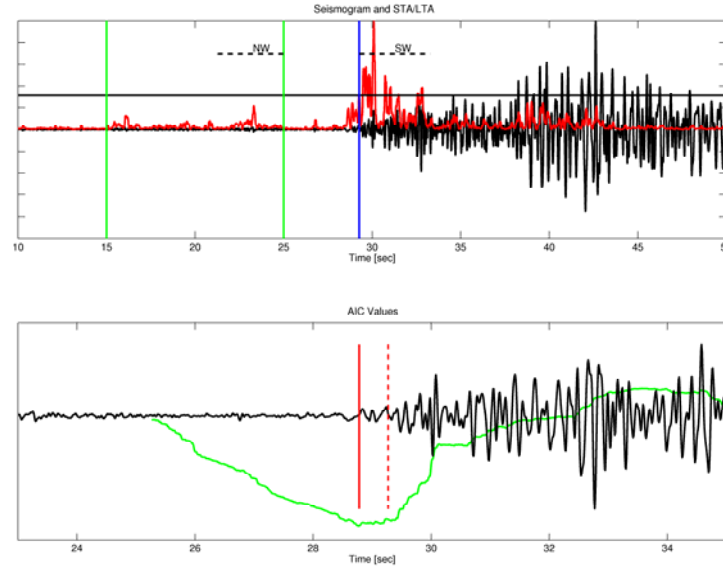
$$AIC(k) = AIC^1 + AIC^2 \quad (16.29)$$

and thus:

$$AIC(k) = (k-M)\log(\sigma_{1,max}^2) + (N-M-k)\log(\sigma_{2,max}^2) + C_1 \quad (16.30)$$

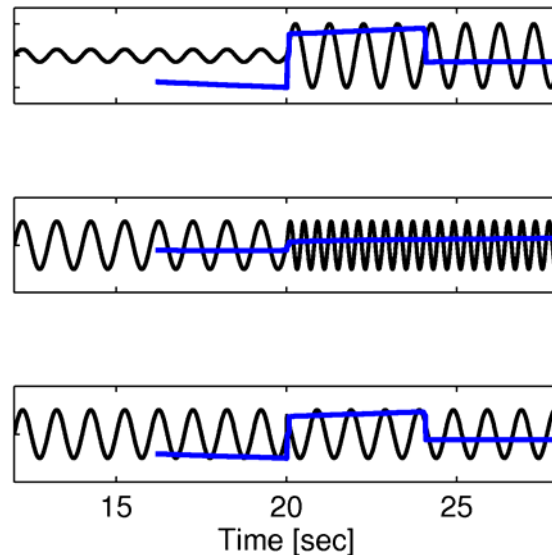
where  $C_1$  is a constant. The minimum of the AIC points then to the optimal separation time of the two stationary time series and indicates the arrival time of the phase. The AR coefficients  $a_m^i$  in eq. (16.27) are estimated in advance separately for the noise and the signal model, respectively. This is usually accomplished by either the Yule-Walker approach (e.g. Box, 1994), Burgs algorithm (Burg, 1975), or the least-squares approach.

In the following the AR-AIC picker is applied to synthetic and real test waveforms. The AR coefficients of the noise and the signal models are estimated in windows of 4 seconds length. The window of the noise model starts 4 seconds before the initial estimate of the arrival time. The window for the estimation of the AR parameters of the signal starts at the initial estimate of the arrival time (Fig. 16.13, top). The picker searches for the minimum in eq. (16.30) in a time window of 12 seconds length starting 8 seconds before the initial STA/LTA detection (Fig. 16.13, bottom). Sleeman and van Eck (1999) fixed the order of the two AR models to 8. Leonard (2000) used an order 4 AR model for both signal and noise.



**Fig. 16.13 Top:** Waveform of a local event (black) and an initial P-onset detection (blue) derived from STA/LTA ratios (red). The green vertical lines indicate the time interval for noise level estimation. The dashed lines denote the time windows for determination of AR coefficients for the noise and the signal model, respectively (NW and SW). **Bottom:** Waveform (black) around the initial STA/LTA pick (red,dashed line) and the automatically determined P onset (red line). Green: The AIC as a function of the merging point  $k$ , determined using eq. (16.30).

Fig. 16.14 shows the results for the synthetic data. For this test, we reduced the assumed order of the AR process to 2. The test shows clearly the potential of AR algorithms to detect changes in amplitude, frequency as well as in the phase.



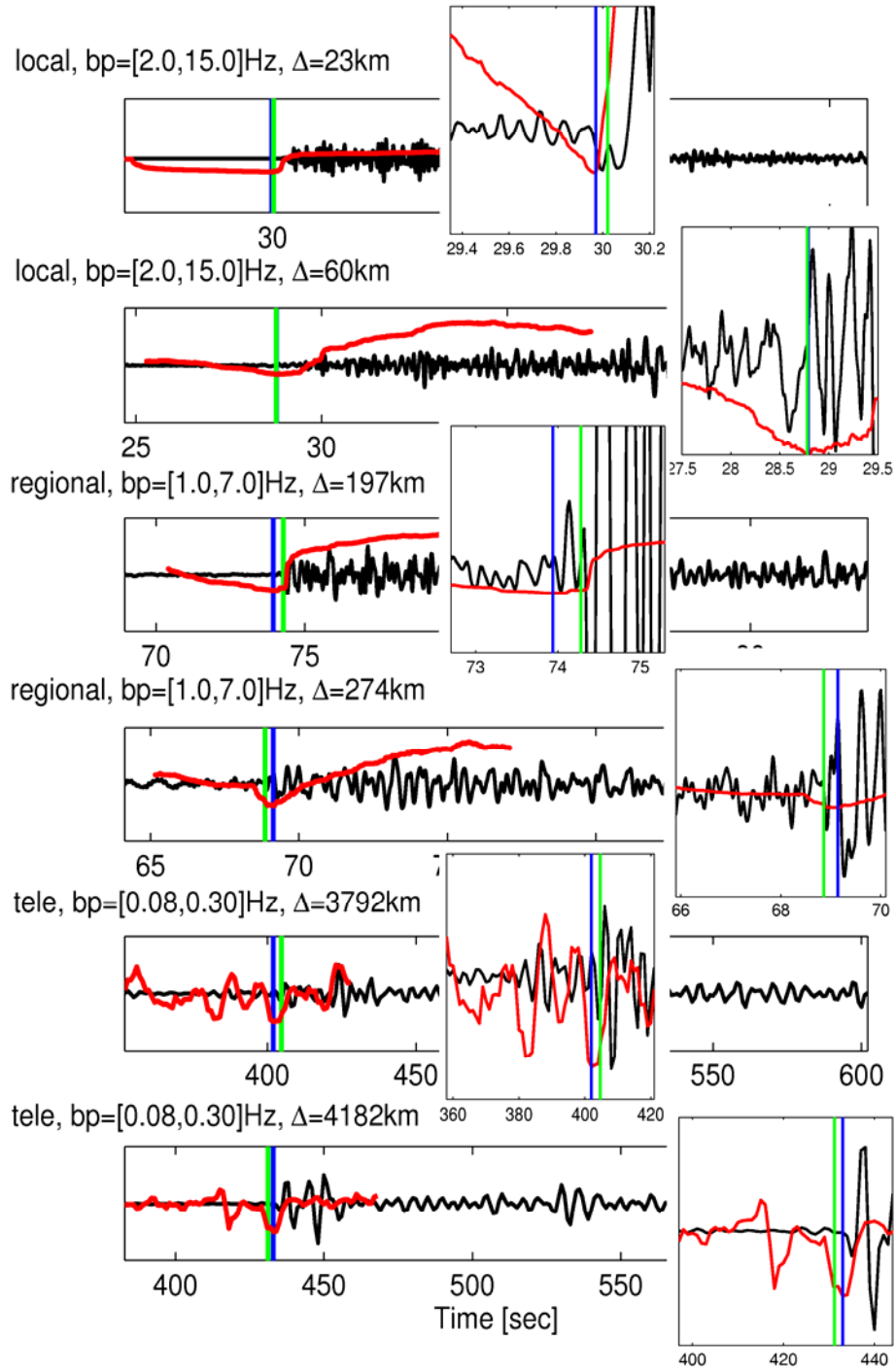
**Fig. 16.14** AR-AIC (red) as function of the merging point  $k$  for synthetic data (black) with changes in amplitude (top), changes in frequency (middle), and changes in phase (bottom). The assumed order of the AR process is reduced to 2. Changes in the time series are recognized in all three cases.

**Tab. 16.4** Parameters to be adjusted for the AR-AIC picker. Parameter settings applied for STA/LTA trigger used to get initial the P onset are not shown. The outer right columns represent the values used for the example waveforms in Fig. 16.15.

| Parameter      | Remark                                                       | Values | Local | Regional | Tele |
|----------------|--------------------------------------------------------------|--------|-------|----------|------|
| <i>tseg1</i>   | length of 1 <sup>st</sup> segment for calculating AIC(k) [s] | 4      | 8     | 8        | 50   |
| <i>tseg2</i>   | length of 2 <sup>nd</sup> segment for calculating AIC(k) [s] | 4      | 4     | 4        | 40   |
| <i>tnoise</i>  | length of noise window for AR-coefficient determination [s]  | 4      | 4     | 4        | 50   |
| <i>tsignal</i> | length of signal window for AR-coefficient determination [s] | 4      | 4     | 4        | 50   |
| <i>offset1</i> | offset between initial pick and <i>tnoise</i> [s]            | 4      | 4     | 4        | 50   |
| <i>offset2</i> | offset between initial pick and <i>tsignal</i> [s]           | 0      | 0     | 0        | 0    |

Fig. 16.15 shows applications of the AR-AIC picker to real local, regional, and teleseismic waveform data. Tab. 16.4 summarizes the parameters used for the example waveforms in Fig. 16.15 and the parameters proposed by Sleeman and van Eck (1999).

One shortcoming of this picker is the dependency on the STA/LTA trigger, which may be exchanged by a more robust procedure. The weighting scheme based on signal-to-noise ratios is useful for detecting false picks, but does not yield reliable quality estimates of the onset. However, the shape of the AIC function depends on the “sharpness” of the onset and could thus serve as a quality criterion (Diehl et al., 2009b, electronic supplement). The AR-AIC picker is more expensive but also more powerful than the Allen picker and the Baer and Kradolfer picker due to its strong sensitivity to changes in the waveform. Furthermore, this picking algorithm represents a robust tool for the identification of a first arrival.



**Fig. 16.15** Waveform examples of local, regional and teleseismic events and the corresponding characteristic functions for the AR-AIC picker proposed by Sleeman and van Eck (1999). Applied filtering and epicentral distances are given at top of each panel. The initial pick is determined by a STA/LTA detector. AIC as a function of the merging point is shown in red. The blue vertical lines denote the automatically determined P-phase arrival times, the green vertical lines the manual P picks. For local event waveform examples, the differences between manual picks and automatically estimated P-arrival times are 0.05 and 0.01 seconds, for regional waveform examples 0.34 and 0.28 seconds, and for teleseismic waveform examples 2.7 and 1.8 seconds, respectively.

### 16.3.5 Discussion of presented P-picking algorithms

The previous section presented four widely used P-picking algorithms in great detail. However, for potential users it might be difficult to decide which algorithm to choose for his certain data set. Therefore, we try to summarize the pros and cons of the four presented algorithms.

The Allen picker is a fast and robust algorithm, which also accounts for automatic quality assessment. However, as this algorithm is amplitude based only, it might miss emergent P onsets. A comparative study by Küperkoch et al. (2010) showed that this algorithm tends to pick somewhat early compared to what an analyst would pick.

The Baer and Kradolfer picker is also very fast and robust and quite user-friendly, as this algorithm only needs 4 input parameters. A shortcoming of this algorithm is the missing automated quality assessment. Several comparative studies (Sleeman and van Eck 1999, Aldersons 2004, Küperkoch et al. 2010) showed that this picking algorithm tends to be somewhat late compared to manual P picks.

Though only amplitude based too, higher order statistics are quite sensitive even to emergent P onsets. In combination with a sophisticated picking algorithm (e.g. Küperkoch et al. 2010), which exploits the entire information provided by the determined CF, yields excellent results. If precisely tuned, the automated quality assessment proposed by Küperkoch et al. (2010) gives similar weights as the analysts. However, the choice of the various parameters needed for this sophisticated algorithm is quite difficult and needs some experience.

The AR-AIC picker is a highly sophisticated picking algorithm based on information theory. The algorithm is computationally quite expensive and hence much slower than the other presented picker. In the discussed version by Sleeman and van Eck (1999) the initial P onset is derived from an STA/LTA detector, which might miss emergent P arrivals or P-phase arrivals dominated by instantaneous changes in frequency. This may limit the performance of this picking algorithm, and it might be worthwhile to replace the STA/LTA detector with a more robust, but nevertheless fast picker like the Baer and Kradolfer picker or a picking algorithm based on higher order statistics. Furthermore, the implemented quality assessment in the Sleeman and van Eck version based on signal-to-noise ratio only is not sufficient for robust quality estimation of the derived P onsets. A more robust quality criterion might be the “sharpness” of the AIC function, as proposed by Diehl et al. (2009b). However, a multivariate improvement of the AR-AIC picker should also be able to pick precisely S-arrival times.

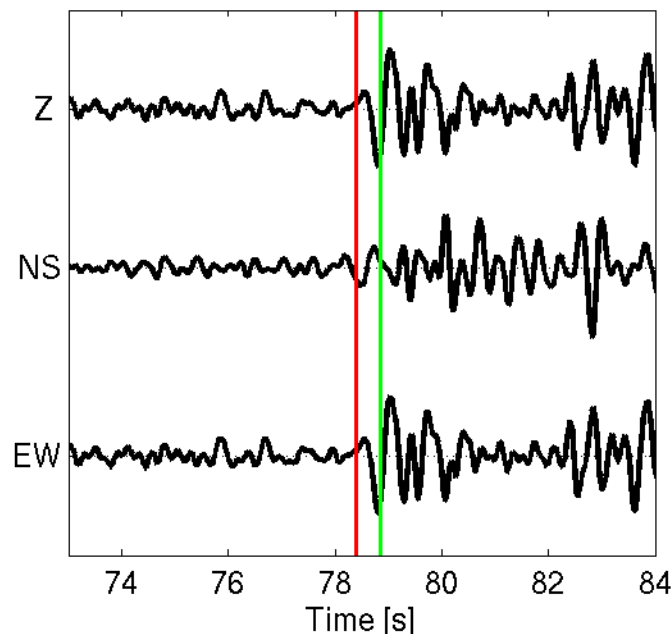
## 16.4 Automated S-onset determination

For robust location of earthquakes and especially for the determination of hypocentral depths, the estimation of S-onset times is crucial (Gomberg et al., 1990). Furthermore, S-wave arrival times may be inverted for models of the S-wave velocity supplementing P-wave velocity models and yielding additional information e.g. for petrological interpretation and seismic hazard models. However, S phase picking is more difficult than P-phase picking due to the character of the later arriving shear wave. Although by far most of the seismic wave energy is contained in the S waves and accordingly their amplitudes are generally larger than those of the related P waves (see record examples of local, regional and teleseismic events in DS 11.1-11.3), the often weaker very initial S-wave onset is usually buried in the preceding P coda.



Statistical properties of the P coda are highly variable and usually not Gaussian distributed. Therefore, the S onset may not be detectable by algorithms based on higher order statistics. S waves show often emergent onsets and may be preceded by S to P conversions. Moreover, while the energy of longitudinally polarized P waves is usually concentrated on the vertical component, the energy of the later arriving transversally polarized S waves is in general distributed over all three components. In addition, the occurrence of S-wave splitting may complicate the determination of the S-wave arrival time (Fig. 16.16). Thus, even manual picking and phase identification is often uncertain and inconsistent and experience is needed to pick and identify reliably the S onset. Hence, automated S-phase arrival time estimation is a challenging task and the development of optimized automated algorithms for picking of arrival times of S phases or other later arriving phases is a topic of ongoing research and applications to large data sets are still rare.

Reliable automatic S-onset determination makes use of all components or at least of both horizontal components. Usually, automatic picking of later phases is an iterative process. First the P onset has to be determined using the vertical component, from which a time window for S-phase picking is derived. Alternatively, the S phase is picked in a time window predicted from a preliminary event location. Algorithms proposed for P phase picking may also be used for S-phase arrival time determination when applied to transversal components. For instance, the AR-AIC picker (Takanami and Kitagawa (1988, 1991), Sleeman and van Eck (1999), see sub-section 16.3.4), has the potential of picking later phases if applied to horizontal components. Leonard and Kennett (1999) investigated single- and multi-component autoregressive modeling techniques for P- as well as for S-onset time determination. For multi-component analysis, the AR coefficients are represented by a second-order tensor. The order of the autoregressive processes representing the noise and the signal parts of the seismogram, respectively, are obtained by fitting AR-power spectra to observed ones.



**Fig. 16.16** Shear-wave splitting observed on a 2-10 Hz bandpass filtered local event recording. The green line indicates the SH-onset time, visible only on the north-south component. The red line indicates the SV-onset time, visible on the vertical and the east-west component.

Here, the algorithms proposed by Cichowicz (1993), Wang and Teng (1997) and Diehl et al. (2009b) are discussed in detail, as these algorithms are specifically developed for S-phase arrival time determination.

Cichowicz (1993) proposed an algorithm for automatic S-phase picking from three-component seismic data based on a polarization analysis, which is a powerful tool to distinguish between longitudinal and transversal energy, as discussed above. The algorithm might be described as follows:

Taking into account the dominating frequency, usually determined from displacement or velocity power spectrum, respectively, the length  $N$  of a moving window is determined. At first, the covariance matrix of the three orthogonal components of ground motion is computed within a window around the P onset, which is assumed to be known:

$$\begin{bmatrix} COV(X,X) & COV(X,Y) & COV(X,Z) \\ COV(Y,X) & COV(Y,Y) & COV(Y,Z) \\ COV(Z,X) & COV(Z,Y) & COV(Z,Z) \end{bmatrix} \quad (16.31)$$

where the estimate of the covariance for  $N$  observations of two variables  $X$  and  $Y$  is given by

$$COV(X,Y) = \frac{1}{N} \sum_i^N (x_i - \bar{x})(y_i - \bar{y}) \quad (16.32)$$

with  $\bar{x}$  and  $\bar{y}$  being average values. The direction of polarization is calculated by considering the eigenvector associated with the largest eigenvalue. Then, the X, Y and Z components are rotated into L, Q and T components, where L coincides with the principle direction of the P wave particle motion, i.e.

$$\begin{bmatrix} L \\ Q \\ T \end{bmatrix} = \begin{bmatrix} u_{11} & u_{12} & u_{13} \\ u_{21} & u_{22} & u_{23} \\ u_{31} & u_{32} & u_{33} \end{bmatrix} \cdot \begin{bmatrix} X \\ Y \\ Z \end{bmatrix} \quad (16.33)$$

where  $u_{i,j}$ ,  $j=1,2,3$ , are the direction cosines of the  $i$ -th principle direction. The covariance matrix and the following parameters are calculated for a moving window:

1) The deflection angle  $F_1(t)$  - the angle between the current polarization and the P-wave polarization - given by

$$F_1(t) = \frac{\cos^{-1} |u_{11}|}{\pi/2} \quad (16.34)$$

with  $u_{11}$  being the direction cosines in the L, Q, T coordinate system.

2) The degree of linear polarization  $F_2(t)$

$$F_2(t) = \frac{(\nu_1 - \nu_2)^2 + (\nu_1 - \nu_3)^2 + (\nu_2 - \nu_3)^2}{2 \cdot (\nu_1 + \nu_2 + \nu_3)^2}, \quad (16.35)$$

with  $\nu_1, \nu_2, \nu_3$  being the eigenvalues of the covariance matrix at time  $t$ .

3) The ratio  $F_3$  between transversal and total energy

$$F_3(t) = \frac{\sum_i^N (Q_i^2 + T_i^2)}{\sum_i^N (Q_i^2 + T_i^2 + L_i^2)} \quad (16.36)$$

These parameters are normalized and their theoretical values will be close to 1 for S waves. The CF is determined using these parameters:

$$CF(t) = F_1^2 \cdot F_2^2 \cdot F_3^2. \quad (16.37)$$

An S phase is declared, if the CF exceeds a threshold  $A$  for a few consecutive samples.  $A$  is calculated from the CF:

$$A = \overline{CF} + 3\sigma, \quad (16.38)$$

where  $\overline{CF}$  is the average value of the CF and  $\sigma$  the variance of the CF in the time window. The uncertainty of the automatic S pick is not evaluated by this algorithm. Sleeman and van Eck (2003) combined the polarization analysis with wavelet tranform (Rioul and Vetterli, 1991) and applied their approach to 313 local events. For their data, they found 44.1% to 47.9% of automatic S picks in the interval  $[-0.5, +0.5]$ s around the manual S onset and 65.5% to 68.9% of automatic S picks in the interval  $[-1.5, +1.5]$ s around the manual S pick.

Wang and Teng (1997) combine several approaches into an artificial neural network (ANN) algorithm. They consider the following properties of the three-component seismogram: (1) short-term to long-term average ratios (STA/LTA); (2) the ratios of transversal to total power (same as  $F_3$  in Cichowicz, 1993); (3) the change of autoregressive model coefficients. As they assume a second-order AR process, this change becomes:

$$ARC(i) = \frac{|a_1(i)a_2(i) - a_1(i-1)a_2(i-1)|}{|a_1(i)a_2(i)|}, \quad (16.39)$$

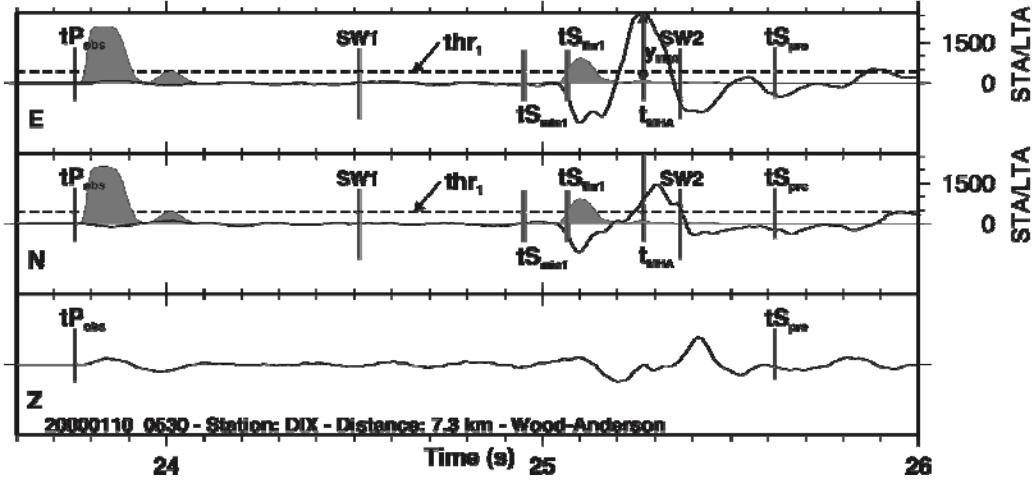
where  $a_1, a_2$  are the estimated AR coefficients.

(4) The fourth criterion is the deflection angle  $D(t)$  or the short-axis incidence angle of the polarization ellipsoid, i.e.

$$D(t) = \frac{\cos^{-1} |u_{31}|}{\pi/2}. \quad (16.40)$$

In their proposed ANN, each attribute is calculated by a neural subnet. The output of the subnets varies between 0 (no S phase) to 1. The output of the four subnets serve as input for the final decision unit, where the inputs are summed up. If the summation of inputs is larger than 3, the estimated arrival time is accepted.

Similar to their approach, Diehl et al. (2009b) propose a picking algorithm that combines STA/LTA ratios, polarization analysis, and AR-AIC picking (see sub-section 16.3.4). Combined STA/LTA ratios are calculated from both horizontal components as illustrated in Fig. 16.17. The picking algorithm is similar to the one proposed by Baer and Krüger (1987, see sub-section 16.3.2), but yielding an earliest ( $tS_{min1}$ ) and latest ( $tS_{thr1}$ ) possible pick, respectively.



**Fig. 16.17** Combined STA/LTA approach used for S-wave detection on horizontal components. Black solid curves represent the Wood-Anderson filtered three-component seismograms (amplitudes normalized by station maximum) of a local earthquake in Switzerland (Ml 3.1, focal depth of 9 km). The gray shaded trace denotes the combined STA/LTA ratio derived from N and E components.  $tP_{obs}$  represents the known P-arrival time, and  $tS_{pre}$  indicates the position of theoretical S arrival predicted from a regional one-dimensional model. The dashed horizontal line denotes the dynamic threshold  $thr_1$  for the picking algorithm. The S-wave arrival time based on the STA/LTA detector in the potential S window ( $SW1$  to  $SW2$ ) is most likely located in the interval between  $tS_{min1}$  (minimum pick) and  $tS_{thr1}$  (threshold pick). Copy of Fig. 2 in Diehl et al. (2009b, p. 1908) with © granted by the Seismological Society of America.

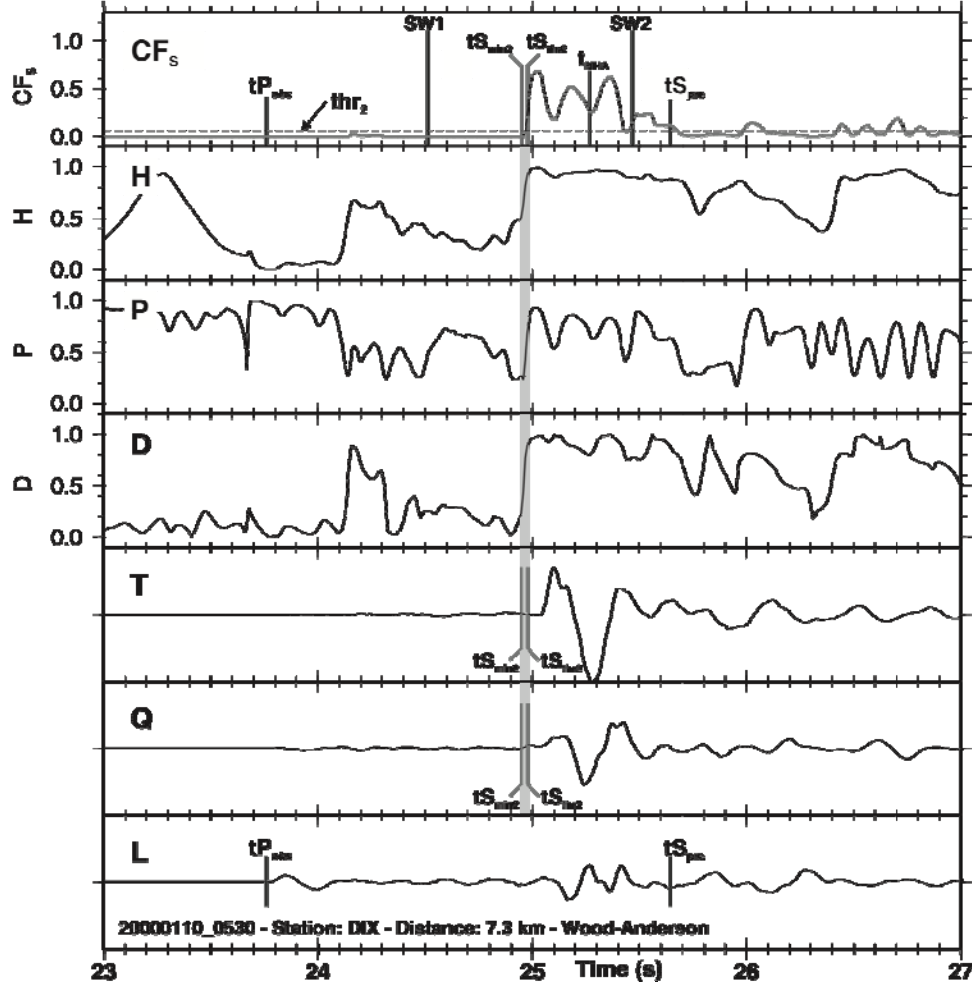
The polarization detector mainly follows the one proposed by Cichowicz (1993). In addition, they calculate a weighting factor  $W(t)$  for each window, which accounts for the absolute amplitude within the centered window with respect to the maximum amplitude derived from the coarse S window, derived from the observed P onset and the theoretical S onset. The characteristic function CF of the polarization detector is:

$$CF(t) = F_1(t)^2 R(t)^2 F_3(t)^2 W(t), \quad (16.41)$$

with  $F_1$  and  $F_3$  being the deflection angle and the ratio between transversal and total energy, respectively, as introduced above.  $R(t)$  is the rectilinearity:

$$R(t) = 1 - \frac{\nu_2 + \nu_3}{2 \cdot \nu_1}, \quad (16.42)$$

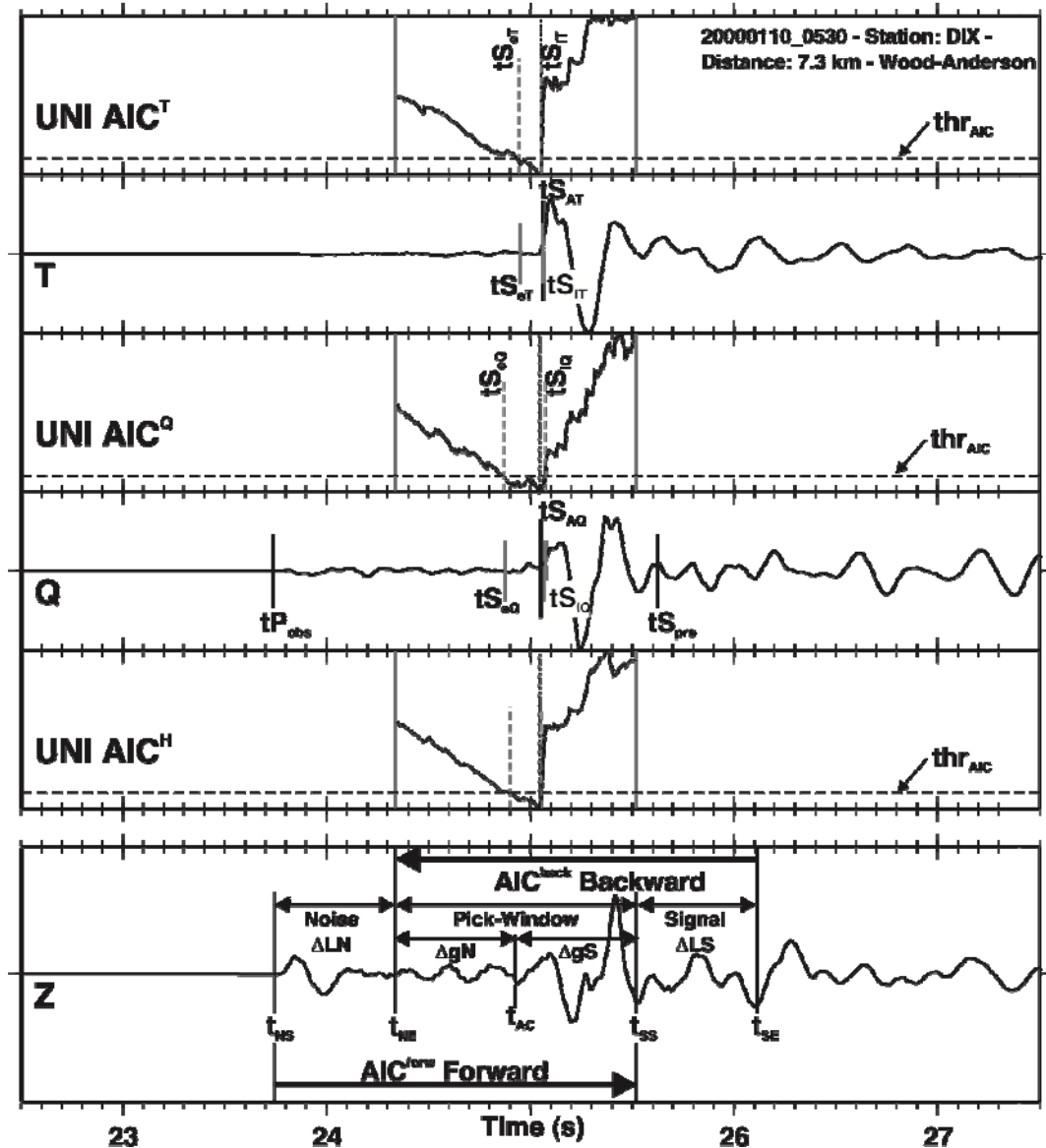
with  $\nu_1, \nu_2, \nu_3$  being again the eigenvalues of the covariance matrix. Fig. 16.18 illustrates the principle of the proposed polarization detector. A picking algorithm is applied to the derived CF, which yields an earliest ( $tS_{min2}$ ) and latest ( $tS_{thr2}$ ) possible estimate of the S phase arrival (see Fig. 16.18).



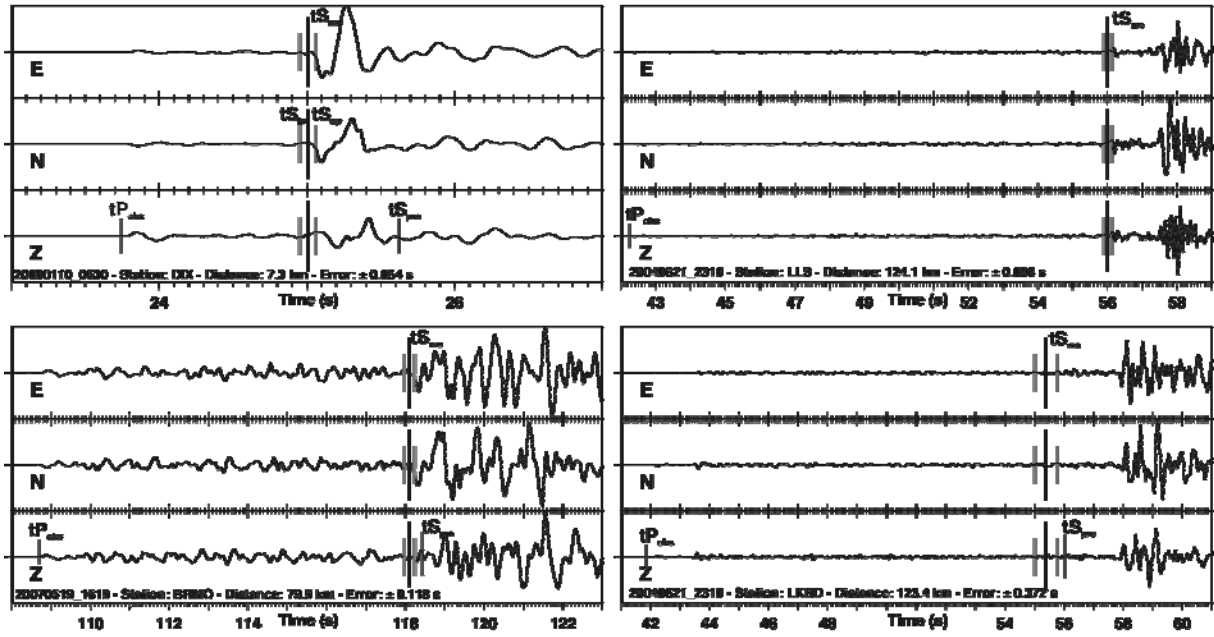
**Fig. 16.18** Example for the polarization detector. L, Q, and T denote the rotated components. The corresponding S-wave operators are  $D(t)$  (directivity),  $P(t)$  (rectilinearity), and  $H(t)$  (transverse to total energy ratio). The uppermost trace represents the amplitude weighted characteristic S-wave function  $CFS$ . The arrival of the S wave (gray band) goes along with the simultaneous increase of  $D(t)$ ,  $P(t)$ ,  $H(t)$ , and  $CFS$ . Compared to the actual arrival on T, the S wave detection is shifted by approximately 0.1 sec to earlier times. This time shift is caused by the finite length of the polarization filter.  $CFS$  is not affected by the P wave. Copy of Fig. 3 in Diehl et al. (2009b, p. 1910) with © granted by the Seismological Society of America.

The information provided by the detectors is used to set up the search window for a AR-AIC picker, applied to single original and rotated components and to both horizontal components as illustrated in Fig. 16.19. The implementation is mainly based on the method of Takanami and Kitagawa (1988). Finally, the earliest and latest possible picks from the STA/LTA detector, polarization detector, and the different AIC minima are used to estimate the ultimate automatic S-wave arrival time and its corresponding uncertainty. Examples of final automatic S-phase picks of different uncertainty classes are shown in Fig. 16.20.

[info/special\\_datasets.htm](http://www.tensorflow.org/info/special_datasets.htm).



**Fig. 16.19** Example of the AR-AIC picker. All amplitudes are trace normalized. The lower box illustrates the search window configuration centered around  $t_{AC}$ . The corresponding univariate AIC functions are shown for the combination of original E+N components ( $AIC^H$ ) and for the rotated components Q ( $AIC^Q$ ) and T ( $AIC^T$ ). AR-AIC picks  $t_{SAQ}$  and  $t_{SAT}$  derived from the minimum on the AIC functions agree very well with the actual arrival of the S wave visible on the seismograms. The uncertainty of the AR-AIC pick is expressed by the earliest and latest possible arrival times  $t_{SEQ}$ ,  $t_{SET}$ ,  $t_{SIQ}$ , and  $t_{SIT}$  derived from intersection of threshold  $thr_{AIC}$  (dashed horizontal lines) with the corresponding AIC functions. Copy of Fig. 4 in Diehl et al. (2009b, p. 1911) with © granted by the Seismological Society of America.



**Fig. 16.20** Examples of automatic S-wave picks at epicentral distances dominated by first arriving Sg phases (left-hand column) and first arriving Sn phases (right-hand column) for different error intervals. The error interval derived from the automatic quality assessment is represented by the vertical gray bars. The vertical long black bars denote the mean position of the S-wave onset. Error interval and mean position agree very well with the actual S-wave arrival observed on the seismograms. Copy of Fig. 5 in Diehl et al. (2009b, p. 1914) with © granted by the Seismological Society of America.

## 16.5 Automated quality assessment, phase identification, and outlier detection

Resolution and reliability of travel-time based inversion techniques, like hypocenter determination and tomography, depend strongly on the consistency of data quality weighting. Arrival times with larger uncertainties have to be down weighted or even rejected in the inversion process. Uncertainties of arrival time picks are traditionally divided into discrete quality classes (e.g. 0 to 4, with 0 corresponding to highest quality and 4 corresponding to lowest quality). Each quality class is associated with a certain weight (between 1 and 0) and should correspond to a measured uncertainty interval (in seconds). Such discrete quality classification is still used in many location and tomography algorithms like HYPOINVERSE (Klein, 2002) or VELEST (Kissling, 1988), which convert the quality class to a data weight during inversion. Quantitative uncertainty measures are usually absent for arrival times in global bulletins. Instead, picks are simply classified as “impulsive”, “emergent”, or “questionable” based on wavelet characteristics. Such qualitative error assessment, however, no longer satisfies the resolution capability of modern digital waveform data.

Although often disregarded, the uncertainty of an arrival time pick consists of two components: uncertainty of the phase timing and uncertainty of the phase identification (e.g. Diehl et al. 2009b). Automated picking and association of later phases is challenging, but provides fundamental information on earth structure and hypocenter locations. Especially phases like PmP, S, pP, and sP are crucial to constrain focal depth of local and teleseismic earthquakes.

In this section, we discuss procedures for automatic quality assessment of phase timing and phase identification. Because automatic quality assessment sometimes overestimates the timing quality or misinterprets the phase, outlier picks have to be detected in post-picking procedures as described at the end of this section.

### 16.5.1 Quality assessment of phase timing

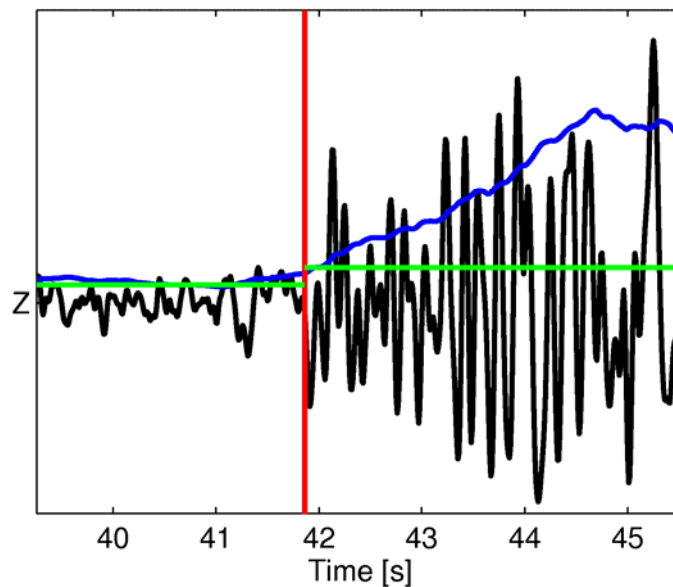
Arrival time estimation of phases needs to be accompanied by uncertainty estimates in order to down weight or even reject uncertain picks. The uncertainty of picks may be increased by low signal to noise ratios or by emergent onsets in contrast to impulsive ones. Emergent onsets are caused by the character of the source time function of the earthquake, the radiation pattern or by complicated wave propagation. Until now, only few papers focus on automatic quality assessment (e.g. Allen 1978; Aldersons 2004; Di Stefano et al. 2006; Diehl et al. 2009a and b; Nippres et al. 2010; Küperkoch et al. 2010) though automated arrival time determination should always be supplemented by automated uncertainty estimations.

Most widespread is the use of the signal-to-noise ratio (SNR) for automatic quality control and uncertainty estimation. Using certain window lengths depending on the dominating frequency, the amplitudes or the energy content right before the considered onset is compared to the amplitudes or energy content right after the onset (e.g. Sleeman and van Eck, 1999). If the SNR is calculated in the frequency domain, power spectral densities are compared (e.g. Leonard and Kennett, 1999). However, even bursts of noise may lead to large SNR, resulting in high qualities for false picks. To overcome this problem, it is necessary to estimate the signal length, which can be done by exploiting the envelope function (see Fig. 16.21) or by counting consecutive zero crossings of previous amplitudes, which exceed a certain threshold, as suggested e.g. by Allen (1978, see sub-section 16.3.1).

In addition to criteria for the SNR and the signal length, Allen (1978) uses a quite sophisticated algorithm for estimating the quality, based on analyses of the seismogram as well as of the CF (see sub-section 16.3.1). Küperkoch et al. (2010) suggest using the slope of the CF (sub-section 16.3.3) to identify emergent onsets. The MannekenPix algorithm of Aldersons (2004) includes a pattern recognition scheme, which weights nine different waveform attributes (predictors) obtained in the time window around the automatic pick and classifies the pick in discrete quality classes. The corresponding weighting factors are called “Fisher coefficients”, which have to be calibrated with a set of manually picked arrival times (reference picks).

A multiple discriminant analysis (MDA) is used to derive appropriate Fisher coefficients from the reference picks. Four predictors characterize the SNR. Predictor 5 is the difference between the dominant frequency of the signal and the dominant frequency of the noise. The remaining 4 predictors characterize properties of the CF around the estimated arrival time. This elaborated picking algorithm has been successfully applied to local earthquake data of the Dead Sea region (Aldersons, 2004), Italy (Di Stefano et al., 2006) and the greater Alpine region, respectively (Diehl et al., 2009a).



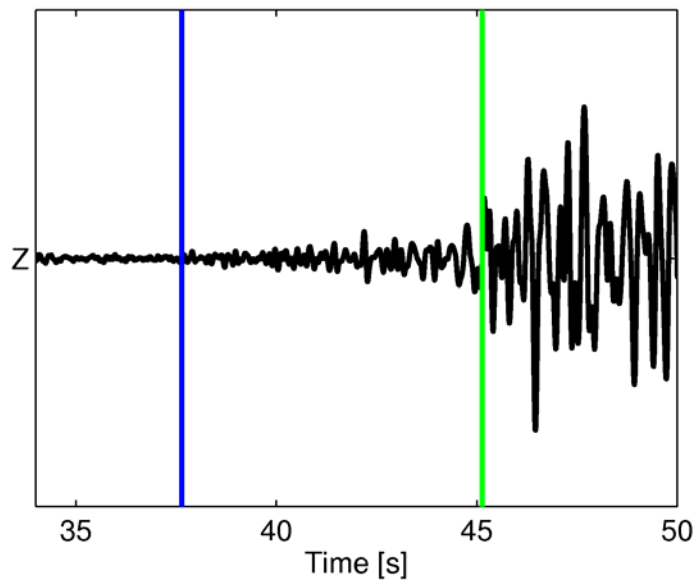


**Fig. 16.21** Short-period seismogram, 2-10 Hz bandpass filtered (black), and corresponding envelope function (blue). The envelope function is used to confirm the automatically derived P onset. The noise level is calculated within a window (green horizontal line) right before the automatically derived P onset (red vertical line). For this regional waveform example the requirement is that at least 70% of the envelope within a window of 4 seconds should exceed the signal threshold (green vertical line after P onset), which is twice the noise level.

### 16.5.2 Phase identification

Automatically determined arrivals have to be identified, that means, phase names have to be assigned to the arrival times, indicating their fundamental type and path through the earth. Most automatic approaches are intended to pick the first arriving phase of either P or S branches. Therefore, any automatic pick is considered as first arriving phase. In practice, picking algorithms can miss the very first phase and misinterpret later phases or larger amplitudes in the coda as the first arrival. Such phase misidentification may result in errors in the arrival times of up to several seconds, as pointed out e.g. by Diehl et al. (2009a). Fig. 16.22 shows an example of a regional event recording, where Pn is picked by the analyst, while the automated procedure picked a more prominent later phase (likely PmP). In case of first-arrival studies, picks should be accompanied by first-order estimates on whether arrival time is associated with first or later arriving phase. So far, most algorithms do not include such uncertainty estimates of phase identification, however, the assessment of phase timing can be tuned to pick targeted phases as described e.g. by Diehl et al. (2009a).

Explicit phase association of arrival times is usually only possible in combination with *a priori* information on the Earth's structure and the location of the event. Phase identification, however, can be accomplished by an iterative procedure: a preliminary location is determined using first arriving P phases, then first (and later arrivals) are associated by comparison with theoretical arrival times. In addition, epicentral distances can be estimated from S-P times and polarization analysis may support the identification of later phases (see section 16.4).



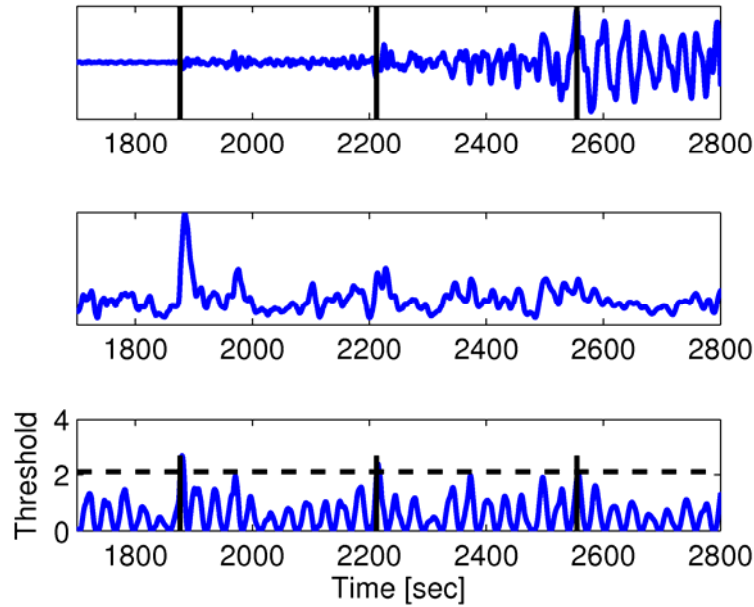
**Fig. 16.22** Example of phase misidentification within a 2-10 Hz bandpass filtered record. First arriving Pn-phase (blue) and intermediate phase (likely Pg) is missed by the automatic approach using kurtosis. Instead, the automatic approach (green) picks a later phase (likely PmP). Such misidentification may result in errors of several seconds.

Recordings of local to regional earthquakes are dominated by crustal phases like Pg, Pn, and PmP. Later arrivals are often buried in the P and S coda and may occur within short time windows. In addition, the correct identification is challenging due to the heterogeneous structure of the crust, which leads to significant travel time residuals, attenuation, scattering, and interference of these phases. In general, identification of first and later arriving crustal phases requires good knowledge of the three-dimensional structure of the crust. On the other hand, teleseismic events are dominated by mantle and core phases (see Chapter 11) and, as discussed later, signal properties like amplitude ratios between components, polarization or slowness as well as travel time differences between phases may be used to identify phases in the record. Global one-dimensional travel time tables can be used to predict and associate arrival times with phases. Crucial later arriving phases like pP can be associated with probability models as e.g. implemented in the EHB (Engdahl, van der Hilst, Buland) relocation procedure of Engdahl et al. (1998).

*In situ* automatic identification of various phases of a seismic record requires detection algorithms sensitive to later arrivals. Earle and Shearer (1994) propose a detection algorithm based on STA/LTA ratios calculated from the envelope function using the Hilbert transform of the seismogram. In order to prevent rapid fluctuations in the ratio function, the ratio function is smoothed by convolution with a Hanning filter. They applied this algorithm to seven years of global data distributed by the National Earthquake Information Center (NEIC). A comparison with the ISC hand-picked phase arrival times encounters difficulties with phase arrivals characterized by a change in frequency and little or no change in amplitude. However, the proposed algorithm is a useful tool for extracting seismograms containing phases desired for further analysis, e.g. for phase association. Fig. 16.23 shows an application of the proposed algorithm to a regional event recording.

Tong and Kennett (1995) and Bai and Kennett (2000) developed an algorithm, which combines automated detection and phase association for first and later arrivals. The proposed procedures are rather sophisticated and it is beyond the scope of this manual to provide a detailed description. The algorithms include many components useful for development of future generations of automatic algorithms (e.g. data adaptive filtering, analysis of three-component records, pattern recognition, etc.) and therefore we summarize only the main features of their approaches here.

Tong and Kennett (1995) define a set of STA/LTA detectors that make use of all three components and compare STAs and LTAs from different components in order to detect later phases. They use an approach of continual updating (Tong, 1995), where the effective lengths of the STA and LTA windows are adapted to the dominating frequency. In addition, they propose optimized filters to enhance the possibility of later phase detections and detection on multiple frequency bands (see also sub-section 16.6.2). Ratios of the energy on the different components, the frequency content, as well as polarization properties are used to characterize and to identify detected phases. Three measures of the energy of the phases are introduced:



**Fig. 16.23** Application of the proposed algorithm by Earle and Shearer (1994) for phase detection to a regional event recording at distance  $\Delta = 3792$  km (blue). Top: 0.01 – 0.1 Hz bandpass filtered waveform with recognized arrivals (vertical, black lines). Middle: corresponding STA/LTA ratios. Bottom: Smoothed ratio function (SRF, blue), applied threshold (dashed horizontal line) and detections (vertical black lines). This algorithm is useful for the detection of later arrivals rather than the precise arrival time determination.

1. the total energy  $E_3 = Z^2 + N^2 + E^2$ ,
2. the vertical component energy  $V_E = Z^2$ ,
3. the energy in the horizontal plane  $H_E = N^2 + E^2$ .

The LTA measure is calculated for the total energy  $E_3$ , while STA terms are calculated for the energy on each of the  $Z$ ,  $N$ , and  $E$  components as well as for the horizontal energy  $H_E$ , and the

total energy  $E_3$ . This results in a set of five different STA/LTA triggers. The output of the STA/LTAs is used to detect as well as to identify phases. Tong and Kennett (1996) define a range of empirical conditions to identify the fundamental type of the detected phase as summarized in Tab. 16.4.

**Tab. 16.4** Empirical conditions to identify the fundamental type of detected (later) phases proposed by Tong and Kennett (1996). Distance range far-regional to teleseismic is defined by slowness less than  $0.12 \text{ km}^{-1}$ .

| P wave<br>(far-regional to teleseismic) | SV-wave<br>(far-regional to teleseismic) | P wave<br>(regional) |
|-----------------------------------------|------------------------------------------|----------------------|
| $H_E < 0.50E_3$                         | $H_E > 0.50E_3$                          | $H_E > 0.50E_3$      |
| $V_E > 0.35E_3$                         | $V_E < 0.35E_3$                          | $V_E > 0.35E_3$      |

Furthermore, simple indicators are introduced to characterize the waveform of the phases:

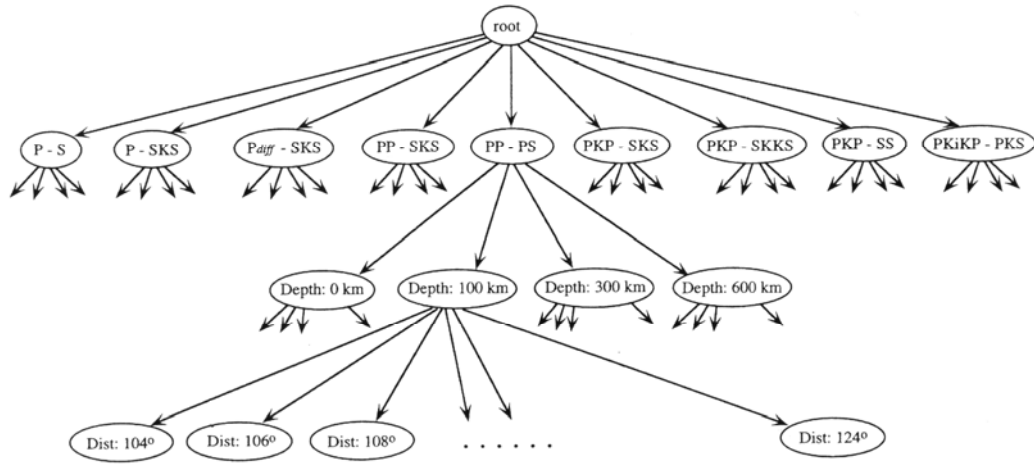
$$\begin{aligned}\tilde{P} &= V_E - 0.35E_3, \\ \tilde{S} &= H_E - 0.5E_3.\end{aligned}$$

The set of five STA/LTA phase detectors is applied to the data adaptive high-pass (corner frequency: two times the dominant background noise frequency) as well as low-pass (corner frequency: half the dominant frequency background noise) filtered data.

Polarization properties of the phases are characterized in terms of azimuth and incidence angles of a "phase vector", which is constructed to describe the average behavior of the phase over a quarter cycle of the dominant period. Using these attributes Tong and Kennett (1996) developed a procedure for the identification of phases. It is assumed that the following six attributes are available for the each phase  $i$ :

1.  $t_i$ , the arrival time;
2.  $a_i$ , the amplitude;
3.  $\nu_i$ , the local frequency;
4.  $\phi_i$ , the azimuth in the horizontal plane;
5.  $\psi_i$ , the angle of incidence to the vertical;
6.  $c_i$ , the P or S type of the arrival.

An assumption tree is used for further identification (Tong and Kennett, 1996). In analogy to rules of thumb employed by expert seismologists a heuristic method is suggested. The phase identification follows the assumption tree given in Fig. 16.24, starting with key phases like P and S first arrivals. At each level of the assumption tree, a set of assumptions about a single factor is tested.



**Fig. 16.24** Assumption tree for seismic event classification hierarchy (after Tong and Kennett, 1996).

At each node, the information of its parent node is gathered and includes extra information, starting with just the observed data. In case of contradictions, the branching associated with the current node is terminated. Otherwise the branching process continues with the evaluation of further hypothesis. When no further assumptions are to be tested, a solution is represented based on a set of hypothesis describing the data.

**Tab. 16.5** Ranges of travel time differences between P and S phases for different epicentral distances for the heuristic method proposed by Tong and Kennett (1996).

| P-S Identity           | Time Interval Range [min] | Distance Range [ $^{\circ}$ ] |
|------------------------|---------------------------|-------------------------------|
| P-S                    | 2.18-10.43                | 12-85                         |
| P-SKS                  | 9.37-10.65                | 82-99                         |
| P <sub>diff</sub> -SKS | 9.39-10.69                | 100-129                       |
| PP-SKS                 | 3.96-7.22                 | 82-129                        |
| PP-PS                  | 9.20-10.14                | 104-125                       |
| PKP-SKS                | 5.84-7.21                 | 114-143                       |
| PKP-SKKS               | 7.97-12.66                | 126-180                       |
| PKP-SS                 | 20.05-27.66               | 136-180                       |
| PKIKP-PKS              | 3.35-3.60                 | 126-141                       |

### 16.5.3 Post-picking and outlier detection

As mentioned before, even the most sophisticated automated algorithms introduce, to some degree, erroneous picks. Phase misinterpretation as shown in Fig. 16.18 or overestimated pick uncertainties are usually inevitable. In the following, we present tools to detect outliers in post-picking procedures.

Most commonly, erroneous picks are identified by large residuals in the event location routine. Frequently applied location routines like HYPOINVERSE (Klein, 2002) or

HYPOSAT (Schweitzer, 2001) are able to exploit additional information like phase names, uncertainty estimates, and other phase properties such as polarization angles in order to identify inconsistent picks. Some of these routines automatically down weight observations associated with large residual. Although it is a common practice, we point out that the velocity model used for location biases residual weights. Therefore, residual weights from location routines should never replace the observation weights assigned by the picker. Otherwise, correct picks might be down weighted due to an insufficient velocity model and might be useless for tomographic studies.

Furthermore, the presence of few outlier picks can have a significant influence on the hypocenter solution derived by traditional least-square approaches. Beamforming techniques as described e.g. by Pinsky (2006) or the Equal Differential Time (EDT) formulation implemented in the NonLinLoc location routine of Lomax et al. (2000) yield very robust hypocenter solutions even in the presence of outliers and may therefore also improve the detection of false picks. In practice, automatic picks associated with large residual are usually flagged in the location routine and might be reviewed by the network analyst.

Identifying outliers in the location routine requires the presence of an appropriate velocity model of the study area. Since automatic pickers are increasingly used to derive such models, simultaneous inversion of hypocenters and seismic velocities have to be performed to detect outlier picks. Such iterative inversion procedures are proposed e.g. by Diehl et al. (2009a). Finally, Wadati diagrams as described, for example, by Kisslinger and Engdahl (1973) or Maurer and Kradolfer (1996) are used to identify mispicked S wave arrivals, independent of the P and S wave velocity structure.

## **16.6 Practical considerations: implementation, calibration, and pitfalls of automatic detection and picking procedures**

Implementation and realization of automatic detection and picking procedures depend on the application purpose and the available data. Commonly, these automatic procedures are used in permanent seismic weak-motion networks to detect and locate events in real or near-real time and the focus is on low detection thresholds in order to generate earthquake catalogs as complete as possible. However, increasingly, such automatic procedures are nowadays also applied on records from very dense strong-motion accelerometer networks such as the Japanese K- and KiK nets (see Chapter 8, sub-section 8.7.3). They trigger events on higher detection thresholds. Therefore, these records are, as compared to very sensitive weak-motion records, less or not at all effected by the preceding microseismic noise. Nevertheless, sophisticated algorithms are used to identify phase onset times. E.g. Akazawa (2004) uses a combination of STA/LTA ratios of a cumulative envelope function and the AR-AIC algorithm to exploit strong motion records of the K-, KiK-, and the CEORKA (Committee of Earthquake Observation and Research in the Kansai Area) networks. In any event, the calculated automatic locations have to be robust and available within minutes after an event. Finally, network applications have to handle signals in a broad frequency range in order to detect and pick events from local to teleseismic distances.

A second class of applications is the use within the increasing number of temporary deployments (aftershock deployments, temporary field deployments, etc.), which generate huge data volumes. Furthermore, the number of available analysts is usually limited. Focus for this class of applications is variable, but usually includes high detection rates for local to

regional events. Computing time of detection, picking, and location is less crucial and teleseismic events are already identified by global bulletins.

Finally, application of “re-picking” studies become increasingly important, in which automatic algorithms are used to improve consistency and quality of existing network picks, or to merge picks of several networks. For this class, event detection is usually not necessary and focus is on high-quality picks. Preliminary locations in combination with available earth models can be used to predict windows of first and later arriving phases. If arrival times are required for high-resolution tomography, consistency of picks is favored over hit-rate of algorithms, since completeness is less relevant. Numbers of erroneous picks, however, have to be as small as possible since residuals of outliers in the post-picking detection might be of the same order than residuals due to velocity anomalies. A suite of iterative detection stages might precede this “high-quality” application.

No matter which algorithm is chosen, automatic procedures have to be calibrated and tested with any new data set, i.e. the parameters of the algorithms have to be adjusted and optimized towards the application purpose. In this section we describe calibration and test procedures for automatic pickers. Furthermore, the pre-processing and quality of waveform data is crucial for reliable automatic procedures. Therefore, common pitfalls related to waveform filters and waveform quality are discussed at the end of this section.

### **16.6.1 Calibration and test of automatic pickers**

Typically, calibration and test of automatic algorithms is achieved by comparison with a set of representative manually picked onset readings (e.g. Baer and Kradolfer 1987; Sleeman and van Eck 1999; Di Stefano et al. 2006). A common practice is the use of network picks (i.e. arrival times routinely picked by network analysts on a day-to-day basis) as reference for the automatic algorithms. These network picks, however, are usually not appropriate to serve as reference picks, because they yield a high level of noise due to mispicks and other inconsistencies, particularly in error assessment. Reference arrival times for calibration and meaningful tests of automatic algorithms have to be picked consistently. In addition, reference picks have to be provided with a consistent measure of uncertainty in terms of phase timing (see sub-section 16.5.1) and phase identification (see sub-section 16.5.2). Seismograms with no identifiable phases (e.g. due to low signal-to-noise ratio) have to be flagged, in order to test the automatic picker’s ability to reject low-quality signals. Finally, consistent use of waveform filters during reference picking is necessary to avoid biases between automatic and manual picks. Additional use of theoretical arrival times, component rotation and tools like polarization analysis might be necessary for reliable reference picking of S waves. A detailed description on reference picking can be found in Diehl et al. (2009a) and IS 11.4. The set of reference picks should be divided in two subsets. The first subset is used to calibrate the algorithm and subsequently, the calibrated algorithm is tested with the other half.

Calibration of algorithms (i.e. adjustment of parameters) is usually accomplished by trial-and-error methods. Parameters are adjusted in order to minimize the difference between automatic and reference picks. As mentioned earlier, the trade-off between hit rate, precision, and number of outliers depends on the application purpose. Additional information on automatic quality assessment can be included in the performance evaluation using a matrix representation as suggested by Di Stefano et al. (2006). The quality assessment of the automatic picker must be tuned to get quality classes, which are similar or even equal to the

manually derived onset weights. Because automated algorithms usually evaluate the weighting classes differently than the analysts (e.g. using SNR determined from the seismogram and/or the CF, slope of CF, combination of both, etc.), the aim of exactly matching manual and automatic weighting classes is hard to achieve. In general, a sufficient calibration is obtained if the automatic picker upgrades only a very few low-quality reference picks to top quality classes and the majority of the high-quality reference picks is correctly classified as top quality. If only high-quality picks are requested (e.g. high-resolution tomography or high-precision relocation studies), the automatic quality assessment must be tuned such that none of the lowest-quality picks is classified as highest-quality (e.g. Di Stefano et al. 2006; Diehl et al. 2009a). As a consequence, a lot of intermediate-quality picks are rejected, resulting in an incomplete but high-quality set of arrival times. In contrast, for coarse location purposes the picker must be trained towards increased hit rate of intermediate-onset readings to assure a sufficient number of observations and azimuthal coverage for robust event location. Finally, to ensure that the calibrated algorithm is appropriate for the entire data set (and not just for the subset used for calibration), it should be tested with a second subset of reference picks as described earlier.

Sophisticated packages like MannekenPix (Aldersons 2004) or the algorithm proposed by Nippres et al. (2010) make use of data adaptive calibration in order to improve performance of picking and quality assessment and to avoid tedious trial-and-error procedures. Parameters are determined from the reference picks by an inverse technique based on pattern recognition. In practice, most automatic algorithms require a minimum of parameter search to determine the optimized choice of essential (data depending) picking parameters, such as search-window definitions or waveform filters.

### **16.6.2 Pre-processing: waveform filters**

The estimation of arrival times depends on the chosen frequency band and differences in arrival times determined in different frequency ranges may be in the order of several seconds. The effect of waveform filters on the arrival time picking is illustrated in Fig. 16.25. On the other hand, application of filters often enhances the signal-to-noise ratio. High-pass filters are used to remove microseismic noise and low-pass filters are commonly used to eliminate anthropogenic noise. To minimize the bias introduced by inconsistent application of filters, only one frequency band should be used for picking. Filters should be of zero-phase-type to avoid systematic time shifts. For a consistent analysis of arrival times, filters used for reference picking should be identical with filters used during automatic picking.

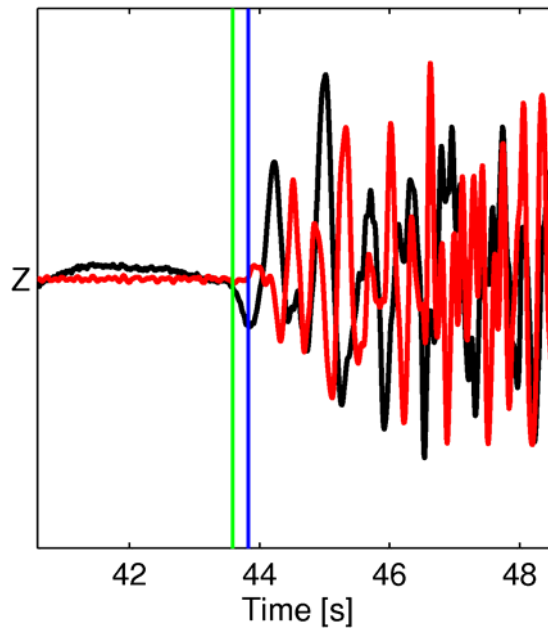
In practice, however, the use of one fixed frequency band for picking is often inapplicable, especially when dealing with data of earthquakes from wide distance and magnitude ranges. Differences in frequency content of seismic signals are mainly caused by the source processes and to minor extend by phase shifts due to wave propagation (attenuation, finite frequency effects). Finite rupture processes of large earthquakes lead, especially in the near field, to very complex signals, dominated by low frequencies. Simpler signals are expected for the point-source characteristics of small and moderate magnitudes. Frequency content varies for local Pg, regional Pn, and teleseismic mantle or core phases and might even change with focal depth. Finally, the noise characteristic can be station dependant, especially for regional and global networks.



To account for the diversity in signal and noise characteristic, data adaptive filters like the Wiener filter can be used to enhance the signal-to-noise ratio of seismic data. The Wiener filter, as implemented in MannekenPix (Aldersons 2004), measures the power spectrum in two windows:  $P_N(f)$  is determined from a window containing background noise and  $P_{SN}(f)$  is measured in a window containing signal and noise. As described by Aldersons (2004) the Wiener filter can be approximated by:

$$W(f) \approx \frac{P_{SN}(f) - P_N(f)}{P_{SN}(f)}$$

The obvious advantage of the Wiener filter is that it affects only frequencies associated with the background noise. In the ideal case, the filter does not modify the signal part, in contrast to traditional high-, low-, or band-pass filters. This assumption, however, is only valid if the signal-spectrum does not overlap with the noise-spectrum. Implementations like in MannekenPix require *a priori* information on the approximate position of the onset (initial pick from predicted arrival time or some sort of preliminary pick) to properly setup the noise and signal+noise windows. A gap between both windows has to account for the uncertainty of the initial picks and the length of the windows have to be appropriate for the expected noise and signal frequencies. Improper choice of these parameters might introduce additional instabilities to the Wiener filter.



**Fig. 16.25** Unfiltered (black) and filtered (red) local event waveform of a broadband record (Guralp CMG, 60 s – 50 Hz) using a 3<sup>rd</sup> order butterworth bandpass (2-10 Hz) and corresponding manual (blue vertical line) and automatic picks (green vertical line) indicating the effect of different filtering on P-onset determination. Note that in the more narrow-band high-frequency filtered record the amplitude of the first oscillations is strongly reduced, the negative first motion, still distinct in the broadband record, no longer recognizable and the phase of the oscillations significantly shifted. The reasons for these signal distortions due to filtering are discussed in Chapter 4.

A simple data adaptive filter is proposed by Tong (1995), who determines the dominant frequency  $f_n$  of the background noise and defines low- and high-pass filters based on  $f_n$ . Picking on multiple (fixed) frequency bands as well as filters depending on epicentral distance or station site might be used to account for differences in the signal and noise characteristics.

### 16.6.3 Pre-processing: waveform quality

The quality of available waveform data is crucial for reliable and robust automatic detection and picking procedures. “Glitches” present in time series can generate a high number of false picks and have to be identified and, if possible, removed prior to the picking. Common glitches are spikes, abrupt offsets in amplitudes (caused by data gaps or sensor re-centering), clipped amplitudes, and precursory oscillations prior to impulsive onsets (caused by instrumental acausal anti-alias filters). If filters are applied to these glitches and if they occur close to a seismic event, they are indistinguishable from real signals for most automatic algorithms. Spikes in the data can be identified and removed by running average routines as implemented e.g. in the SAC software. Clipped amplitudes usually inhibit reliable S wave detection and therefore, clipped seismograms have to be identified and removed prior to automatic S picking as described e.g. by Diehl et al. 2009b. Effects of acausal anti-alias filters can be removed by an inverse filtering process described e.g. in Scherbaum (2001) or may be minimized by application of certain low-pass filters. Problems with data quality are usually specific to networks and procedures to detect and remove them have to be developed and adjusted from case to case. Therefore, tests with reference data have to be used to identify these problems and to develop and calibrate tools to remove them.

## Appendix

Some internet addresses providing source codes or binaries of some picking algorithms:

- Several seismological software and useful links:  
<http://www.orfeus-eu.org/Software/software.html>
- MannekenPix (MPX):  
<http://faldersons.net/Software/MPX/MannekenPix.html>
- S-picker by Diehl et al. (2009b):  
<http://www.ldeo.columbia.edu/~tdiehl/Data2Download/spicker1.3.4.publ.tar.gz>

## Acknowledgments

The authors would like to thank Peter Bormann, Joachim Saul, and Reinould Sleeman for careful proof-reading and valuable suggestions.

## References

Akaike, H. (1970). Statistical predictor estimation. *Ann. Inst. Statist. Math.*, **22**, 203-217.

- Akaike, H. (1971). Autoregressive model fitting for control. *Ann. Inst. Statist. Math.*, **23**, 163-180.
- Akaike, H. (1974). Markovian representation of stochastic process and its application to the analyses of autoregressive moving average processes. *Ann. Inst. Statist. Math.*, **26**, 363-387.
- Akaike, H. (1974b). A new look at the statistical model identification. *IEEE Transactions on Automatic Control*, **19** (6), 716-723.
- Akazawa, T. (2004). A technique for automatic detection of onset time of P- and S-phases in strong motion records, *13<sup>th</sup> World Conference on Earthquake Engineering*, Vancouver, B.C., Canada, Paper No. 786.
- Aldersons, F. (2004). Toward three-dimensional crustal structure of the Dead Sea region from local earthquake tomography. *PhD thesis*, Tel Aviv University, Israel.
- Allen, R. V. (1978). Automatic earthquake recognition and timing from single traces. *Bull. Seism. Soc. Am.*, **68**, 1521-1532.
- Allen, R. V. (1982). Automatic phase pickers: Their present and future prospects. *Bull. Seism. Soc. Am.*, **72**, 225-242.
- Baer M., and Kradolfer U. (1987). An automatic phase picker for local and teleseismic events. *Bull. Seism. Soc. Am.*, **77**, 1437-1445.
- Bai, Chao-ying, and Kennett, B. L. N. (2000). Automatic phase-detection and identification by full use of a single three-component broadband seismogram. *Bull. Seism. Soc. Am.*, **90**, 187-198.
- Basseville, M., and Nikiforov, I. V. (1993). Detection of abrupt changes: Theory and application. *Prentice Hall Information and System Science Series*, Prentice Hall, Englewood Cliffs, New Jersey.
- Box, G. E. P., Jenkins, G. M., and Reinsel, G. C. (1994). Time series analysis - Forecasting and control. *Prentice-Hall*.
- Burg, J. P. (1975). Maximum entropy spectral analysis. *PhD thesis*, Department of Geophysics, Stanford University.
- Cichowicz, A. (1993). An automatic S-phase picker, *Bull. Seism. Soc. Am.*, **83**, 1, 180-189.
- Diehl, T., Kissling, E., Husen, S., Aldersons, F. (2009a). Consistent phase picking for regional tomography models: application to the greater Alpine region. *Geophys. J. Int.*, **176**, 542-554.
- Diehl, T., Deichmann, N., Kissling, E. Husen, S., 2009b. Automatic S-wave picker for local earthquake tomography. *Bull. Seism. Soc. Am.*, **99**(3), 1906-1920.
- Dietz, L. (2002). Notes on Configuring BINDER\_EW: Earthworm's Phase Associator; [http://www.isti2.com/ew/ovr/binder\\_setup.html](http://www.isti2.com/ew/ovr/binder_setup.html).
- Di Stefano, R., Aldersons, F., Kissling, E., Baccheschi, P., Chiarabba, C., Giardini, D. (2006). Automatic seismic phase picking and consistent observation error assessment: Application to the Italian seismicity. *Geophys. J. Int.*, **165**, 121-134.
- Douglas, A., Bowers, D., and Young, J. B. (1997). On the onset of P seismograms. *Geophys. J. Int.* **129**, 681-690.
- Earle, P. S., and Shearer, P. M. (1994). Characterization of global seismograms using an automatic-picking algorithm. *Bull. Seism. Soc. Am.*, **84**, 366-376.
- Engdahl, E. R., Van der Hilst, R. D., and Buland, R. P. (1998). Global teleseismic earthquake relocation with improved travel times and procedures for depth determination. *Bull. Seism. Soc. Am.*, **88**, 722-743.
- Freiberger, W. F. (1963). An approximation method in signal detection. *Quart. J. App. Math.* **20**, 373-378.
- Gentili, S., and Bragato, P. (2006). A neural-tree-based system for automatic location of earthquakes in Northeastern Italy. *J. Seismology*, **10**, 73-89.

- Gentili, S., and Michelini, A. (2006). Automatic picking of P and S phases using a neural tree. *J. Seismology*, **10**, 39-63.
- Goforth, T., and Herrin, E. (1981). An automatic seismic signal detection algorithm based on the Walsh transform. *Bull. Seism. Soc. Am.*, **71**, 1351-1360.
- Gomberg, J. S., Shedlock, K. M., and Roecker, W. W. (1990). The effect of S-wave arrival times on the accuracy of hypocenter estimation. *Bull. Seism. Soc. Am.*, **80**, 6, 1605-1628.
- Groos, J. C., and Ritter, J. R. R. (2009). Time domain classification and quantification of seismic noise in an urban environment. *Geophys. J. Int.*, accepted for publication.
- Hartung, J. (1991). Statistik - Lehr- und Handbuch der angewandten Statistik. *Oldenbourg Publishing Company*, Vienna, Munich.
- Hildyard, M. W., Nippres, S. E. J., and Rietbrock, A. (2008). Event detection and phase picking using a time-domain estimate of predominant period,  $T^{pd}$ . *Bull. Seism. Soc. Am.*, **98**(6), 3025-3032.
- Hildyard, M. W., and Rietbrock, A. (2010).  $T^{pd}$ , a damped predominant period function with improvements for magnitude estimation. *Bull. Seism. Soc. Am.*, **100**(2), 684-698.
- Johnson, C. E., Bittenbinder, A., Bogaert, B., Dietz, L., and Kohler, W. (1995). Earthworm: A Flexible Approach to Seismic Network Processing. *IRIS Newsletter*, **14**, 2, 1-4.
- Joswig, M. (1987). Methoden zur automatischen Erfassung und Auswertung von Erdbeben in seismischen Netzen und ihre Realisierung beim Aufbau des lokalen 'Bochum University Germany'-Netzes. *Berichte des Instituts für Geophysik der Ruhr-Universität Bochum*, Reihe A, No. 23.
- Kao, H., and Shan, S.-J. (2004). The source-scanning algorithm: mapping the distribution of seismic sources in time and space. *Geophys. J. Int.*, **157**, 589-594.
- Kissling, E. (1988). Geotomography with local earthquake data. *Rev. Geophys.*, **26**, 659-698.
- Kisslinger, C., and Engdahl, E. R. (1973). The interpretation of the Wadati diagram with relaxed assumptions. *Bull. Seismol. Soc. Am.*, **63**, 1723-1736.
- Kitagawa, G., Takanami, T., and Matsumoto, N. (2001). Signal extraction problems in seismology. *Int. Statist. Rev.*, **69**(1), 129-152.
- Klein, F. (2002). User's Guide to HYPOINVERSE-2000, a Fortran program to solve for earthquake locations and magnitudes, *USGS, Open File Report 02-171*, 123pp.
- Klein, F. W. (2003). The HYPOINVERSE2000 earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam*, 1619-1620.
- Küperkoch, L., Meier, T., Friederich, W., and EGELADOS working group (2010). Automated P-wave arrival time determination using higher order statistics. *Geophys. J. Int.*, **181**(2), 1159-1170; doi: 110.1111/j.1365-246X.2010.04570.x.
- Lay, T., and Wallace, T. C. (1995). *Modern global seismology*. ISBN 0-12-732870-X, *Academic Press*, 521 pp.
- Le Bras, R., Swanger, H., Sereno, T., Beall, G., Jenkins, R., Nagy, W., and Henson, A. (1994). Global association; Final Report. *Science Applications International Corporation Technical Report*, SAIC-94/1155.
- Leonard, M. (2000). Comparison of manual and automatic onset time picking. *Bull. Seism. Soc. Am.*, **90**, 1384-1390.
- Leonard, M., and Kennett, B. L. N. (1999). Multi-component autoregressive techniques for the analysis of seismograms. *Phys. Earth Planet Int.*, **113**, 247-263.
- Lomax, A., Virieux, J., Volant, P., and Berge, C. (2000). Probabilistic earthquake location in 3D and layered models: Introduction of a Metropolis-Gibbs method and comparison with linear locations. In: Thurberand, C. H., and Rabinowitz, N. (2000). *Advances in Seismic Event Location. Kluwer, Amsterdam*, 101-134.

- Longbottom, J., Walden, A. T., and White, R. E. (1988). Principles and application of maximum kurtosis phase estimation. *Geophysical Prospecting*, **36**, 115-138.
- Maeda, N. (1985). A method for reading and checking phase times in autoprocesing system of seismic data. *Zisin=Jihin (J. Seism. Soc. Japan)*, **38**, 365-379.
- Maurer, H., and Kradolfer, U. (1996). Hypocentral parameters and velocity estimation in the western Swiss Alps by simultaneous inversion of P- and S-wave data. *Bull. Seismol. Soc. Am.*, **86**, 32-42.
- Michael, A. J., Gildea, S. P., and Pulli Jay, J. (1982). A real-time digital seismic event detection and recording system for network applications. *Bull. Seism. Soc. Am.*, **72**, 6, 2339-2348.
- Morita, Y., and Hamaguchi, H. (1984). Automatic detection of onset time of seismic waves and its confidence interval using the autoregressive model fitting. *Zisin*, **37**, 281-293.
- Nippres, S. E. J., Rietbrock, A., and Heath, A. E., (2010). Optimized automatic pickers: application to ANCORP data set. *Geophys. J. Int.*, **181**, 911-925.
- Pinsky, V. (2006). Using beamforming for the global network location. *Phys. Earth Planet Int.*, **158**, 75-83.
- Rawlinson, N., and Kennett, B. L. N. (2004). Rapid estimation of relative and absolute delay times across network by adaptive stacking. *Geophys. J. Int.*, **157**, 332-340.
- Rioul, O., and Vetterli, M. (1991). Wavelets and signal processing. *IEEE Signal Proc. Magazine*, October 1991, 14-38.
- Rowe, C. A., Aster, R. C., Borchers, B., and Young, C. J. (2002). An automatic, adaptive algorithm for refining phase picks in large seismic data sets. *Bull. Seism. Soc. Am.*, **92**, 5, 1660-1674.
- Saragiotis, C. D., Hadjileontiadis, L. J., and Panas, S. M. (2002). PAI-S/K: A robust automatic seismic P phase arrival identification scheme. *IEEE Trans. Geosci. Remote Sensing*, **40**, 1395-1404.
- Scherbaum, F. (2001). Of poles and zeros: Fundamentals of digital seismology. Modern Approaches in Geophysics, *Kluwer Academic Publishers*, 2<sup>nd</sup> edition, 265 pp. (including an CD-ROM by Schmidtke, E., and Scherbaum, F. with examples written in Java).
- Scherbaum, F., and Johnson, J. (1992). Programmable Interactive Toolbox for Seismological Analysis (PITSA). IASPEI Software Library Volume 5, Seismological Society of America, El Cerrito.
- Schweitzer, J. (2001). HYPOSAT – An enhanced routine to locate seismic events. *Pure and Appl. Geophys.*, **158**, 277-289.
- Sleeman, R., and van Eck, T. (1999). Robust automatic P-phase picking: an on-line implementation in the analysis of broadband seismogram recordings. *Phys. Earth Planet. Int.*, **113**, 265--275.
- Sleeman, R., and van Eck, T. (2002). Single station real-time P and S phase pickers for seismic observatories. In: Methods and applications of signal processing in seismic network operations. Takanami, T. and Kitagawa, G. (Eds.). Springer-Verlag, Berlin, 173-194.
- Stewart, S. W. (1977). Real-time detection and location of local seismic events in Central California. *Bull. Seism. Soc. Am.*, **67**, 2, 433-452.
- Takanami, T., and Kitagawa, G. (1988). A new efficient procedure for estimation of onset times of seismic waves. *J. Phys. Earth*, **36**, 267-290.
- Takanami, T., and Kitagawa, G. (1991). Estimation of the arrival times of seismic waves by multivariate time series model. *Ann. Inst. Stat. Math.*, **43**, 407--433.
- Tong, C. (1995). Characterization of seismic phases - an automated analyser for seismograms. *Geophys. J. Int.*, **123**, 937-947.

- Tong, C., and Kennett, B. L. N. (1995). Towards the identification of later seismic phases. *Geophys. J. Int.*, **123**, 948-958.
- Tong, C., and Kennett, B. L. N. (1996). Automatic seismic event recognition and later phase identification for broadband seismograms. *Bull. Seism. Soc. Am.*, **86**, 6, 1896-1909.
- VanDecar, and Crosson, R. S. (1990). Determination of teleseismic relative phase arrival times using multi-channel cross-correlation and least squares. *Bull. Seism. Soc. Am.*, **80**, No. 1, 150-169.
- Wang, J, and Teng, T. (1997). Identification and picking of S phase using an artificial neural network. *Bull. Seism. Soc. Am.*, **87**(5), 1140-1149.
- Withers, M. M., Aster, R. C., and Young, Ch. (1998). An automated local and regional seismic event detection and location system using waveform correlation. *Bull. Seism. Soc. Am.*, **88**, No. 3, 657-669.
- Zeiler, C., and Velasco, A. A. (2009). Seismogram picking error from analyst review (SPEAR): single-analyst and institution analysis. *Bull. Seism. Soc. Am.*, **99**, 5, 2759-2770.
- Zhang, H., Thurber C., and Rowe C. (2003). Automatic P-wave arrival detection and picking with multiscale wavelet analysis for single-component recordings. *Bull. Seism. Soc. Am.*, **93**, 1904-1912.

## Download programs and files

| Name               | Download programs and files                                                                                                                                  |
|--------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Compiled by</b> | <b>Peter Bormann</b> , (formerly GFZ German Research Centre for Geosciences, 14473 Potsdam, Germany); E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| <b>Version</b>     | November, 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_Downloads">10.2312/GFZ.NMSOP-2_Downloads</a>                                               |

### Introduction

The alphabetical listing below gives in bold letters the names of **Programs** or complementary educational **Files** (animations, PPTs, pdfs), mentioned in the NMSOP-2 texts, that are offered for free download.

Program folders may contain several files, such as program descriptions/manuals, sub-programs, source codes and executable files. Files or folders that are hyperlinked with the respective program archive administered by the GFZ library are written in **bold blue letters**. Authors of the programs and downloadable files have direct access via password (to be requested from the GFZ Chief Librarian, Mr. Roland Bertelmann, [rab@gfz-potsdam.de](mailto:rab@gfz-potsdam.de)) for uploading new or revised/upgraded programs or file versions.

Interested users may **download** the respective **Programs** or **Files** by left mouse click on the respective folder or file name typed in **bold blue letters**. **Animation and PPT files**, when typed in **bold blue letters**, may also **be activated** directly from the related NMSOP-2 texts via left mouse click on the file names.

Files that are part of a folder package are also listed below in alphabetical order, yet typed in **bold black letters**. They are downloaded together with the hyperlinked **folders in bold blue**.

**Note:** If animations are running for your purpose or comprehension too fast or too slow then catch the running movie cursor with a left mouse button click, hold the button pressed and then control the movie speed according to the time you need to read the texts and to follow the movie.

### Programs

[calex](#) (Author: E. Wieland; program package containing the program file **trical**)

[calibrat](#) (Author: J. Bribach; program package with program files **response**, **caliseis** and **seisfil**)

## Download programs and files

**caliseis** (Author: J. Bribach; program in [calibrat](#))

[dispcal](#) (Author: E. Wieland; program )

[filtdemo](#) (Author: E. Wieland; program)

[fourierdemo](#) (Author: E. Wieland, program)

[hyposat\\_hypomod](#) (Author: J. Schweitzer; program package for locating seismic sources)

[inverseif](#) (Author: E. Wieland; program)

[laufze\\_laufps](#) (Author: J. Schweitzer; program package for calculating travel-time curves for horizontally layered or spherical P- and/or S-velocity models)

[lincomb](#) (Author: E. Wieland; program package containing the programs **triax** and **rectac**)

[linreg3](#) (Author: E. Wieland; program package with folder **linregress**)

**linregress** (Author: E. Wieland; program in [linreg3](#))

[libraries](#) (Author: E. Wieland)

[lok-sept2011.tar](#) (source code; Authors: M. Živčić and J. Ravnik)

[noisecon](#) (Author: E. Wieland; program)

[optinet.ppt](#) (Author: P. Bormann; PPT presentation complementing IS 8.7)

[polzero](#) (Author: E. Wieland; program package with program files **resp** and **winresp**)

**rectax** (Author: E. Wieland; program in [lincomb](#))

**response** (Author: J. Bribach; program in [calibrat](#))

[sincal](#) (Author: E. Wieland; program package containing the program file **sinfit**)

**sinfit** (Author: E. Wieland; program in [sincal](#))

[seisan](#) (program package for the SEISAN earthquake analysis software; authors: J. Havkov, L. Ottemöller, and P. Voss)

**seisan.pdf** (Manual of version 9.0.1 of the [seisan](#) earthquake analysis software; authors: J. Havkov, L. Ottemöller, and P. Voss)

**seisan\_explorer** (zip folder in [seisan](#); authors: J. Havkov, L. Ottemöller, and P. Voss)



## Download programs and files

**seisan\_training\_data** (three zip, rar and tar folders in [seisan](#) with training data;  
authors: J. Havskov, L. Ottemöller, and P. Voss)

**seisan\_v9.0.1** (three [seisan](#) program files, tar.gz and msi, for Windows, Linux 32bit and  
Linux 64bit systems; authors: J. Havskov, L. Ottemöller, and P. Voss)

**seisan\_v9.0.2** (seisan program file for Solaris 32bit systems; authors: J.  
Havskov, L. Ottemöller, and P. Voss)

**seitrain.pdf** (Introduction to [seisan](#) and computer exercises in processing earthquake data;  
authors: J. Havskov, L. Ottemöller, and P. Voss)

**seisfilt** (Author: J. Bribach; program in [calibrat](#))

[sleeman](#) (program for determining instrumental noise of digitizers and seismic sensors;  
author: E. Wieland)

[tiltcal](#) (Author: E. Wieland; program)

**triax** (Author: E. Wieland; program in [lincomb](#))

**trical** (Author: E. Wieland; program in [calex](#))

**tricrosp** (Author: E. Wieland?; program in [sleeman](#))

**twocrosp** (Author: E. Wieland?; program in [sleeman](#))

[upgrade](#) (folder in [seisan](#) with the files **codaq.exe**, **codaq.for** and **read me**;  
authors: J. Havskov, L. Ottemöller, and P. Voss)

**winresp** (Author: E. Wieland; program in [polzero](#))

## Files (complementary educational material and data)

[3D-wave-prop travel-time qt](#) (Animation; local wavefronts and travel-time curves;  
Author: S. Wendt; see EX 11.1 )

[csem\\_newsletter ml scales](#) (pdf file of the CSEM-EMSC Newsletter of November 15,  
1999, with a compilation of different institutional procedures of magnitude determination)

[circles\\_depth](#) (Animation; local seismic wavefronts and source depth; Author: S. Wendt;  
see  
EX 11.1)

[gfz\\_ma\\_chile2010 rupture](#) (Animation of the seismic energy release in space and time  
during the 2010 Mw8.8 Chile earthquake; Author: Yanlu Ma, GFZ Potsdam; see IS 1.1)

[optinet:ppt](#) (PPT tutorial, complementary to IS 8.7; Author: P. Bormann)

[ohrnberger\\_sumatra2004\\_rupture](#) (Animation of the seismic energy release in space and time during the 2004 Mw9.3 Sumatra-Andaman earthquake; Authors: M. Ohrnberger and F. Krueger, 2005; see IS 1.1)

[reference events](#) (data folder compiled by S. Wendt)

[ref\\_evt\\_00\\_05.tar.gz](#) (data file in folder [reference events](#) by S. Wendt)

[ref\\_evt\\_06\\_09.tar.gz](#) (data file in folder [reference events](#) by S. Wendt)

[ref\\_evt\\_94\\_99.tar.gz](#) (data file in folder [reference events](#) by S. Wendt)

[s1.tar](#) (waveform data files for onset picking in the SPEAR project by C. Zeiler, see IS 11.5)

[s2.tar](#) (waveform data files for onset picking in the SPEAR project by C. Zeiler, see IS 11.5)

[wendt\\_fiji\\_qt](#) (Animation of ray propagation; Author: S. Wendt; see IS 11.3)

[wendt\\_hindukush\\_qt](#) (Animation; mantle and depth phases; Author: S. Wendt; see IS 11.3)

[wendt\\_india\\_qt](#) (Animation; rays of mantle and core phases; Author: S. Wendt; see IS 11.3)

[wendt\\_newbritain\\_qt](#) (Animation; rarely observed multiple core phases; Author: S. Wendt)

[wendt\\_nz\\_qt](#) (Animation; teleseismic seismic phases and wavefronts; Author: S. Wendt)

[wendt\\_peru\\_qt](#) (Animation, rarely observed core phases; Author: S. Wendt; see IS 11.3)

[wendt\\_russlchina\\_qt](#) (Animation; depth and other phases; Author: S. Wendt; see IS 11.3)

[wendt\\_slovenia\\_qt](#) (Animation; regional seismic wavefronts; Author: S. Wendt; see IS 11.3)

[wendt\\_source\\_migra\\_around](#) (Animation with text on the 2- and 3-D viewing of the development of an earthquake swarm in space and time in its relation to dominating regional tectonic trends; Author: S. Wendt; see IS 1.1).

[wendt\\_source\\_rad\\_pat](#) (Animation with text on the dependence of source directivity and radiation pattern of P- and S- wave amplitudes on distance, strike, dip and rake angle for different types of source mechanism; Author: S. Wendt; see IS 1.1 and Chapter 3)

[wendt\\_vogtland1\\_qt](#) (Animation; local seismic wavefronts; Author: S. Wendt; see IS 11.3)

[wendt\\_vogtland2\\_qt](#) (Animation; local seismic wavefronts; Author: S. Wendt; see IS 11.3)

## Download programs and files

[wendt\\_waveprop\\_circles](#) (Animation with text on the P- and S-wave propagation from a local earthquake through a regional seismic network, the formation of the related travel-time curves and seismic records at the seismic stations, the determination of the epicentre by means of the circle and chort method and of the hypocenter depth by iterative reduction of the circle overshoot (Author: S. Wendt; see IS 1.1 and EX 11.1).

| Topic       | Acronyms                                                                                                                                                                                 |
|-------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Compiled by | <b>Peter Bormann</b> (formerly GFZ German Research Centre for Geosciences, Potsdam, Telegrafenberg, D-14473 Potsdam, Germany);<br>E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a> |
| Version 1   | November 2011; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_Acronyms">10.2312/GFZ.NMSOP-2_Acronyms</a>                                                                              |

The list below has been compiled by the Editor, with contributions made by several NMSOP authors. It is open for upgrading and corrections. Additional NMSOP-2 related entries or necessary corrections are welcome and should be sent by E-mail to the Editor.

|            |                                                                                                                                                                                                                        |
|------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| AAL        | average annualized loss                                                                                                                                                                                                |
| AC         | alternating current or auto correlation                                                                                                                                                                                |
| ACF        | Auto Correlation Function                                                                                                                                                                                              |
| ACH method | Aki-Christofferson-Husebye method                                                                                                                                                                                      |
| ADC        | analog-to-digital converter; a device that converts data from analog to digital form (see Chapter 6)                                                                                                                   |
| AFTAC      | United States Air Force Technical Applications Center<br>( <a href="http://www.aftac.gov/">http://www.aftac.gov/</a> )                                                                                                 |
| AGU        | American Geophysical Union ( <a href="http://www.agu.org/">http://www.agu.org/</a> ).                                                                                                                                  |
| AH         | ad hoc format                                                                                                                                                                                                          |
| AIC        | Akaike Information Criterion (see Chapter 16)                                                                                                                                                                          |
| ANSS       | Advanced National Seismic System (USA)<br>( <a href="http://www.anss.org/">http://www.anss.org/</a> )                                                                                                                  |
| APA        | average peak amplitude                                                                                                                                                                                                 |
| AR         | auto regression (coefficient; model; power; process)(see Chapter 16)                                                                                                                                                   |
| ARAME      | automatic radionuclide air monitoring equipment (see Chapter 17)                                                                                                                                                       |
| ARCES      | code for a small aperture array in northern Norway, originally:<br>ARCESS – Arctic Experimental Seismic<br>System( <a href="http://www.norsardata.no/NDC/stations/ARC">http://www.norsardata.no/NDC/stations/ARC</a> ) |
| ARF        | array response function (Chapters 9 and 14)                                                                                                                                                                            |
| ASC        | Asian Seismological Commission ( <a href="http://www.asc1996.com">http://www.asc1996.com</a> ).                                                                                                                        |
| ASL        | Albuquerque Seismological Laboratory<br>( <a href="http://earthquake.usgs.gov/regional/asl/">http://earthquake.usgs.gov/regional/asl/</a> )                                                                            |
| ASNR       | amplitude-based signal-to-noise ratio                                                                                                                                                                                  |
| ASP        | analog signal preparation                                                                                                                                                                                              |
| ASRO       | Abbreviated Seismic Research Observatories (seismograph network,<br>also known as the Modified High Gain Long Period Seismograph<br>Network                                                                            |
| ASTM       | American Society for Testing and Materials ( <a href="http://www.astm.org/">http://www.astm.org/</a> )                                                                                                                 |
| AutoDRM    | Automatic Data Request Manager ( <a href="http://seismo.ethz.ch/autodrm/">http://seismo.ethz.ch/autodrm/</a> )                                                                                                         |
| AZI        | azimuth                                                                                                                                                                                                                |
| AZM        | azimuth                                                                                                                                                                                                                |
| BAZ        | backazimuth                                                                                                                                                                                                            |
| BB         | broadband                                                                                                                                                                                                              |
| BBN        | Bayesian belief networks                                                                                                                                                                                               |
| BCIS       | Bureau Central International de Séismologie                                                                                                                                                                            |
| BDSN       | local format in standard analysis software; in use at individual stations<br>and networks (Chapter 10)                                                                                                                 |
| BER        | bit error rate                                                                                                                                                                                                         |

## Acronyms

|         |                                                                                                                                                                                                                                                   |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| BFM     | beam-forming method (Chapters 9 and 14)                                                                                                                                                                                                           |
| BGR     | Federal Institute for Geosciences and Natural Resources (Hannover, Germany; <a href="http://www.bgr.bund.de/">http://www.bgr.bund.de/</a> ; for seismology topics see <a href="http://www.seismologie.bgr.de">http://www.seismologie.bgr.de</a> ) |
| BGS     | British Geological Survey (see <a href="http://www.bgs.ac.uk">http://www.bgs.ac.uk</a> and the BGS seismology website <a href="http://www.earthquakes.bgs.ac.uk">http://www.earthquakes.bgs.ac.uk</a> )                                           |
| BFM     | beam-forming method                                                                                                                                                                                                                               |
| BP      | band pass                                                                                                                                                                                                                                         |
| BSSA    | Bulletin of the Seismological Society of America ( <a href="http://www.seismosoc.org/publications/bssa.html">http://www.seismosoc.org/publications/bssa.html</a> )                                                                                |
| CALTECH | California Institute of Technology, Pasadena, California ( <a href="http://www.caltech.edu">http://www.caltech.edu</a> )                                                                                                                          |
| CANDIS  | Canadian Digital Seismograph Network                                                                                                                                                                                                              |
| CAP     | software developed within the framework of the SESAME project                                                                                                                                                                                     |
| CAV     | cumulative absolute velocity                                                                                                                                                                                                                      |
| CC      | cross-correlation coefficient (see Chapter 9)                                                                                                                                                                                                     |
| CERESIS | Centro Regional de Sismología para América del Sur (Lima, Peru)                                                                                                                                                                                   |
| CD      | compact disk or continuous data                                                                                                                                                                                                                   |
| CD-ROM  | compact disk-read only memory                                                                                                                                                                                                                     |
| CDSN    | China Digital Seismograph Network                                                                                                                                                                                                                 |
| CEB     | calibration event bulletin (of the PIDC)                                                                                                                                                                                                          |
| CF      | characteristic function (non-linear transformation of the seismogram, used in automatic picking; Chapter 16)                                                                                                                                      |
| CGPS    | continuously operating GPS receivers and networks                                                                                                                                                                                                 |
| CSEM    | Centre Sismologique Euro-Méditerranéen (see EMSC)                                                                                                                                                                                                 |
| CLVD    | compensated linear vector dipole                                                                                                                                                                                                                  |
| CMB     | core-mantle boundary                                                                                                                                                                                                                              |
| CMR     | former Center for Monitoring Research (USA)                                                                                                                                                                                                       |
| CMT     | centroid moment tensor                                                                                                                                                                                                                            |
| COSMOS  | Consortium of Organizations for Strong Motion Observation Systems ( <a href="http://www.cosmos-eq.org/">http://www.cosmos-eq.org/</a> )                                                                                                           |
| CoP     | former IASPEI Commission on Practice                                                                                                                                                                                                              |
| CoSOI   | current IASPEI Commission on Seismological Observation and Interpretation ( <a href="http://www.iaspei.org/commissions/CSOI.html">http://www.iaspei.org/commissions/CSOI.html</a> )                                                               |
| CPU     | Central processing unit (in a computer)                                                                                                                                                                                                           |
| CRC     | cyclic redundancy check                                                                                                                                                                                                                           |
| CSM     | cross spectral matrix (see Chapter 9)                                                                                                                                                                                                             |
| CSS     | The former Center for Seismic Studies (USA)                                                                                                                                                                                                       |
| CSS 3.0 | tabulated CSS waveform and database format                                                                                                                                                                                                        |
| CTBT    | Comprehensive Nuclear-Test-Ban Treaty                                                                                                                                                                                                             |
| CTBTO   | Comprehensive Nuclear-Test-Ban Treaty Organization (see PTS) with headquarters in Vienna, Austria ( <a href="http://www.ctbto.org/">http://www.ctbto.org/</a> )                                                                                   |
| CUBE    | An acronym for the Caltech-US Geological Survey Broadcast of Earthquakes, a program to develop and distribute real-time earthquake information in southern California                                                                             |
| CVFK    | conventional semblance-based frequency-wavenumber method after Kvaerna and Ringdahl (1986)                                                                                                                                                        |
| DAC     | digital-to-analog converter; a device which takes a digital value and outputs a voltage which is proportional to the input value                                                                                                                  |
| dB      | acronym for <i>decibel</i> , in acoustics a calibrated measure of signal strength.                                                                                                                                                                |

|         |                                                                                                                                                                                                                                                                     |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| DBMS    | database management system                                                                                                                                                                                                                                          |
| DC      | direct current                                                                                                                                                                                                                                                      |
| DCF     | long-period time signal, which can be received in large parts of Europe                                                                                                                                                                                             |
| DFX     | detection and feature extraction (see Chapter 17)                                                                                                                                                                                                                   |
| DDL     | data description language                                                                                                                                                                                                                                           |
| DLESE   | digital libraries for Earth science education (USA)                                                                                                                                                                                                                 |
| D-GPS   | differential GPS interferometry                                                                                                                                                                                                                                     |
| D-INSAR | differential INSAR                                                                                                                                                                                                                                                  |
| DMC     | data management center                                                                                                                                                                                                                                              |
| DOY     | day of the year                                                                                                                                                                                                                                                     |
| DP      | detection processing                                                                                                                                                                                                                                                |
| DQC     | data quality control (see Chapter 17)                                                                                                                                                                                                                               |
| DRM     | data request manager                                                                                                                                                                                                                                                |
| DS      | acronym for datasheets in NMSOP                                                                                                                                                                                                                                     |
| DSHA    | deterministic seismic hazard assessment                                                                                                                                                                                                                             |
| DSP     | digital signal processor (see Chapter 6)                                                                                                                                                                                                                            |
| DTED    | digital terrain elevation data                                                                                                                                                                                                                                      |
| DTM     | digital terrain model                                                                                                                                                                                                                                               |
| DTRA    | Defense Threat Reduction Agency ( <a href="http://www.dtra.mil/Home.aspx">http://www.dtra.mil/Home.aspx</a> )                                                                                                                                                       |
| DVD     | digital versatile disc                                                                                                                                                                                                                                              |
| DWWSSN  | Digital World-Wide Standard Seismograph Network                                                                                                                                                                                                                     |
| EAEE    | European Association for Earthquake Engineering                                                                                                                                                                                                                     |
| ECOSOC  | Economic and Social Council of the United Nations<br>( <a href="http://www.un.org/esa/coordination/ecosoc/">http://www.un.org/esa/coordination/ecosoc/</a> )                                                                                                        |
| EDM     | electronic distance meter                                                                                                                                                                                                                                           |
| EEFIT   | earthquake engineering field investigation team<br>( <a href="http://www.cen.bris.ac.uk/civil/research/eerc/links/eefit.htm">http://www.cen.bris.ac.uk/civil/research/eerc/links/eefit.htm</a> )                                                                    |
| EERI    | Earthquake Engineering Research Institute ( <a href="http://www.eeri.org">http://www.eeri.org</a> )                                                                                                                                                                 |
| EEWS    | Earthquake early warning system                                                                                                                                                                                                                                     |
| EDUSEIS | EDUcational SEISmological European Network<br>( <a href="http://www.eduseis.com">http://www.eduseis.com</a> )                                                                                                                                                       |
| EHV     | spectral ratio of horizontal-to-vertical component earthquake records                                                                                                                                                                                               |
| EKA     | station code for the UKAEA array in Eskdalemuir in Scotland                                                                                                                                                                                                         |
| EMI     | electromagnetic interference                                                                                                                                                                                                                                        |
| EMS     | electromagnetic seismograph                                                                                                                                                                                                                                         |
| EMS-98  | European Macroseismic Scale 1998( <a href="http://www.gfz-potsdam.de/portal/gfz/Struktur/Departments/Department+2/sec26/resources/Images/EMS-98-eng">http://www.gfz-potsdam.de/portal/gfz/Struktur/Departments/Department+2/sec26/resources/Images/EMS-98-eng</a> ) |
| EMSC    | European-Mediterranean Seismological Centre<br>( <a href="http://www.emsc-csem.org/">http://www.emsc-csem.org/</a> )                                                                                                                                                |
| EOS     | Earth Observing System (centerpiece of ESE)<br>( <a href="http://eospsso.gsfc.nasa.gov/">http://eospsso.gsfc.nasa.gov/</a> )                                                                                                                                        |
| EP      | event processing                                                                                                                                                                                                                                                    |
| EQ      | earthquake                                                                                                                                                                                                                                                          |
| ESAC    | extended spatial auto-correlation (see Chapter 14)                                                                                                                                                                                                                  |
| ESC     | European Seismological Commission of IASPEI ( <a href="http://www.esc-web.org/">http://www.esc-web.org/</a> )                                                                                                                                                       |
| ESE     | Earth's Science Enterprise (of NASA) ( <a href="http://www.earth.nasa.gov/">http://www.earth.nasa.gov/</a> )                                                                                                                                                        |
| ESPAC   | extended spatial auto-correlation method (see Ohori et al., 2002)                                                                                                                                                                                                   |
| ESSTF   | European Standard Seismic Tape Format                                                                                                                                                                                                                               |
| ETAS    | epidemic-type of aftershock sequences                                                                                                                                                                                                                               |

## Acronyms

|          |                                                                                                                                                                                                 |
|----------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| ETS      | episodic tremor and slip                                                                                                                                                                        |
| EX       | acronym for exercises in NMSOP                                                                                                                                                                  |
| FARM     | technique for data exchange                                                                                                                                                                     |
| FBA      | force-balance accelerometer                                                                                                                                                                     |
| FDSN     | Federation of Digital Broad-Band Seismograph Networks<br>( <a href="http://www.fdsn.org/">http://www.fdsn.org/</a> )                                                                            |
| FEC      | forward error-correction                                                                                                                                                                        |
| FEIS     | fast earthquake information system (see Chapter 9)                                                                                                                                              |
| FEMA     | Federal Emergency Management Agency (USA)<br>( <a href="http://www.fema.gov">http://www.fema.gov</a> )                                                                                          |
| FFT      | fast Fourier transformation                                                                                                                                                                     |
| FINES    | code for a small aperture array in southern Finland, originally:<br>FINESA: Finnish Experimental Seismic Array; later: FINESS –<br>Finnish Experimental Seismic System                          |
| FITESC   | field investigation team of the ESC                                                                                                                                                             |
| FIR      | finite impulse response filter, also named acausal- or zero-phase filter                                                                                                                        |
| f-k      | frequency-wavenumber (analysis)                                                                                                                                                                 |
| FM       | frequency modulated                                                                                                                                                                             |
| FOCMEC   | program for the determination of focal mechanisms (Chapter 3)                                                                                                                                   |
| PPFIT    | program for the determination of fault-plane solutions (Chapter 3)                                                                                                                              |
| FRF      | frequency response function                                                                                                                                                                     |
| FSNR     | frequency-based signal-to-noise-ratio                                                                                                                                                           |
| FTAN     | Frequency-time analysis (Chapter 14)                                                                                                                                                            |
| FTP      | fast transfer protocol                                                                                                                                                                          |
| FUNIMAR  | fast invariate case of the minimum Aikake information criterion (see<br>AIC and Chapter 16)                                                                                                     |
| FZ       | fault zone                                                                                                                                                                                      |
| GA       | genetic algorithm                                                                                                                                                                               |
| GBF      | generalized beam forming location algorithm developed at NORSAR<br>(see Chapter 9)                                                                                                              |
| GCF      | Guralp compressed format                                                                                                                                                                        |
| GCI      | Global Communication Infrastructure (see Chapter 17))                                                                                                                                           |
| GDSN     | Global Digital Seismographic Network (see GSN)                                                                                                                                                  |
| GEM      | Global Earthquake Model ( <a href="http://www.globalquakemodel.org/">http://www.globalquakemodel.org/</a> )                                                                                     |
| GEOFON   | GeoForschungsNetz (of broadband seismographs; run by the GFZ;<br><a href="http://geofon.gfz-potsdam.de">http://geofon.gfz-potsdam.de</a> )                                                      |
| GEOSCOPE | French program and station network for global seismological<br>investigations ( <a href="http://geoscope.ipgp.fr/">http://geoscope.ipgp.fr/</a> )                                               |
| GERES    | Code for a small aperture array in southern Germany, originally:<br>GERESS - German Experimental Seismic System (for weblink see<br>GRSN)                                                       |
| GFZ      | GeoForschungsZentrum Potsdam (Germany)<br>( <a href="http://www.gfz-potsdam.de/">http://www.gfz-potsdam.de/</a> )                                                                               |
| GIS      | geographical information system                                                                                                                                                                 |
| GIT      | generalized inversion technique (Chapter 14)                                                                                                                                                    |
| GLONAS   | Russian global (satellite-based) navigation system (as GPS)                                                                                                                                     |
| GPS      | US (satellite-based) global positioning system<br>( <a href="http://www.colorado.edu/geography/gcraft/notes/gps/gps_f.html">http://www.colorado.edu/geography/gcraft/notes/gps/gps_f.html</a> ) |
| G-R      | Gutenberg-Richter (magnitude-frequency relationship)                                                                                                                                            |
| GRF      | station code for the broadband array near Gräfenberg, Germany;<br>see weblink GRSN with array outlay and <a href="http://www.szgrf.bgr.de/">http://www.szgrf.bgr.de/</a>                        |

|              |                                                                                                                                                                                                                                                                                                                                                              |
|--------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| GRFO         | Former Gräfenberg Observatory, also code for the SRO station co-located with site GRA1 of the GRF array                                                                                                                                                                                                                                                      |
| GRSN         | German Regional Seismic Network<br>( <a href="http://www.bgr.bund.de/DE/Themen/Erdbeben-Gefaehrdungsanalysen/Seismologie/Seismologie/Seismometer_Stationen/Stationsnetze/d_stationsnetz_node.html">http://www.bgr.bund.de/DE/Themen/Erdbeben-Gefaehrdungsanalysen/Seismologie/Seismologie/Seismometer_Stationen/Stationsnetze/d_stationsnetz_node.html</a> ) |
| GSE          | Group of Scientific Experts of the Conference on Disarmament in Geneva (meetings 1976 – 1996)                                                                                                                                                                                                                                                                |
| GSE (format) | formats designed for seismic data exchange and archiving by the GSE (e.g., GSE1.0, GSE2.1, GSE2.X)                                                                                                                                                                                                                                                           |
| GSETT        | Group of Scientific Experts Technical Test                                                                                                                                                                                                                                                                                                                   |
| GSETT-3      | Group of Scientific Experts Third Technical Test January 1995 – March 2000                                                                                                                                                                                                                                                                                   |
| GSHAP        | Global Seismic Hazard Assessment Programme                                                                                                                                                                                                                                                                                                                   |
| GSN          | Global Seismographic Network (of IRIS and the FDSN)<br>( <a href="http://www.iris.edu/hq/programs/gsn">http://www.iris.edu/hq/programs/gsn</a> )                                                                                                                                                                                                             |
| GT           | ground truth                                                                                                                                                                                                                                                                                                                                                 |
| GUI          | graphical user interface                                                                                                                                                                                                                                                                                                                                     |
| HF           | high-frequency                                                                                                                                                                                                                                                                                                                                               |
| HGLP         | high-gain long period system                                                                                                                                                                                                                                                                                                                                 |
| HHM          | hidden Markov model (applied in speech recognition techniques)                                                                                                                                                                                                                                                                                               |
| HP           | high pass                                                                                                                                                                                                                                                                                                                                                    |
| HRFK         | high resolution frequency-wavenumber (K) method (see Chapter 9)                                                                                                                                                                                                                                                                                              |
| HRVD         | Harvard University, USA ( <a href="http://www.harvard.edu/">http://www.harvard.edu/</a> )                                                                                                                                                                                                                                                                    |
| H/V          | amplitude ratio between horizontal and vertical component observations of seismic noise, used to investigate site effects                                                                                                                                                                                                                                    |
| IAEE         | International Association for Earthquake Engineering<br>( <a href="http://www.iaee.or.jp/">http://www.iaee.or.jp/</a> )                                                                                                                                                                                                                                      |
| IAGA         | International Association of Geomagnetism and Aeronomy<br>( <a href="http://www.iugg.org/">http://www.iugg.org/</a> )                                                                                                                                                                                                                                        |
| IASP91       | IASPEI seismic travel-time model (see DS 2.1)                                                                                                                                                                                                                                                                                                                |
| IASPEI       | International Association of Seismology and Physics of the Earth's Interior (of the IUGG) ( <a href="http://www.iaspei.org/">http://www.iaspei.org/</a> )                                                                                                                                                                                                    |
| IDP          | intensity data point (see Chapter 12)                                                                                                                                                                                                                                                                                                                        |
| ICB          | inner core boundary                                                                                                                                                                                                                                                                                                                                          |
| IDA          | International Deployment of Accelerometers<br>( <a href="http://ida.ucsd.edu/index.html">http://ida.ucsd.edu/index.html</a> )                                                                                                                                                                                                                                |
| IDC          | international data center (specifically that of the CTBTO)                                                                                                                                                                                                                                                                                                   |
| IDE          | integrated drive electronics                                                                                                                                                                                                                                                                                                                                 |
| IDNDR        | United Nations International Decade for Natural Disaster Reduction (by United Nations General Assembly resolution 44/236 of December 22, 1989, designated decade of the 1990s)                                                                                                                                                                               |
| IERRS        | Istanbul Earthquake Rapid Response System                                                                                                                                                                                                                                                                                                                    |
| IGS          | international GPS Service; an organization responsible for worldwide coordination of continuous GPS measurements                                                                                                                                                                                                                                             |
| IIR          | infinite impulse response filter, also named causal filter; all analog filters are of this type                                                                                                                                                                                                                                                              |
| IISEE        | International Institute of Seismology and Earthquake Engineering, Tsukuba, Japan ( <a href="http://iisee.kenken.go.jp/">http://iisee.kenken.go.jp/</a> )                                                                                                                                                                                                     |
| IMS          | International Monitoring System (of PTS/CTBTO); (see Chapter 17 and <a href="http://www.ieer.org/sdfiles/vol_8/8-2/ver-ctbt.html">http://www.ieer.org/sdfiles/vol_8/8-2/ver-ctbt.html</a> )                                                                                                                                                                  |



## Acronyms

|        |                                                                                                                                                                                            |
|--------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| IMS1.0 | format for the exchange of parameter data adopted by the IMS                                                                                                                               |
| INGV   | Istituto Nazionale di Geofisica e Vulcanologia, Rome, Italy<br>( <a href="http://www.ingv.it/">http://www.ingv.it/</a> )                                                                   |
| INSAR  | interferometric (analysis of) synthetic aperture radar (images)                                                                                                                            |
| IP     | Internet protocol; combination of numbers which is associated with a computer in the Internet, for explicit identification of a computer                                                   |
| IPGP   | Institut de Physique du Globe de Paris<br>( <a href="http://www.ipgp.fr/index2.php?Largeur=1280">http://www.ipgp.fr/index2.php?Largeur=1280</a> )                                          |
| IRIS   | Incorporated Research Institutions for Seismology (USA)<br>( <a href="http://www.iris.edu/hq/">http://www.iris.edu/hq/</a> )                                                               |
| ISAM   | indexed sequential access method                                                                                                                                                           |
| ISC    | International Seismological Centre (Thatcham, UK)<br>( <a href="http://www.isc.ac.uk/">http://www.isc.ac.uk/</a> )                                                                         |
| ISN    | independent sub-network of the IMS (see Chapter 17)                                                                                                                                        |
| ISS    | International Seismological Summary (see Chapters 4, 41, and 88 in Lee et al., 2002).                                                                                                      |
| ISDN   | Integrated Services Digital Network                                                                                                                                                        |
| ISF    | IASPEI Seismic Format (Chapter 10);<br><a href="http://www.isc.ac.uk/standards/isf">http://www.isc.ac.uk/standards/isf</a>                                                                 |
| ISOP   | International Seismic Observing Period (important international project proposed by IASPEI in the 1990s; only partially realized)                                                          |
| ITRF   | International Terrestrial Reference Frame (defined by a combination of VLBI, SLR, DORIS and GPS currently used to represent absolute coordinates for sub-centimeter geodetic measurements) |
| IUGG   | International Union of Geodesy and Geophysics<br>( <a href="http://www.iugg.org/">http://www.iugg.org/</a> )                                                                               |
| JB     | Joyner-Boore, JB distance is the distance to the nearest point on the surface projection of a fault rupture                                                                                |
| J-B    | Jeffrey-Bullen (travel-time curves or Earth model)                                                                                                                                         |
| JMA    | Japanese Meteorological Agency<br>( <a href="http://www.jma.go.jp/jma/indexe.html">http://www.jma.go.jp/jma/indexe.html</a> )                                                              |
| JPL    | Jet Propulsion Laboratory at Pasadena, California.                                                                                                                                         |
| KAPG   | former Commission of the Academies of Sciences of Socialistic Countries for Planetary Geophysical Research                                                                                 |
| KNMI   | The Royal Netherlands Meteorological Institute (for Research Seismology Division see: <a href="http://www.knmi.nl/research/seismology/">http://www.knmi.nl/research/seismology/</a> )      |
| LAB    | lithosphere-asthenosphere boundary                                                                                                                                                         |
| LASA   | Large Aperture Seismic Array (Montana, USA, in operation from 1965 to 1978)                                                                                                                |
| LF     | low frequency                                                                                                                                                                              |
| LFE    | low-frequency earthquakes                                                                                                                                                                  |
| LDEO   | Lamont-Doherty Earth Observatory, Palisades, N.Y., USA, formerly LDGO (see <a href="http://www.ldeo.columbia.edu">http://www.ldeo.columbia.edu</a> ),                                      |
| LDGO   | Lamont-Doherty Geological Observatory, Palisades, N.Y. USA                                                                                                                                 |
| LP     | stands for either low pass (filter) or long-period (seismographs)                                                                                                                          |
| LTA    | long-term average (of noise and/or signal amplitudes) (Chapter 9 and IS 8.1)                                                                                                               |
| LTI    | Linear time-invariant                                                                                                                                                                      |
| LVZ    | low-velocity zone                                                                                                                                                                          |
| Ma     | Mega annum; an abbreviation for million years ago                                                                                                                                          |

|                |                                                                                                                                                                         |
|----------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| $M_0$          | scalar seismic moment                                                                                                                                                   |
| mb             | seismic body-wave magnitude; determined from short-period P wave amplitude and period measurements                                                                      |
| mB             | seismic body-wave magnitude; determined from amplitude and period measurements on medium-period or broadband records of P, PP and S waves                               |
| MCS            | Mercalli-Cancani-Sieberg seismic intensity scale.                                                                                                                       |
| MCSP           | multi-channel seismic profiling                                                                                                                                         |
| MDA            | multiple discriminant analysis (Chapter 16)                                                                                                                             |
| MDP            | macroseismic data point                                                                                                                                                 |
| Me             | energy magnitude                                                                                                                                                        |
| MEDNET         | MEDiterranean NETwork ( <a href="http://www.mednet.ingrm.it/">http://www.mednet.ingrm.it/</a> )                                                                         |
| MEMS           | micro-electromechanical systems                                                                                                                                         |
| miniSEED       | SEED format without any of the associated control header information                                                                                                    |
| $M_l$ or $M_L$ | local magnitude (according to the original definition by Richter (1935))                                                                                                |
| MLM            | maximum-likelihood method                                                                                                                                               |
| $M_m$          | magnitude derived from observation of mantle surface waves                                                                                                              |
| MMI            | modified Mercalli intensity                                                                                                                                             |
| MM56           | Modified Mercalli Scale of 1956                                                                                                                                         |
| $M_{max}$      | maximum magnitude                                                                                                                                                       |
| $M_{ms}$       | magnitude derived from macroseismic intensity observation                                                                                                               |
| MOHO (Moho)    | abbreviation for the Mohorovičić-discontinuity (boundary between the Earth's crust and mantle)                                                                          |
| MP events      | multi-phase events (observed at volcanoes)                                                                                                                              |
| MPX            | program for automated phase picking                                                                                                                                     |
| $M_s$          | surface-wave magnitude                                                                                                                                                  |
| MSB            | most significant bit                                                                                                                                                    |
| MSEED          | mini-SEED (data format)                                                                                                                                                 |
| MSK            | macroseismic intensity scale according to Medvedev, Sponheuer and Karnik                                                                                                |
| MSOP           | Manual of Seismological Observatory Practice (1979 edition) (see <a href="http://www.iaspei.org/projects/NMSOP.html">http://www.iaspei.org/projects/NMSOP.html</a> )    |
| MSPAC          | modified SPAC method (see SPAC and Bettig et al., 2001)                                                                                                                 |
| $M_t$          | tsunami magnitude                                                                                                                                                       |
| MUSIC          | multiple signal classification (see Chapter 9)                                                                                                                          |
| $M_w$          | seismic moment magnitude according to Kanamori (1977)                                                                                                                   |
| NASA           | National Aeronautics and Space Administration (USA) ( <a href="http://www.nasa.gov/">http://www.nasa.gov/</a> )                                                         |
| NCSN           | Northern California Seismic Network (USA) ( <a href="http://quake.geo.berkeley.edu/ncsn/ncsn.overview.html">http://quake.geo.berkeley.edu/ncsn/ncsn.overview.html</a> ) |
| NDC            | national data centers (in the framework of the PTS/CTBTO)                                                                                                               |
| NDPC           | NORSAR data processing center                                                                                                                                           |
| NEHRP          | National Earthquake Hazards Reduction Program of the United States                                                                                                      |
| NEIC           | National Earthquake Information Center of the USGS; acts as WDC for earthquake data ( <a href="http://neic.usgs.gov/">http://neic.usgs.gov/</a> )                       |
| NEIS           | U.S. Geological Survey National Earthquake Information Service (no longer used abbreviation)                                                                            |
| NERC           | Natural Environment Research Council (USA) ( <a href="http://www.nerc.ac.uk/">http://www.nerc.ac.uk/</a> )                                                              |
| NERIES         | Network of Research Infrastructures for European Seismology                                                                                                             |

## Acronyms

|         |                                                                                                                                                                                                                                  |
|---------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| NEWS    | NORSAR's earthquake warning system (see Chapter 9)                                                                                                                                                                               |
| NetDC   | Networked Data Center Protocol (see Chapter 10)<br>( <a href="http://www.iris.washington.edu/manuals/netdc/netdcuser.htm">http://www.iris.washington.edu/manuals/netdc/netdcuser.htm</a> )                                       |
| NGA     | next generation attenuation                                                                                                                                                                                                      |
| NGNM    | New Global Noise Model (see section 4.1.2)                                                                                                                                                                                       |
| NHNM    | New High-Noise Model (see section 4.1.2)                                                                                                                                                                                         |
| NHV     | seismic noise spectral ratio between horizontal-to-vertical component (Chapter 14)                                                                                                                                               |
| NMSOP   | New Manual of Seismological Observatory Practice (2002, 2009 and 2011-12 editions; see <a href="http://nmsop.gfz-potsdam.de">http://nmsop.gfz-potsdam.de</a> )                                                                   |
| NNSN    | Norwegian National Seismic Network<br>( <a href="http://www.geo.uib.no/seismo/naar-jorden-skjelver/index.php?topic=nnsn&amp;lang=es">http://www.geo.uib.no/seismo/naar-jorden-skjelver/index.php?topic=nnsn&amp;lang=es</a> )    |
| NOA     | since 1978 station code for the large aperture NORSAR array<br>( <a href="http://www.norsardata.no/NDC/stations/NOA/">http://www.norsardata.no/NDC/stations/NOA/</a> )                                                           |
| NOAA    | National Oceanic and Atmospheric Administration (USA)<br>( <a href="http://www.noaa.gov/">http://www.noaa.gov/</a> )                                                                                                             |
| NORES   | station code of a small aperture array in southern Norway, originally NORESS – Norwegian Experimental Seismic Station<br>( <a href="http://www.norsardata.no/NDC/stations/NRS/">http://www.norsardata.no/NDC/stations/NRS/</a> ) |
| NORSAR  | Norwegian seismological institute, originally used for the large aperture seismic array in southern Norway (from NORwegian Seismic ARray) ( <a href="http://www.norsar.no/">http://www.norsar.no/</a> )                          |
| NPEF    | noise prediction-error filter                                                                                                                                                                                                    |
| NSF     | National Science Foundation (USA)                                                                                                                                                                                                |
| NTS     | Nevada Test Site ( <a href="http://www.nv.doe.gov/main.aspx">http://www.nv.doe.gov/main.aspx</a> )                                                                                                                               |
| OBS     | ocean-bottom seismograph (see sub-chapter 7.5 of Chapter 7)                                                                                                                                                                      |
| OBH     | ocean-bottom hydrophone                                                                                                                                                                                                          |
| ODC     | ORFEUS Data Center (De Bilt, Netherlands)                                                                                                                                                                                        |
| ODP     | Ocean Drilling Program (1983-2003)                                                                                                                                                                                               |
| ORFEUS  | Observatories and Research Facilities for European Seismology (De Bilt, Netherlands; <a href="http://www.orfeus-eu.org/">http://www.orfeus-eu.org/</a> )                                                                         |
| OSI     | on-site inspection capabilities (of the PTS/CTBTO)                                                                                                                                                                               |
| OT      | origin time                                                                                                                                                                                                                      |
| PACF    | partial auto-correlation function                                                                                                                                                                                                |
| PASSCAL | Program for Array Seismic Studies of the Continental Lithosphere (see: <a href="http://www.passcal.nmt.edu/content/usarray">http://www.passcal.nmt.edu/content/usarray</a> )                                                     |
| PCEQ    | pulse-coded earthquake data format (Chapter 10)                                                                                                                                                                                  |
| PD      | program descriptions in NMSOP                                                                                                                                                                                                    |
| PDE     | preliminary determination of epicenters (NEIC event data reports)                                                                                                                                                                |
| PDF     | probability density function                                                                                                                                                                                                     |
| PDR-2   | local format in use at individual stations and networks                                                                                                                                                                          |
| PGA     | peak ground acceleration                                                                                                                                                                                                         |
| PGD     | peak ground displacement                                                                                                                                                                                                         |
| PGV     | peak ground velocity                                                                                                                                                                                                             |
| PIDC    | the former Prototype International Data Center (USA); many PIDC products are still available at the DTRA Verification Data Base ( <a href="http://www.rdss.info/">http://www.rdss.info/</a> )                                    |
| PITSA   | programmable interactive toolbox for seismological analysis                                                                                                                                                                      |
| PMCC    | progressive multichannel cross-correlation method                                                                                                                                                                                |
| PML     | probable maximum loss                                                                                                                                                                                                            |

|            |                                                                                                                                                                                                                                        |
|------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| POSEIDON   | Pacific Orient Seismic Digital Observation Network (Japanese seismic network for global seismological studies which includes OBS sites)                                                                                                |
| PREM       | Preliminary Reference Earth Model (see DS 2.1)                                                                                                                                                                                         |
| PREPROC    | program for preprocessing of digital seismic data                                                                                                                                                                                      |
| PSA        | pseudo-spectral acceleration                                                                                                                                                                                                           |
| PSD        | power spectral density                                                                                                                                                                                                                 |
| PSHA       | probabilistic seismic hazard assessment                                                                                                                                                                                                |
| PSRV       | pseudo relative velocity response spectrum                                                                                                                                                                                             |
| PTS        | Preparatory Technical Secretary of the Preparation Commission of the CTBTO in Vienna, Austria; organization running the IMS and IDC before the CTBT is coming into force                                                               |
| PVC        | Poly-Vinyl-Chloride                                                                                                                                                                                                                    |
| QED        | quick epicenter determinations (NEIC event data reports)                                                                                                                                                                               |
| Q          | quality factor; Q is inverse proportional to the attenuation of seismic waves, i.e., to the relative loss of energy per wave cycle                                                                                                     |
| RASA       | radionuclide aerosol sampler/analyzer (see Chapter 17)                                                                                                                                                                                 |
| RBW        | relative bandwidth                                                                                                                                                                                                                     |
| RC         | resistivity-capacity (electric circuit, filter, element etc.)                                                                                                                                                                          |
| RDBMS      | relational data base management system (see Chapter 17)                                                                                                                                                                                |
| rdseed     | SEED-reading program                                                                                                                                                                                                                   |
| RDSS       | former Research and Development Support System (at the CMR, USA), now integrated in the DTRA Verification Data Base ( <a href="http://www.rdss.info/">http://www.rdss.info/</a> )                                                      |
| REB        | reviewed event bulletin                                                                                                                                                                                                                |
| REDB       | reference event data base (of the IDC)                                                                                                                                                                                                 |
| Reftek     | Refraction Technology Inc.<br>( <a href="http://www.reftek.com/products/accelerometers-147-01.html">http://www.reftek.com/products/accelerometers-147-01.html</a> )                                                                    |
| RF         | radio frequency or Rossi-Forel intensity scale                                                                                                                                                                                         |
| RMS or rms | root mean square                                                                                                                                                                                                                       |
| ROSINE     | resolution of site response issues in the Northridge earthquake                                                                                                                                                                        |
| ROV        | remotely operated vehicle (see <i>Glossary</i> for reference)                                                                                                                                                                          |
| RSAM       | real-time seismic amplitude measurement                                                                                                                                                                                                |
| RSM        | reference site method (Chapter 14)                                                                                                                                                                                                     |
| RSTN       | remote seismic telemetered network                                                                                                                                                                                                     |
| SA         | sSpectral acceleration                                                                                                                                                                                                                 |
| SAC        | seismic analysis code (format used in standard analysis software)                                                                                                                                                                      |
| SAP        | signal attribute processing (s. Chapter 9, section 9.13.3)                                                                                                                                                                             |
| SAR        | successive approximation register<br><b>Also:</b> synthetic aperture radar (see <i>Glossary</i> for reference)                                                                                                                         |
| SAR-ADC    | Successive approximation analog-to-digital converter                                                                                                                                                                                   |
| SASCs      | slowness and azimuth station corrections (see Chapter 17)                                                                                                                                                                              |
| SCEC       | Southern California Earthquake Center<br>( <a href="http://www.scec.org/">http://www.scec.org/</a> )                                                                                                                                   |
| SCRs       | stable continental regions                                                                                                                                                                                                             |
| SCSI       | small computer system interface                                                                                                                                                                                                        |
| SCSN       | Southern California Seismic Network<br>( <a href="http://www.trinet.org/scsn/scsn.html">http://www.trinet.org/scsn/scsn.html</a> )                                                                                                     |
| SEB        | standard event bulletin (of the PIDC)                                                                                                                                                                                                  |
| SEED       | standard for the exchange of earthquake data (format designed for data exchange and archiving (for download the SEED manual see: <a href="http://www.iris.edu/manuals/SEED_chpt1.htm">http://www.iris.edu/manuals/SEED_chpt1.htm</a> ) |

## Acronyms

|                  |                                                                                                                                                                                                                                                                |
|------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| SEEDlink         | program package to exchange SEED data via Internet                                                                                                                                                                                                             |
| SEG Y            | data format                                                                                                                                                                                                                                                    |
| SEISAN           | Seismogram Analysis Software (manual and software can be downloaded from <a href="http://ftp.geo.uib.no/pub/seismo/SOFTWARE/SEISAN/">http://ftp.geo.uib.no/pub/seismo/SOFTWARE/SEISAN/</a> )                                                                   |
| SEL3             | standard event list 3; the final fully automatic bulletin of the IDC (PTS/CTBTO)                                                                                                                                                                               |
| SESAME           | Site Effects uSing AMBient Excitations; a European Union project (also the name of a project producing a unified hazard model for the Euro-Mediterranean region: <a href="http://www.ija.csic.es/gt/earthquakes/">http://www.ija.csic.es/gt/earthquakes/</a> ) |
| SGM              | strong ground motion                                                                                                                                                                                                                                           |
| SH               | Seismic Handler (data analysis program), command-line version                                                                                                                                                                                                  |
| SHI              | seismic hydroacoustic and infrasonic (data products see IS 10.3)                                                                                                                                                                                               |
| SHM              | Seismic Handler (data analysis program), motif version (see <a href="http://seismic-handler.org/portal">http://seismic-handler.org/portal</a> )                                                                                                                |
| SI               | Système International d'Unités (international system of standard units)                                                                                                                                                                                        |
| SKD              | classical medium- to long-period (in Russian: Dlinnoperiodnij) displacement-proportional seismograph constructed by D. P. Kirnos                                                                                                                               |
| SNR or S/N ratio | signal-to-noise ratio (see sub-chapters 4.4 and 9.6)                                                                                                                                                                                                           |
| SOFAR            | sound fixing and ranging                                                                                                                                                                                                                                       |
| SP               | short-period                                                                                                                                                                                                                                                   |
| SPAC             | spatial auto-correlation method, after Aki (1957) (Chapter 14)                                                                                                                                                                                                 |
| SPITS            | code for a small aperture array on Spitsbergen (Svalbard Archipelago)                                                                                                                                                                                          |
| SPT              | split-barrel sampling; also acronym for the Semipalatinsk test site in Kazakhstan                                                                                                                                                                              |
| SRL              | satellite laser ranging                                                                                                                                                                                                                                        |
| SRO              | Seismic Research Observatory (in a global USA network) using very specific 20 s bandpass and 6 s noise notch filtered LP records. For data and links see NMSOP-2 <i>Glossary</i> , Chapter 11, DS 11.2 and 11.3                                                |
| SS-1/SSR-1       | Kinematics "Ranger" seismometer and data logger, respectively                                                                                                                                                                                                  |
| SSA              | Seismological Society of America ( <a href="http://www.seismosoc.org/">http://www.seismosoc.org/</a> )                                                                                                                                                         |
| SSAM             | spectral seismic amplitude measurement                                                                                                                                                                                                                         |
| SSR              | standard spectral ratio method (Chapter 14)                                                                                                                                                                                                                    |
| SSSC             | source specific station correction (see Chapter 17)                                                                                                                                                                                                            |
| SSZ              | seismic source zone                                                                                                                                                                                                                                            |
| STA              | Short-term average (in a trigger algorithm) (see Chapter 9 and IS 8.1)                                                                                                                                                                                         |
| STA/LTA          | ratio of short-term to long-term average (trigger algorithm) (see IS 8.1 and Chapters 8 and 9)                                                                                                                                                                 |
| STA2             | sStation line in parameter data format (Chapter 10)                                                                                                                                                                                                            |
| StaPro           | station processing (see Chapter 17)                                                                                                                                                                                                                            |
| STEIM1 or 2      | Algorithms proposed by J. M. Steim for data compression                                                                                                                                                                                                        |
| STS1 or STS2     | Streckeisen seismometers, type 1 and 2, resp. (see DS 5.1)                                                                                                                                                                                                     |
| SUDS             | seismic unified data system (format designed for database systems)                                                                                                                                                                                             |
| SZGRF            | Seismological Central Observatory GRF (Gräfenberg, Germany; see <a href="http://www.szgrf.bgr.de/">http://www.szgrf.bgr.de/</a> )                                                                                                                              |
| TCP              | transmission control protocol                                                                                                                                                                                                                                  |
| TCP/IP           | transmission control protocol over Internet protocol                                                                                                                                                                                                           |
| TERRAScope       | very broadband seismographic network in Southern California                                                                                                                                                                                                    |
| TF               | telegraphic format                                                                                                                                                                                                                                             |
| THR              | SNR threshold used to define a detection                                                                                                                                                                                                                       |
| TR               | transient response                                                                                                                                                                                                                                             |

|          |                                                                                                                                |
|----------|--------------------------------------------------------------------------------------------------------------------------------|
| UHF      | ultra high-frequency range (around 450 MHz)                                                                                    |
| UHS      | uniform hazard spectrum                                                                                                        |
| ULF      | ultra low frequency (volcanic events with $f < 0.01$ Hz)                                                                       |
| ULP      | ultra long period                                                                                                              |
| UNE      | underground nuclear explosion                                                                                                  |
| UNESCO   | United Nations Educational, Scientific and Cultural Organization                                                               |
| UPS      | unbreakable power supply                                                                                                       |
| URL      | uniform resource locator; it is the address for locating a Web site, and consists of the protocol and the Internet address.    |
| URS      | uniform risk spectrum (actually the same as UHS)                                                                               |
| USESN    | United States Educational Seismology Network<br>( <a href="http://www.indiana.edu">http://www.indiana.edu</a> )                |
| USCGS    | United States Coast and Geodetic Survey                                                                                        |
| USGS     | United States Geological Survey ( <a href="http://www.usgs.gov/">http://www.usgs.gov/</a> )                                    |
| USNSN    | United States National Seismograph Network                                                                                     |
| UT       | universal time                                                                                                                 |
| UTC      | universal time coordinated; a time scale defined by the International Time Bureau and agreed upon by international convention. |
| VBB      | very broadband                                                                                                                 |
| VEI      | volcanic explosivity index                                                                                                     |
| VELEST   | program to derive 1-D seismic velocity models                                                                                  |
| VESPA    | velocity spectrum analysis                                                                                                     |
| VHF      | very-high frequency range (160-200 MHz)                                                                                        |
| VLBI     | very long baseline interferometry; uses radar frequencies around 9 GHz with wavelengths around 3 cm.                           |
| VLF      | very low frequency (volcanic events with $f \sim 0.01 - 1$ Hz)                                                                 |
| VLP      | very long period; in reference to the 27-second to 600-second band of seismic signals recorded by a modern GSN station.        |
| $v_p$    | P-wave velocity                                                                                                                |
| $v_s$    | S-wave velocity                                                                                                                |
| VSAT     | very small aperture terminals                                                                                                  |
| VT-A     | deep volcanic-tectonic events                                                                                                  |
| VT-B     | shallow volcanic-tectonic events                                                                                               |
| WA       | Wood-Anderson torsion seismometer                                                                                              |
| WAN      | wide area network                                                                                                              |
| WDC      | world data center                                                                                                              |
| WIDC     | waveform identification line                                                                                                   |
| WWSSN    | World-Wide Standard Seismograph Network                                                                                        |
| WWSSN-LP | long-period seismographs of the WWSSN                                                                                          |
| WWSSN-SP | short-period seismographs of the WWSSN                                                                                         |
| YBP      | years before present                                                                                                           |
| YKA      | Yellowknife array (Canada)                                                                                                     |

| Title                   | Glossary of interest to earthquake and engineering seismologists                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|-------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Compiled and amended by | <p><b>Peter Bormann</b>, retired from GFZ German Research Centre for Geosciences, Telegrafenberg, 14473 Potsdam, Germany; E-mail: <a href="mailto:pb65@gmx.net">pb65@gmx.net</a></p> <p><b>Keiiti Aki</b> (†) (1930-2005)</p> <p><b>William H. K. Lee</b>, retired from U.S. Geological Survey, Menlo Park, CA 94025, USA; E-mail: <a href="mailto:willielee88@gmail.com">willielee88@gmail.com</a></p> <p><b>Johannes Schweitzer</b>, NORSAR, Gunnar Randers vei 15, P.O. Box 53, N-2027 Kjeller, Norway; E-mail: <a href="mailto:johannes.schweitzer@norsar.no">johannes.schweitzer@norsar.no</a></p> |
| Version 2               | December 2013; DOI: <a href="https://doi.org/10.2312/GFZ.NMSOP-2_Glossary">10.2312/GFZ.NMSOP-2_Glossary</a>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |

## Acknowledgment

This second version of the NMSOP-2 Glossary is based on a very careful proof-reading, uniform formatting, and the spotting of missing links, references and entries by Ms. Dr. Myrto Pirli whom the authors owe great thanks. This tremendously eased the revision.

## 1 Introduction

**This is a preliminary glossary**, still under development, revision and amendment, also with respect to co-authorship. **Comments, corrections as well as proposals for complementary entries are highly welcome and should be sent to the editor, P. Bormann.** He has compiled the current version by combining, harmonizing, amending and sometimes correcting entries from three major earlier published glossaries of terms that are (see References):

- 1) frequently used in seismological observatory practice (glossary in the first edition of the IASPEI New Manual of Seismological Observatory Practice (NMSOP; Bormann, 2002 and 2009; see <http://nmsop.gfz-potsdam.de>);
- 2) of interest to earthquake and engineering seismologists (glossary compiled by Aki and Lee, 2003, for the International Handbook of Earthquake and Engineering Seismology (see Lee et al., 2003a);
- 3) now widely used in some rapidly developing new fields such as rotational seismology, (compiled by Lee, 2009).

The following glossary includes some 1,500 specialized terms. Not included were abbreviations and acronyms (with the exception of just a few that required more explanation). Acronyms have been added as an extra file to this Manual (see NMSOP-2 website cover page). Alphabetization is strictly letter-by-letter of the glossary terms (in bold face), ignoring hyphenation and space between words. Words included in parenthesis are also ignored in the alphabetization. Words which are written in *italics* in the explanatory text refer to other glossary terms for which more explanation is given elsewhere.

Terms in this glossary may be encountered when reading literature in seismological observatory practice, earthquake and engineering seismology and related fields. The purpose



of the glossary is to provide a short definition and/or a starting point for readers to learn more from cited references. Although rather extensive, this glossary is by no means complete. We attempt to give commonly used definitions, understandable also for non-specialists. But readers should realize that terms may be differently defined in some literature. We avoid giving mathematical formulas (except for a small number of cases). For key formulas in seismology we refer to Yehuda Ben-Zion (2003), as well as to Aki and Richards (2002 and 2009).

When compiling the above mentioned earlier main glossaries many earlier and widely scattered ones were consulted, such as glossaries published by the US Geological Survey (<http://www.usgs.gov/>), the Montana Bureau of Mines and Geology (<http://www.mbmng.mtech.edu/>), the International Data Center of the CTBTO (<http://www.ctbto.org>), and in addition several standard reference sources, including Aki and Richards (1980, 2002 and 2009), Ben-Menahem and Singh (2000), Bormann (2002), Clark (1966), EERI (1989), Hancock and Skinner (2000), Jackson (1997), James and James (1976), Lee and Stewart (1981), Lide (2002), Meyers (2002), Naeim (2001), Richter (1958), Runcorn (1967), Sigurdsson (2000), USGS Photo Glossary (2002), and Yeats *et al.* (1997). Some glossary terms and texts have also been contributed by chapter/section/entry authors of this Manual, amongst them J. Schweitzer, T. Diehl and L. Kueperkoch, as well as by authors of the International Handbook of Earthquake and Engineering Seismology (eds. Lee *et al.*, 2002 and 2003a) and of the BSSA special issue on rotational seismology (eds. Lee *et al.*, 2009, Lee *et al.*, 2009a, and 2009b), amongst them (for the latter) R. DeSalvo, J. R. Evans, E. F. Grekova, C. R. Hutt, H. Igel, C. A. Langston, B. Lantz, E. Majewski, R. Nigbor, J. Pujol, P. Spudich, R. Teisseyre, M. D. Trifunac, R. J. Twiss, and Z. Zembaty. Further terms and/or suggestions for revisions were provided by R. Adams, J. Anderson, J. Andrews, Y. Ben-Zion, M.G. Bonilla, D. Boore, K. Campbell, Ch. Chapman, C.B. Crouse, A. Curtis, A. Douglas, L. Esteva, B. Hall, D. Hill, B. Hutt, P. Irwin, K.-H. Jäckel, R. Jeanloz, P. Jennings, C. Kisslinger, O. Kulhanek, A. McGarr, M. Pirli, C. Prodehl, H. Sato, J. Savage, J. Schweitzer, A. Snoke, K. Suyehiro, B. Tilling, and S. Uyeda. More names may be added here with future entries.

Nevertheless, words of caution are required at this point. Although precise definitions are critical for fostering collaboration and reducing misunderstanding between different disciplines, there are no unique definitions of terms. Every scientific field seems to develop its own “jargon” that is sometimes difficult for outsiders to understand. Even within a given field, different authors often define technical terms differently, or a given term has different usage. This is particularly the case in rapidly developing new fields such as rotational seismology. Therefore, the compilers tried to use a usage that they judged to be “common”, or to indicate different meanings of the same term in different fields of Earth or technical sciences and sometimes by providing references to alternative and more specialized sources. With respect to rotational seismology these are especially the tutorials by Peters (2009), Pujol (2009), Teisseyre (2009), and Zembaty (2009b), and the review by Trifunac (2009). Evans *et al.* (2010) describe notation conventions for rotational seismology, while Grekova and Lee (2009) provide suggested readings in continuum mechanics (including elasticity theories), and earthquake seismology.



## 2 Glossary terms

### A

|                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
|-----------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>aa</b>             | A common type of <i>lava flow</i> composed of a jumble of <i>lava blocks</i> , with a rough, jagged, spiny, and generally clinkery or slaggy surface. Of Hawaiian origin, the term aa is applied also to similar lava flows at other volcanoes.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>acausal filter</b> | See <i>zero-phase filter</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>acausal signal</b> | Output signal that exists also for times prior to the application of an input <i>signal</i> , a non-physical phenomenon of digital data processing.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>acceleration</b>   | When an object, e.g., a car, changes its speed from one speed to another, it is accelerating (moving faster) or decelerating (moving slower). This change in <i>velocity</i> is called acceleration. In seismology this term specifically means the rate of change of <i>ground motion</i> particle velocity per unit time when the ground is shaking due to an <i>earthquake</i> or another kind of <i>seismic source</i> process. The peak acceleration is the largest acceleration recorded by a particular station during an earthquake. In <i>strong-motion</i> seismology the ground acceleration is commonly expressed as a fraction or percentage of the acceleration due to gravity ( $g$ ) where $g = 981 \text{ cm/s}^2$ . For the strongest earthquakes ground accelerations of more than 1.5 $g$ have been recorded. Since the weight of an object, e.g., of a building, is equal to its mass multiplied by the gravity $g$ , the additional acceleration due to ground shaking causes an extra load. This extra load may exceed the strength of the building and may cause damage or collapse. |
| <b>accelerogram</b>   | A record (time history) of ground <i>acceleration</i> as a function of time produced by an <i>accelerograph</i> during a seismic event.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>accelerograph</b>  | A compact, rugged, and relatively inexpensive <i>seismograph</i> designed to record the <i>signal</i> from an <i>accelerometer</i> , especially of strong ground shaking on ground or in structures, caused by nearby large earthquakes (see, e.g., Trifunac, 2009). Film used to be the most common analog recording medium. Modern accelerographs, however, record digitally with much larger <i>dynamic range</i> . Whereas a traditional <i>strong-motion</i> accelerograph begins recording when the motion exceeds a certain specified trigger level, at present, some <i>seismic networks</i> record acceleration within a large dynamic range continuously.                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>accelerometer</b>  | A sensor or <i>transducer</i> that converts <i>acceleration</i> of its base into an analog trace or electrical <i>signal</i> . The acceleration signal is normally obtained from the feedback current necessary to                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |

maintain an *inertial sensor* at its mechanical zero point. As in a linear actuator the current is equal to the applied force, this current is proportional to the acceleration within the bandwidth limits of the feedback. Outside the band the acceleration can still be extracted from the position signal, normally through a complicated *transfer function*. Accelerometers may be enclosed within a self-contained *accelerograph*, but for studies of structural response they can be located remotely, with their signals being transmitted to a central recorder. Unless otherwise specified, an accelerometer measures translational acceleration. However, accelerometers are also sensitive to *rotational ground motions*.

**acceptable risk**

The probability associated with a social or economic consequence of an earthquake that is low enough (in comparison with other risks) to be judged acceptable by appropriate authorities. It is often used to represent a realistic basis for determining design requirements for engineered structures, or for taking certain social or economic actions. What kind and level of risk is considered to be acceptable by a society depends also on its socio-economic and political conditions.

**ACH method**

ACH stands for Aki, Christoffersson and Husebye (1977) who developed a method for determining the three-dimensional structure of the Earth using the *teleseismic* observations from a two-dimensional *seismic array*. The method was extended to the *local earthquake* data by Aki and Lee (1976). It opened one way to the research area now called “*seismic tomography*”.

**accretionary prism**

See *accretionary wedge*.

**accretionary wedge**

*Trench* sediments and volcanics that accumulate and deform where oceanic and continental *plates* collide. These sediments are scraped off the top of the downgoing oceanic crustal plate and are added as a generally wedge-shaped mass to the leading edge of the continental plate (Von Huene and Scholl, 1991; Duff, 1993; see *subduction* and *tectonic plates*).

**accuracy**

According to Bevington and Robinson (2003) a measure of how close the results of an experiment (observed values or values calculated from observations via model assumptions) come to the “true” value.

**acoustic emission**

A term in rock mechanics indicating *seismic wave* radiation from dynamic formation of a microcrack. Due to the small crack lengths, the *frequencies* of the radiated waves are in the kHz (acoustic or ultrasound) range.

|                               |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|-------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>acoustic impedance</b>     | Product of wave propagation <i>velocity</i> $v$ and <i>density</i> $\rho$ of the medium in which the acoustic (or elastic seismic) waves propagate. See <i>seismic impedance</i> and <i>impedance contrast</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>active fault</b>           | A <i>fault</i> that is considered likely to undergo renewed movement within a period of concern to humans. Faults are commonly considered to be active if they have moved one or more times in the last 10,000 years, but they may also be considered active when assessing the hazard for very critical installations such as nuclear power plants even if movement has occurred in the last 500,000 years. Although faults that move in earthquakes today are active, not all active faults generate earthquakes - some are capable of moving aseismically (see <i>creep</i> , <i>silent earthquake</i> , <i>slow earthquake</i> ; also, e.g., Johnston and Linde, 2002). More precise attempts (usually unsatisfactory) have been made to define “active” faults for legal or regulatory purposes. See Yeats et al. (1997, p. 449). |
| <b>active margin</b>          | A continental margin that is also a <i>plate boundary</i> with earthquakes and/or volcanic activity.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>active tectonic regime</b> | A term that refers to regions where <i>tectonic deformation</i> is relatively large and earthquakes are relatively frequent, usually near <i>plate boundaries</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>active tectonics</b>       | <i>Tectonic</i> movements that are expected to occur or have occurred within a time span of concern to society.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>active volcano</b>         | By a widely used but poor definition, an active volcano is one that is erupting or has erupted one or more times in recorded history (see Decker and Decker, 1998). Active and potentially active (i.e., presently dormant) volcanoes occur in narrow belts that collectively comprise less than 1 percent of the Earth’s total surface (Simkin et al., 1994).                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>aftershock</b>             | An earthquake occurring as a consequence of a larger earthquake (referred to as the <i>mainshock</i> ) at roughly the same location. Aftershocks are smaller than the mainshock and within 1-2 <i>fault lengths</i> distance from the mainshock fault. The sequence of such earthquakes following a larger one generally shows a regular decrease in the rate of occurrence, first discovered by Omori (1894), indicating a stress relaxation and redistribution process as the rocks accommodate to their new post-earthquake state. Aftershocks can continue over a period of weeks, months, or years, decreasing in frequency with time. In general, the larger the mainshock, the larger and more numerous the aftershocks, and the longer they will continue. See Utsu (2002b), and Eq. (7.4) in Ben-Zion (2003).                 |
| <b>afterslip</b>              | The increase in <i>displacement</i> on a <i>fault</i> by <i>creep</i> following a sudden <i>coseismic slip</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|-------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>AIC</b>                          | <p>Akaike Information Criterion, applied in statistical modeling to get the most appropriate probability distribution that describes the observed data. Definition:</p> $AIC = -2 \cdot \max(L) + 2(p),$ <p>where <math>\max(L)</math> is the maximum likelihood function and <math>p</math> the number of free parameters. See Akaike (1970, 1971).</p>                                                                                                                                                                                                                                                                              |
| <b>airgun</b>                       | <p>A non-explosive <i>seismic source</i> normally used at sea. Compressed air (about 15 MPa) is fed into an air chamber of variable size (1-30 L) and is repetitively released into water to generate sound waves to be emitted.</p>                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>air-coupled surface wave</b>     | <p>Despite the great <i>density</i> contrast between air and ground, atmospheric <i>pressure</i> disturbances traveling over the Earth's surface can produce <i>surface waves</i> if the <i>phase velocity</i> is equal to the acoustic velocity in the air. Such a coupling with air has been observed for flexural waves in ice sheets floating on the ocean and for <i>Rayleigh waves</i> in ground with low-velocity surface layers. A simultaneous arrival of <i>air waves</i> and tidal disturbances was observed after the well known explosion of the volcano Krakatoa in 1883.</p>                                           |
| <b>air wave</b>                     | <p>Audible sounds or <i>infrasound</i> waves are sometimes generated by earthquakes; a <i>local earthquake</i> may sound like distant thunder, canon shots or the rumbling from a passing truck or tank on a bumpy road. Instrumental measurements show that these sounds arrive simultaneously with the first <i>P waves</i> (Hill et al., 1976). More long-<i>period</i> (seconds to hours) acoustic-gravity waves may also be excited by <i>great earthquakes</i>, as well as by volcanic explosions, meteorite falls, and nuclear blasts in the atmosphere. See Ben-Menahem and Singh (2000).</p>                                 |
| <b>Airy phase</b>                   | <p>Portions of dispersed wave trains associated with the maxima or minima of the <i>group velocity</i> as a function of <i>frequency</i>. The Airy function can be used for an approximate calculation of the <i>waveform</i>. Examples are continental <i>Rayleigh waves</i> at <i>periods</i> around 15 seconds, <i>mantle</i> Rayleigh waves at periods of 200 to 250 seconds, and <i>surface waves</i> of periods between some 6 and 10 seconds which travel across an ocean at a velocity near 1 km/sec. (see Fig. 2.10 of Chapter 2 in this Manual). See also Kulhanek (2002, p. 345), and Aki and Richards (1980, p. 256).</p> |
| <b>Akaike Information Criterion</b> | See <i>AIC</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>aleatory uncertainty</b>         | <p>The uncertainty in <i>seismic hazard analysis</i> due to inherent random variability of the quantity being measured. Aleatory uncertainties cannot be reduced by refining modeling or analytical techniques. See <i>epistemic uncertainty</i>.</p>                                                                                                                                                                                                                                                                                                                                                                                 |

- aliasing** A serious error in the *analog-to-digital conversion*, pointed out by Blackman and Tukey (1958), that occurs when the sampling rate is less than twice the highest *frequency* (*Nyquist frequency*) contained in the *signal* to be digitized. A long *period* ghost which does not exist in the original *signal* may appear after digitization (see Chapter 6 of this Manual; also Aki and Richards, 1980, p. 576-579).
- alluvium** Loose gravel, sand, silt, or clay deposited by streams after the last ice age (see *Holocene*).
- Alpide belt** Mountain belt with frequent earthquakes that extends from the Mediterranean region, eastward through Turkey, Iran and northern India.
- altitude of ambiguity** The topographic relief (or error) required to create one fringe in an *interferogram*. The height of ambiguity is expressed in units of meters.
- ambient noise** See *noise*.
- amplification** (by seismograph) Most earthquakes are relatively small, in fact, so small that no one feels them. In order for seismologists to see the recording of the ground movement from smaller earthquakes, the recording has to be made larger. Modern *seismographs* are able to magnify the *ground motion*  $10^6$  times or even more, i.e., they are able to resolve ground motion *amplitudes* as small as the diameter of molecules or even atoms. The frequency-dependent amplification of a *seismograph* depends on the input spectrum of the *seismic signal*, the *eigenfrequency* and *attenuation* of the *seismic sensor* or *seismometer*, and of the filter parameters and gain of the *seismic recorder* (see Chapters 4 to 6 of this Manual).
- amplification** (by recording site) The term is also used to describe the increase in *amplitude* of *seismic waves* and/or levels of shaking at the site of recording or observation. Increase (or decrease) of amplitudes may be due to focusing (or defocusing) of seismic wave energy caused by the *geometry/heterogeneity* of the *velocity structure* of sediments and other rock formations, by basin subsurface topography, or by surface topography. Most *seismographs* are installed at a site on or near the surface with irregular topography and lithologic structures of heterogeneous material created by aeons of weathering, erosion, deposition and other geological processes. These complex near-surface structures also tend to amplify the amplitude of incident seismic waves (see *site effect* and *site response*). The degree of amplification is usually dependent on the *frequency* and thus on the *wavelength* of the *seismic signal* with respect to the linear dimensions of the surface and subsurface heterogeneities.

|                                           |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|-------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>amplitude</b>                          | The size of the wiggles on an earthquake recording; more general and correct in a physical sense the height of a wave-like disturbance - such as the seismic record trace (called <i>waveform</i> ) - from the medium (zero) level to its peak, respectively trough. When measured from peak to trough it is called double or “peak-to-peak” (more correct “peak-to-trough”) amplitude. In seismology <i>ground motion</i> amplitudes are usually measured in nanometers ( $10^{-9}$ m) or micrometers ( $10^{-6}$ m). |
| <b>amplitude response</b>                 | In general, the complex <i>frequency response</i> (e.g., calculated as <i>Fourier transform</i> of the output <i>signal</i> divided by that of the input signal) of a system. Specifically, the frequency-dependent amplification and phase response of <i>ground motion</i> by a <i>seismograph</i> , a site, a building etc., depending on their eigenoscillation and <i>attenuation</i> properties (for <i>seismographs</i> see Chapters 4 and 5 and for <i>site response</i> Chapter 14).                          |
| <b>analog-to-digital converter</b>        | An electronic device that converts a voltage produced by a sensor to digital values by applying specific digitization rates depending on the frequency content of the original signal.                                                                                                                                                                                                                                                                                                                                 |
| <b>anelastic attenuation</b>              | <i>Amplitude</i> reduction of a <i>signal</i> (e.g., a seismic or electro-dynamic wave) due to energy dissipation (friction, heat). See <i>intrinsic attenuation</i> .                                                                                                                                                                                                                                                                                                                                                 |
| <b>anelasticity</b>                       | Loss of <i>strain</i> energy in a stressed medium due to internal <i>friction</i> caused by imperfections in the elasticity of the medium. A dimensionless measure of the internal friction or anelasticity of a medium is the <i>quality factor</i> <i>Q</i> .                                                                                                                                                                                                                                                        |
| <b>angular coherence function</b>         | Measure of <i>coherency</i> of two transmitted <i>wave fields</i> as a function of incidence angles.                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>anisotropic</b>                        | A medium is anisotropic if its physical properties depend on the direction. See <i>seismic anisotropy</i> .                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>annual probability (of exceedence)</b> | The probability that a given level of <i>seismic hazard</i> (typically some measure of <i>ground motion</i> , e.g., seismic <i>magnitude</i> , <i>intensity</i> or ground <i>acceleration</i> ), or <i>seismic risk</i> (typically economic loss or casualties) can be equaled or surpassed within an exposure time of one year.                                                                                                                                                                                       |
| <b>ansatz</b>                             | A German word used by mathematicians to describe an initial or trial solution.                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>anthropogenic earthquake</b>           | An <i>earthquake</i> that is a byproduct of human activities.                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>anticline</b>                          | A fold, generally convex upward, whose core contains the stratigraphically older rocks.                                                                                                                                                                                                                                                                                                                                                                                                                                |

|                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|--------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>anti-plane strain</b> | Assumed two-dimensional symmetry, used to simplify mathematical analysis or numerical calculation, in which <i>displacement</i> depends on only two spatial coordinates and its direction is perpendicular to the plane of those coordinates. <i>Strike-slip</i> displacement that does not vary along strike is an example of anti-plane <i>deformation</i> . Examples are a Love wave from a line source and mode 3 crack (see Figure 6 in IS 3.1., <i>plane strain</i> , <i>mode 3 crack</i> , <i>Love wave</i> , and Love, 1944). |
| <b>aperture</b>          | Largest horizontal distance between two sensors of a <i>seismic array</i> . See <i>array aperture</i> .                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>apparent stress</b>   | Defined by Wyss and Brune (1968) as the product of the <i>rigidity</i> and the ratio of the <i>seismic energy</i> to the <i>seismic moment</i> for an earthquake. It is the fictitious <i>stress</i> (radiation resistance) which acting through the actual earthquake <i>slip</i> would do the work equivalent to the seismic energy radiated in the faulting. More simply, it is the product of the <i>seismic efficiency</i> and the average absolute stress as explained in Aki and Richards (2002, p. 55). See also Chapter 3.   |
| <b>apparent velocity</b> | If $T$ (in s) and $D$ (in km) denote the <i>travel times</i> and <i>epicentral distances</i> , respectively for a <i>seismic phase</i> arriving at points of the Earth's surface, then the apparent velocity $v_{app}$ is defined as $dD/dT$ . The apparent velocity is for a spherical Earth model the reciprocal of the seismic <i>ray parameter</i> $p$ or more generally the reciprocal of the horizontal (radial) component $s_R$ of the <i>slowness vector</i> $s$ (see Chapters 2 and Chapter 9 of this Manual).               |
| <b>arc</b>               | See <i>volcanic arc</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>Archaean</b>          | An eon in geologic time, from 4 to 2.5 billion years ago. Archaean is from Greek and means “beginning” (see <i>geologic time</i> ).                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>Arias intensity</b>   | A <i>ground-motion</i> parameter derived from an <i>accelerogram</i> and proportional to the integral over time of the acceleration squared. Expressed in units of <i>velocity</i> (m/s or cm/s). More specifically a broadband measure of the strength of strong ground motion observed in an accelerogram, defined by the integral over all <i>natural frequencies</i> of the energy input to a damped <i>single degree of freedom</i> oscillator in response to an accelerogram.                                                   |
| <b>array (seismic)</b>   | See <i>seismic array</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>array analysis</b>    | The joint analysis of <i>seismograms</i> recorded in a network of <i>seismometers</i> , which all record with a common time base. <i>Seismic arrays</i> allow the estimation of the propagation direction of <i>seismic waves</i> and also the estimation of the <i>wave field's</i>                                                                                                                                                                                                                                                  |

|                                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|---------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                       | gradient and thus the estimation of various components of the <i>strain tensor</i> and the vector of <i>rotational motions</i> . See Chapter 9 in this Manual.                                                                                                                                                                                                                                                                                                                    |
| <b>array aperture</b>                 | The size of an <i>array</i> measured by the maximum separation between pairs of <i>seismometers</i> in an array. Traditionally arrays are divided up into small aperture (maximum separation around 3 km), medium aperture (maximum separation 10-25 km), and large aperture (maximum separation 100-200 km) depending on their size (see Chapter 9 in this Manual).                                                                                                              |
| <b>array beam</b>                     | The result of the summation of the output from several elements of an <i>array</i> ( <i>beamforming</i> by <i>delay-and-sum processing</i> ) to enhance <i>signals</i> with a given <i>apparent velocity</i> (or <i>slowness</i> ) and <i>backazimuth</i> of approach. By choosing velocity and azimuth expected for a <i>hypocenter</i> of the source, this process can enhance <i>seismic waves</i> coming from this particular source location (see Chapter 9 in this Manual). |
| <b>array processing</b>               | Techniques of <i>array analysis</i> , which are mainly based on <i>wave field coherence</i> and <i>correlation</i> techniques (see Chapter 9 in this Manual).                                                                                                                                                                                                                                                                                                                     |
| <b>arrival</b>                        | The appearance of a particular <i>seismic phase</i> (e.g., <i>P wave</i> ) on a <i>seismogram</i> .                                                                                                                                                                                                                                                                                                                                                                               |
| <b>arrival time</b>                   | The time of the first <i>onset</i> of a <i>seismic wave</i> (e.g., <i>P wave</i> ) as recorded by a <i>seismograph</i> . Arrival times are the basic data for locating earthquake <i>hypocenters</i> and for determining seismic <i>velocity</i> structures of the Earth.                                                                                                                                                                                                         |
| <b>ASC</b>                            | See <i>Asian Seismological Commission</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>Asian Seismological Commission</b> | The second regional commission of <i>IASPEI</i> . The ASC was founded in 1996, has Member States from all over Asia and holds General Assembly meetings every second year since its foundation. See <a href="http://www.asc1996.com">http://www.asc1996.com</a> .                                                                                                                                                                                                                 |
| <b>aseismic</b>                       | In seismology, an adjective for characterizing a <i>fault</i> (an area) on which (where) no earthquakes have been observed and are not likely to occur. Aseismic behavior may be due to lack of <i>shear stress</i> across the fault, a <i>locked fault</i> condition with or without shear stress, or release of stress by fault <i>creep</i> . In earthquake engineering, an adjective for structures which are designed and built to withstand earthquakes.                    |
| <b>aseismic slip</b>                  | <i>Slip</i> on a <i>fault</i> which is not associated with the radiation of <i>seismic waves</i> . See <i>fault creep</i> .                                                                                                                                                                                                                                                                                                                                                       |
| <b>ash</b>                            | See <i>volcanic ash</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                         |



|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
|----------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>ashfall</b>                   | See <i>fallout</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>asperity</b>                  | A region on a <i>fault</i> of higher strength than the surroundings, at which stress concentrates prior to fault rupture. In seismology, this term (and also the term <i>barrier</i> ) is often used to describe the <i>heterogeneity</i> of a fault. It was first introduced by Lay and Kanamori (1981) to describe regions of earthquake ruptures where relatively high release of <i>seismic moment</i> or release of <i>seismic energy</i> occurs [see <i>barrier</i> ; Scholz (2002)] or it may refer to locked sections of the fault that cause fault <i>segmentation</i> . An asperity may be produced by one or more of the following conditions: increased <i>normal stress</i> , high <i>friction</i> , low <i>pore pressure</i> , or geometric changes in the fault such as fault bends, offsets, or roughness. |
| <b>association (of arrivals)</b> | To assign a seismic wave <i>arrival</i> to a specific <i>seismic event</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>asthenosphere</b>             | The <i>ductile</i> part of the Earth's <i>upper mantle</i> , just below the brittle <i>lithosphere</i> . More specifically, the weak lower portion of the near-surface <i>thermal boundary layer</i> and underlying mantle which undergoes significant ductile <i>deformation</i> under long-term loads.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>asymptotic ray theory</b>     | The mathematical theory used to describe high-frequency <i>seismic rays</i> . The ray <i>ansatz</i> is constructed as a series in inverse powers of <i>frequency</i> in the frequency domain, or integrals of the source impulse in the time domain, with <i>amplitude</i> coefficients and <i>travel times</i> only as functions of position along the ray. See <i>Maslov asymptotic ray theory</i> .                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>attenuation</b>               | Waves are largest where they are formed and gradually get smaller as they move away from their generating source. The decrease in wave <i>amplitude</i> is caused by <i>geometrical spreading</i> and by attenuation of wave energy. The latter is due to two different processes: a) intrinsic attenuation (also termed anelastic attenuation or absorption) in different Earth materials and/or b) <i>scattering</i> of <i>seismic energy</i> at <i>heterogeneities</i> in the Earth (e.g., <i>faults</i> or small-scale anomalous geological bodies; see Figures 2.37 and 2.38 in Chapter 2). $Q$ and $\kappa$ (kappa) are attenuation parameters used in modeling the attenuation of <i>ground motions</i> (see section 2.5.4.2 in Chapter 2).                                                                         |
| <b>attenuation relation</b>      | Term used particularly in the fields of <i>strong ground motion</i> and earthquake engineering to describe the decay of peak spectral ground motion parameters (such as displacement, <i>peak velocity</i> , or <i>acceleration</i> ) as a function of distance from the <i>source</i> or a ruptured <i>fault</i> , its <i>magnitude</i> , fault type, and sometimes also the specific travel path and site conditions. Its coefficients are usually derived from statistical analysis of earthquake records (see, e.g., Chapter 15 in this Manual and                                                                                                                                                                                                                                                                     |

Campbell (2003). Similar relations are expected for peak ground rotation parameters.

**attribute (of an arrival)** A quantitative measure of a seismic *arrival* such as *onset* time, *(back)azimuth*, direction of first motion, *slowness*, *period* and *amplitude*.

**A-type earthquake** Earthquakes with clear *P* and *S* waves, occurring beneath volcanoes. They are indistinguishable from normal shallow *tectonic earthquakes*. In the original definition by Minakami (1961), it was contrasted to the *B-type earthquake* attributed to an extremely shallow *focal depth*. In recent years the contrast has been attributed to the source process, and it is more often called volcano-tectonic (VT), or *high-frequency* (HF) *earthquake* (see Chapter 13 of this Manual).

**autocorrelation** Calculating the *correlation* function of a *signal* with itself.

**autocorrelation coefficient** Parameter that describes the *correlation* between values of a function or samples of a discrete time series separated by time shifts in case of continuous functions or time lags in case of discrete time series. The autocorrelation  $c$  is the autocovariance normalized by the variance in case of a stationary process.

**autoregressive process** Model of a time series, where the actual value  $x_j$  is described by a linear combination of the  $p$  predecessors:

$$x_j = \sum_{i=1}^p a_i x_{j-i} + e_j,$$

where  $a_i$  denotes the  $j - i$ -th autoregressive (AR) coefficient and  $e_i$  denotes white noise or the prediction error.  $p$  denotes the order of the AR process.

**average annualized loss (AAL)** The average economic loss expected per year for a specific property, portfolio of properties, or region as a result of one or more damaging earthquakes.

**axial deformation** A *deformation* having an axial symmetry, characterized by a lengthening or shortening parallel to a given direction or axis, with a shortening or lengthening perpendicular to the axis that is the same in all directions.

**axis of rotation** An axis parallel to the *rotation vector*. Each rotation can be presented in terms of angle of rotation and axis of rotation. See *rotation tensor*.

**axial tensor, vector or scalar** A tensor (vector or scalar) object which changes the sign when we change the orientation of a system of coordinates from right-handed to the left-handed or vice versa. They do not depend on the coordinate system itself. For instance, *couple stress tensor*, *Levi-Civita tensor*, *torque*, *rotational velocity*, and mixed

product of basis vectors of the coordinate system are axial objects. Many axial objects are related to (see different entries under) *rotation*. Axial objects sometimes are called pseudo objects. They cannot be equal to polar objects, which do not depend on the orientation of a coordinate system.

**azimuth** In general, a direction measured clock-wise in degrees against the North. In seismology, used to measure the direction from a *seismic source* to a *seismic station* recording this event. The azimuth in the opposite direction from the station to the source is called *backazimuth*.

**azimuthal anisotropy** A term used when the *seismic wave* properties depend on the azimuth of propagation in the horizontal plane. For example, *Pn* waves in the neighborhood of an oceanic spreading center exhibit azimuthally anisotropic velocity but also *SKS* and *SKKS* waves due to *upper mantle* flow anisotropy (see Fig. 2.8 in Chapter 2 of the Manual).

## B

**back arc** The region adjacent to a volcanic arc, and on the opposite side of the oceanic *trench* and subducting *plate* (see also *volcanic arc*).

**back-arc basin** A basin floored by oceanic *crust* behind a *volcanic arc* (See *back arc*; also Duff, 1993; Uyeda, 1982).

**backazimuth** The direction from the *seismic station* towards a *seismic source*, usually measured in degrees clock-wise from North (cf. *azimuth*). The backazimuth of shear energy can be estimated from co-located measurements of *translation* and *rotations*.

**background noise** Permanent movements of the Earth as seen on seismic records caused by ocean waves, wind, rushing waters, turbulences in air-pressure, etc. (ambient natural noise), and/or by traffic, hammering or rotating machinery, etc. (man-made noise). See *microseisms*.

**background seismicity** *Seismicity* that occurs at a more or less steady rate in the absence of an *earthquake sequence*, such as a *swarm* or *foreshock-mainshock-aftershock* sequence. In *seismic hazard analysis*, it is the seismicity that cannot be attributed to a specific *fault* or source zone.

**back-scattering** See *scattering* and *scattering theory*.

**backstop** Continental rocks in the *back arc* that are landward from the trace of the *subduction thrust fault* and that are strong enough to support *stress* accumulation. These rocks are both igneous and

dewatered, lithified, consolidated sediments that probably were part of the *accretionary wedge*. The softer accretionary-wedge rocks are strongly deformed as they accumulate against the backstop. The exact position and dip direction of the backstop is not well determined, and more than one backstop may exist.

**ballistic projectiles**

A collective term for fragmental volcanic products (*tephra*) of diverse shapes greater than 64 mm in size. Such projectiles can exit the volcanic vent at speeds of tens to hundreds of meters per second on trajectories that are little affected by the eruption dynamics or the prevailing wind direction; thus, they are typically restricted to within 5 km of vents. See *bomb*, *volcanic*, *block*, *volcanic*, and *cinder*.

**band-pass filter**

A *filter* which removes low and high *frequency* portions of the input *signal* outside of the *passband* (see, e.g., Figs. 4.10, 4.14, 4.16, 4.18, and 4.42-4.45 in Chapter 4 of this Manual).

**bandwidth**

A range between high-pass and low-pass cut-off frequencies (see *high-pass filter* and *low-pass filter*). In seismology, bandwidth is often meant as the portion of the *frequency-response* amplitude spectrum that is approximately flat; and it is often defined as the frequency band between the upper and lower frequencies, where the amplitude falls 3 *dB* below its maximum value (called the -3 dB points).

**barrier**

A site on a *fault* surface of higher strength than the surroundings that is capable of terminating rupture or across which rupture may jump. In seismology, this term (and also the term *asperity*) is often used to describe *heterogeneity* of a fault, and was first introduced by Das and Aki (1977) to construct a two-dimensional fault model. See Scholz (2002).

**basalt**

A general term for mafic igneous rocks (or corresponding melts) produced by partial melting of the *upper mantle*, and comprising most of the *crust* beneath the oceans.

**base isolation**

A technique to reduce the earthquake forces in a structure by the installation of horizontally flexible devices at the foundation level. Such devices greatly increase the lowest natural periods of the structure for horizontal motions and thereby lower the accelerations experienced by the structure. The most common applications of base isolation are to old historic buildings, hospitals and bridges.

**base-line correction**

A term used in the processing of time series analysis. It corrects a recorded *signal* for the bias in the zero value, and any long-period drift in the zero level that may arise from *instrumental noise*, or from the digitization of analog signals.

|                                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
|--------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>basement</b>                      | In geology, the igneous and metamorphic rocks that underlie the sedimentary deposits and extend downward to the base of the <i>crust</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>base shear</b>                    | The horizontal shearing force at the base of the structure. The maximum base shear, typically in the form of a fraction of the weight of the structure, is an important parameter in earthquake response studies and in earthquake resistant design.                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>base surge</b>                    | Highly turbulent, dilute cloud of <i>volcanic ash</i> , air, and steam, which expands radially away from the base of an <i>eruption column</i> at high velocity. It commonly forms when an eruption starts in a lake or coastal environment and involves the explosive interaction of hot <i>magma</i> with water.                                                                                                                                                                                                                                                                                                                           |
| <b>basin-induced surface wave</b>    | <i>Surface waves</i> generated at the edges of a basin or at a strong lateral <i>discontinuity</i> inside the basin by incident <i>body</i> ( <i>S</i> or <i>P</i> ) <i>waves</i> . Their <i>amplitudes</i> and <i>waveforms</i> depend on the basin structure along the ray path from the edge (or the discontinuity) to the site. If the site lies in the central part of a large sedimentary basin, the basin-induced surface waves appear much later than the direct and reverberated <i>S</i> waves. The term “secondary surface waves” sometimes used for them is not appropriate since their amplitudes can be of primary importance. |
| <b>basin-transduced surface wave</b> | <i>Surface waves</i> transduced at edges of a basin or at a strong lateral <i>discontinuity</i> inside the basin from incident surface waves. They are distinguished from <i>basin-induced surface waves</i> by the difference in incident waves. These surface waves are observed if the earthquake source is shallow and distant. <i>Refraction</i> and mode conversion of both <i>Love</i> and <i>Rayleigh waves</i> and transformation among them occur at the edge of a basin.                                                                                                                                                          |
| <b>Bayesian method</b>               | Any method accepting not only the interpretation of a probability as the limit of an experimental histogram, but also as the representation of subjective information. See Mosegaard and Tarantola (2002), and the biography of Thomas Bayes in Howell (2003).                                                                                                                                                                                                                                                                                                                                                                               |
| <b>beamforming</b>                   | See <i>array beam</i> , <i>delay-and-sum processing</i> and Chapter 9 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>bedrock</b>                       | A relatively hard, solid rock that commonly underlies soil or other softer unconsolidated sedimentary materials. The bedrock is a subset of the <i>basement</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>bending-moment fault</b>          | <i>Fault</i> formed due to bending of a flexed layer during folding. <i>Normal faults</i> characterize the convex side, placed in tension,                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |

and *reverse faults* characterize the concave side, placed in compression.

**Benioff zone**

Also called *Wadati-Benioff zone* (see also *subduction zone*).

**blind fault**

A *fault* that does not extend upward to the Earth's surface, being "buried" under the uppermost layers of sediments or rock in the *crust* with no surface evidence. It usually terminates upward in the axial region of an *anticline*. If its *dip* is  $<45^\circ$ , it is a blind thrust. The Northridge thrust fault responsible for the 1994 Northridge, California, earthquake is an example of a blind fault. See Yeats et al. (1997).

**block-and-ash flow**

Small-volume *pyroclastic flow* characterized by a large fraction of dense to moderately vesicular juvenile blocks in a matrix of the same composition. Such flows are typically associated with collapse of *Vulcanian eruption* columns or are produced from partial or complete collapse of unstable, viscous *lava domes* as they over-steepen at their fronts during active growth (e.g., as during the eruption of Unzen Volcano, Japan, in 1991). See Freundt et al. (2000).

**block (or blocky) lava**

A term sometimes applied to all *lava flows* with fractured surfaces and are covered with angular fragments (including *aa*). However, as emphasized by Macdonald (1972), the term is more appropriately restricted to those lava flows, which, even though similar in overall structure to *aa*, lack the jaggedly spinose features of *aa* and have fragments characterized by more regular forms (often approaching cube-like) and smoother surfaces.

**block rotation**

Rotation of crustal blocks in continental areas. Block rotation about a vertical axis is common along *strike-slip faults*, but horizontal-axis block rotation commonly occurs in *thrust* and *normal fault* systems, and presumably oblique-axis block rotation can occur along *oblique-slip faults*. See Twiss et al. (1993).

**block, volcanic**

A ballistic projectile ( $> 64$  mm in size) but more angular in shape than volcanic bombs, indicating that it was thrown out in a state too solid to allow any changes in shape during flight. See *bomb, volcanic*.

**body wave**

Waves that propagate through the interior of a body are called body waves, as opposed to *surface waves*, which propagate along the boundary surface of a body. In seismology, the Earth is usually considered an infinite, linear elastic and *homogeneous* continuum. In such a medium there exist only two basic types of seismic body waves with different *polarization*: (1) dilatational (or *longitudinal*) *P waves* and (2) translational (*shear*) *S waves* (see Chapter 2 in this Manual).

- body-wave magnitude** Earthquake *magnitude* calculated from *amplitude/period* ratios of *body waves*. *mB* and *mb* are based on data recorded by relatively broadband and short-period *seismographs*, respectively. The original body-wave magnitude introduced by B. Gutenberg (1945) and Gutenberg and Richter (1956) is denoted by  $m_B$ . It has been determined by measuring the largest ratio of displacement amplitudes and periods of vertical and/or horizontal component *P* and *PP* waves as well as horizontal component *S* waves. Different from this, the more or less equivalent new IASPEI (2011) standard, denoted *mB(BB)* or *mB\_BB*, is based only on direct measurement of the largest vertical component broadband velocity amplitude in the whole P-wave train (Bormann and Saul, 2008). Body-wave magnitudes assigned by USGS and ISC were up to 2011 only short-period *mb* (denoted *Mb* by the ISC), however, both agencies will in future additionally determine *mB\_BB*. Also *mb* has been re-defined in a unique way (IASPEI 2005 and 2013) in order to avoid the sometimes significant differences between the *mb* values issued by different agencies. See Chapter 3 and IS 3.3 in this Manual, Bormann et al. (2009) and Bormann (2011).
- bomb, volcanic** A ballistic projectile greater than 64 mm in size that is erupted in a semi-solid state and fluid enough to change shape aerodynamically while flowing through the air. Volcanic bombs can take on a great variety of shapes (see also *block, volcanic*).
- borehole sensor** A sensor installed below the Earth's surface in a borehole (see, e.g., section 7.4.6 in Chapter 7 of this Manual).
- Born approximation** A perturbation method that iteratively solves an *inhomogeneous* wave equation, where waves are decomposed into primary waves which satisfy the *homogenous* wave equation and scattered waves of small *amplitude*. See Sato et al. (2002), and the biography of Max Born in Howell (2003).
- branch (of travel-time curve)** Term used in seismology for “branching” *travel-time curves* that are related to discrete ray paths of the same type of wave and due to strong velocity *gradients* and/or *low-velocity layers* in the Earth's interior. For a *prograding* travel-time branch the rate of travel time *t* increases with increasing *epicentral distance* *D* (in km) (or  $\Delta$  in degree), i.e., the derivative  $dt/dD = \text{ray parameter } p = \text{horizontal component of slowness } s_R$  is decreasing, in contrast to a *retrograde* (or *receding*) travel-time branch, for which this derivative is increasing. This is equivalent with  $dD/dp < 0$  for prograde and  $dD/dp > 0$  for retrograde branches. Thus, at the distance where the prograde and the retrograde branches merge (i.e., where the sign of  $dD/dp$  changes), wave energy is focussed, resulting in strongly increased amplitudes (*caustic point*; see also Figs. 2.30 - 2.33 in

Chapter 2 of this Manual). Retrograde branches may be formed by strong positive velocity increases with depth. When this gradient weakens again at larger depth, a new prograde branch is forming. Such a sequence of prograde, retrograde and prograde travel-time branches in the order of decreasing  $dt/dD = p$  for the same wave type is called a *triplication*. Examples are the two *P* wave triplications due to the *upper mantle discontinuities* (strong velocity gradient zones) at 410 km and 660 km depth (see Figs. 2.32 and 2.5.3.2 in Chapter 2 of this Manual) and the *PKP* triplication with the branches PKPab, PKPbc, PKiKP and PKPdf (Fig. 11.73 in Chapter 11) due to the strong velocity increase from the outer to the inner Earth *core* (see Fig. 2.79 in Chapter 2. For this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

|                                        |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
|----------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>breccia</b>                         | A coarse-grained clastic rock composed of angular broken rock fragments in a fine-grained matrix.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>brittle</b>                         | Unable to accommodate inelastic <i>strain</i> without loss of strength and tending to break.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>brittle-ductile boundary</b>        | See <i>brittle-plastic transition</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>brittle-ductile transition</b>      | A zone in the Earth's <i>crust</i> across which the thermo-mechanical properties of the crust change from <i>brittle</i> behavior above (tending to break) to <i>ductile</i> behavior below (plastic or quasiplastic, tending to bend). Its depth is usually identified as the maximum focal depth of local earthquakes and occurs at depths at which the temperature lies between 200 to 400 degrees. This is typically around 15 km but depends on the <i>heat flow</i> conditions and age of the crust. Most earthquakes initiate at or above this depth on steep (high-angle) <i>faults</i> . Below this depth, fault slips may be <i>aseismic</i> and may grade from high angle to low angle faulting (e.g., Ito 1990). |
| <b>brittle-failure (BF) earthquake</b> | Sometimes used in <i>volcano seismology</i> (see Chapter 13 of this Manual) to distinguish earthquakes dominated by <i>brittle</i> processes (frictional slip or formation of a tensional crack) from those in which fluids are thought to play an active role (such as " <i>LP</i> " earthquakes or " <i>harmonic tremor</i> ". The designation "BF earthquake" is generally synonymous with "high-frequency", " <i>volcano-tectonic</i> ", or " <i>A-type</i> " earthquake.                                                                                                                                                                                                                                                |
| <b>brittle-plastic transition</b>      | A simplified, more precise model of the <i>brittle-ductile transition</i> in which the plastic fracture is considered as the <i>deformation</i> mechanism in the <i>ductile</i> part.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>broadband magnitude</b>             | <i>Body-wave</i> and <i>surface-wave magnitudes</i> that are based (in contrast to <i>amplitude/period</i> ratios measured on more or less narrow-band records of ground <i>displacement</i> ) on direct                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |



measurements of ground motion velocity on broadband records. Examples are the IASPEI (2011) recommended *P*-wave magnitude standard  $mB(BB)$  or  $mB\_BB$ , measured in the period range between 0.2 s to 30 s, and the new *surface-wave* magnitude standard  $Ms(BB)$  or  $Ms\_BB$ , measured at periods between 3 s and 60 s. As compared to more narrow-band magnitudes, such as  $mb$ , measured at periods around 1 s, and the surface-wave magnitude  $Ms$ , measured at periods around 20 s, broadband magnitudes are less prone to the underestimation of magnitude for large *seismic events* (in the case of  $mb$ ) or smaller events (in the case of  $Ms(20) = Ms\_20$ ). See Chapter 3 and IS 3.3 in this Manual and Bormann et al. (2009).

**broadband seismogram** *Seismograms* recorded by a *broadband seismograph*. See Chapter 5 of this Manual.

**broadband seismograph** To avoid the strong *ambient noise* caused by ocean waves (*microseisms*), two types of *seismograph* have traditionally been used to record seismic *signals*: one for *periods* longer than about 10 s and the other for those shorter than 3 s. However, this kind of filtering may result in significant signal distortions and underestimation of earthquake *magnitude*. A broadband seismograph can record faithfully seismic signals in a *frequency* range of 3 decades or some 10 octaves (e.g., between 0.1 s to 100 s) or even wider, thanks to the improved linearity range of the seismometer and *dynamic range* of the recorder. See Chapter 5 of this Manual.

**broadband seismometer** *Seismic sensor* used in a *broadband seismograph*.

**Brownian (molecular) motion** Completely irregular (stochastic) motion of smallest particles suspended in a liquid or gas, first described by the British botanist Robert Brown (1773-1858). The diffusion intensity of B.m.m. increases with temperature and is explained in a physically and mathematically tractable form by the theories of Albert Einstein (1879-1955) and Marian Smoluchowski (1872-1917). It plays a role in the temperature-dependent diffusive internal *friction* and thus *anelasticity* of Earth materials, the related *attenuation* of *seismic waves*, and may also be one of the causes of *instrumental noise* of *seismic sensors* (see Section 5.5.5 in Chapter 5 of this Manual).

**Brune model** The omega-squared circular-crack model proposed by Brune (1970). The physical basis of this model was refined by Madariaga (1976), who solved the mathematical problem of seismic wave radiation from an expanding circular crack that stops. See *omega-squared model*, Chapter 3, Fig. 3.5 in this Manual and Aki and Richards (2002 or 2009, p. 560-565),

***b*-slope** Another name for *b-value*.

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|-------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>B-type earthquake</b>            | Earthquakes occurring under volcanoes with emergent onset of <i>P</i> waves, no distinct <i>S</i> waves, and a lower <i>frequency</i> content as compared with the usual <i>tectonic earthquakes</i> of the same <i>magnitude</i> . In the original definition by Minakami (1961) its character difference from the <i>A-type earthquake</i> was attributed to the extremely shallow <i>focal depth</i> (< 1 km). More recently the difference has been attributed to the source process and now called more often <i>long-period (LP)</i> or <i>low-frequency earthquakes</i> . See Chapter 13 in this Manual. |
| <b>bulk density</b>                 | The mass of a material divided by its volume, including the volume of its pore spaces.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>bulk modulus</b>                 | The modulus of volume elasticity which relates a change in volume to a change in hydrostatic pressure. It is the reciprocal of <i>compressibility</i> . See Birch (1966) for an introduction and a compilation of values, section 2.2 in Chapter 2 of this Manual and Eq. (1.6) in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                             |
| <b>Burridge &amp; Knopoff model</b> | A spring-and-block model proposed by Burridge and Knopoff (1967) for earthquakes. See Madariaga and Olsen (2002, p. 176-7).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b><i>b</i>-value</b>               | A coefficient in the frequency-magnitude relation, $\log N(M) = a - bM$ , obtained by Gutenberg and Richter (1941; 1949), where <i>M</i> is the earthquake <i>magnitude</i> and <i>N(M)</i> is the number of earthquakes with magnitude greater than or equal to <i>M</i> . Estimated <i>b-values</i> for most seismic zones fall between 0.6 and 1.1, but may be higher in volcanic regions, earthquake <i>swarms</i> and sometimes in <i>aftershock</i> sequences. See also the biographies of Beno Gutenberg and Charles Francis Richter in Howell (2003).                                                   |
| <b>Byerlee's law</b>                | The common observation from laboratory <i>friction</i> experiments that the coefficients of friction for nearly all rocks that comprise the <i>lithosphere</i> fall in the range 0.6 to 1.                                                                                                                                                                                                                                                                                                                                                                                                                      |

## C

|                                |                                                                                                                                                                                                                                                                                                                                                                                                     |
|--------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>c</b>                       | This symbol is used to indicate the <i>reflection</i> at the <i>core-mantle boundary</i> for waves incident from the <i>mantle</i> . For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).                                                                                                                              |
| <b><sup>14</sup>C age date</b> | An absolute age obtained for geologic materials containing bits or pieces of carbon using measurements of the proportion between the radioactive carbon ( <sup>14</sup> C) and the non-radioactive carbon ( <sup>12</sup> C). These dates are independently calibrated with calendar dates. The method is used to determine when past earthquakes occurred on a fault. See <i>paleoseismology</i> . |

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|-------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>caldera</b>                      | A circular or ovoid depression, generally 5 to 20 km in diameter, formed by collapse or subsidence of the central part of a volcano, often associated with large explosive eruptions.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>calibration</b> (of seismograph) | The process of determining the <i>response function</i> and <i>sensitivity</i> of a seismic instrument. See Chapter 5 in this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>calibration pulse</b>            | An electronic <i>signal</i> used to calibrate seismic instruments. See <i>calibration</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>capable fault</b>                | A mapped <i>fault</i> that is deemed a possible site for a future earthquake with <i>magnitude</i> greater than some specified threshold.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>cascade model of earthquakes</b> | An extended version of the <i>characteristic earthquake model</i> in which several consecutive <i>fault segments</i> can rupture in a variety of combinations. The <i>slip</i> on each segment is assumed to be characteristic to the segment.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>catalog of earthquakes</b>       | A chronological listing of earthquakes. Early catalogs were purely descriptive, i.e., they gave the location and date of each earthquake and some description of its effects. Modern catalogs are usually quantitative, i.e., earthquakes are listed as a set of numerical parameters describing <i>origin time</i> , <i>hypocenter</i> location, <i>magnitude</i> , <i>moment tensor</i> , etc. See Engdahl and Villasenor (2002) for an instrumentally determined earthquake catalog (1900-1999), Utsu (2002a) for a catalog of deadly earthquakes (1500-2000), and Ambraseys et al. (2002, p. 759-761) and Schweitzer and Lee (2003) for a discussion of historical and modern earthquake catalogs and <i>seismic bulletins</i> . |
| <b>causal signal</b>                | An output <i>signal</i> that does not exist for times prior to the application of an impulsive input. In other words, an impulse response $h(t)$ that vanishes for $t < 0$ . See <i>impulse response</i> , <i>linear system</i> and <i>convolution</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>caustic point</b>                | A point on the Earth's surface where the rate of change of the <i>epicentral distance</i> traveled by a <i>seismic ray</i> , respectively of its <i>slowness</i> , with the change of its <i>take-off angle</i> changes its sign. A large concentration of arriving <i>seismic energy</i> will generally be observed at the caustic point. See Figs. 2.30 to 2.3 in Chapter 2 of this Manual.                                                                                                                                                                                                                                                                                                                                        |
| <b>Cenozoic</b>                     | An era in geologic time, from 66.4 million years ago to the present. See <i>geologic time</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>cgs system of units</b>          | An obsolete system of units based on: (1) centimeter for length, (2) gram for mass, and (3) second for time. Some derived cgs units with special names are: (1) force in dyne ( $\text{cm g s}^{-2}$ ), (2)                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |

energy in erg ( $\text{cm}^2 \text{ g s}^{-2}$ ), and (3) acceleration of free fall in gal ( $\text{cm s}^{-2}$ ). See *international system of units*.

- channel** In observational seismology, it is the *signal* output of a component of a *seismic sensor*. Modern seismic recorders (see *data logger*) are able to record simultaneously the signals from many channels.
- chaos** A term that had been used since antiquity to describe various forms of randomness. It is now a commonly accepted technical term referring to the irregular, unpredictable, and apparently random behavior of deterministic dynamic systems. Solutions to deterministic equations are chaotic if adjacent solutions diverge exponentially in phase space. See Socolar (2002) or Wolfram (2002) for a general discussion of chaos, and Turcotte and Malamud (2002) for an application in seismology.
- characteristic earthquake model** A fault-specific earthquake model in which a given *fault segment* generates an earthquake of a size and mechanism determined from the geometry of the segment. At a specific location along a fault, the *displacement (slip)* is the same in successive characteristic earthquakes. Other earthquakes occurring on the fault are much smaller than these. In its application to *seismic hazard analysis*, it refers to an earthquake of a specific size that is known or inferred to recur at the same location, usually at a less frequent rate than that extrapolated from the frequency-magnitude relation for smaller earthquakes in the area. See Aki (2002, p. 44-45), Grant (2002, p. 482-484), and Eq. (7.3) and Figure 15 in Ben-Zion (2003).
- characteristic function** Non-linear transformation of the *seismogram*, its information being exploited by a picking algorithm. Introduced by Rex Allen (1978). See Chapter 16 in this Manual.
- Christoffel matrix** A symmetric real-valued matrix, whose eigenvalues give the seismic *velocities* and eigenvectors give the *polarization* directions of the waves. See Chapman (2002), Cara (2002), and the biography of Elwin Brun Christoffel in Howell (2003).
- cinder** A common fragmental volcanic product (*tephra*), classified as ranging between 2 and 64 mm in size.
- cinder cone** A relatively small, cone-shaped hill or mountain, rarely exceeding 1 km in height, that nearly always has a truncated top in which there is a bowl-shaped crater (see Macdonald, 1972). As its name implies, this volcanic edifice is built entirely of *tephra* ejected during moderately explosive eruptions. See *cinder*.

|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|----------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>circular fault</b>            | <i>Fault plane</i> geometry in which the rupture front starts from a point on the fault and spreads as a circle. See <i>rupture front</i> and <i>Brune model</i> .                                                                                                                                                                                                                                                                                                                            |
| <b>Circum-Pacific belt</b>       | The zone surrounding the Pacific Ocean that is characterized by frequent and strong earthquakes and many volcanoes, as well as high <i>tsunami</i> hazard. Also called the <i>Ring of Fire</i> .                                                                                                                                                                                                                                                                                              |
| <b>classical continuum</b>       | An idealized material medium that is assumed in classical mechanics (see e.g., Goldstein, 1950). The medium has a continuously distributed mass, and the <i>stress</i> is a function of 6 independent strain measures, characterizing purely <i>translational</i> (volumetric and shear one) <i>strain</i> .                                                                                                                                                                                  |
| <b>clip level</b>                | The maximum output of a <i>seismograph</i> . A “hard clip” is defined as the maximum voltage output of a <i>seismometer</i> regardless of input. A “soft clip” is defined as the <i>amplitude</i> at which the output begins to distort into a nonlinear representation of the input waveform. For an ideal sensor, the two clip levels are (nearly) the same, with no distortion or nonlinearity before the hard clip is reached.                                                            |
| <b>coda</b>                      | The “tail” of a <i>seismic signal</i> , which follows a well defined wave <i>arrival</i> . For example, the teleseismic P-coda follows the arrival of teleseismic <i>P waves</i> . Coda waves are due to <i>scattering</i> and superposition of multi-path arrivals, with usually exponentially decaying <i>amplitudes</i> (e.g., Figs. 2.44 and 2.45 of Chapter 2 and Figure 1b of DS 11.1 in this Manual, and Kulhanek (2002, p. 333-334).                                                  |
| <b>coda attenuation</b>          | See <i>coda Q</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>coda normalization method</b> | A method to use the power of local earthquake S-coda at a given lapse time as a measure of earthquake radiation energy, which is used for measurements of <i>site amplification</i> factors, <i>attenuation</i> per unit travel distance, and source energy. This method is based on the assumption of a uniform spatial distribution of coda energy density of a local earthquake at a long lapse time measured from the earthquake <i>origin time</i> . See Sato et al. (2002, p. 200-201). |
| <b>coda Q</b>                    | A parameter characterizing the <i>amplitude</i> decay of S-coda of a local earthquake with the <i>lapse time</i> defined by Aki and Chouet (1975) assuming single S to S <i>scattering</i> . The coda lasts longer for higher coda Q.                                                                                                                                                                                                                                                         |
| <b>coda phase</b>                | A <i>detection</i> of a single phase of unknown path found within the coda signal envelope, designated as tx, Px or Sx.                                                                                                                                                                                                                                                                                                                                                                       |
| <b>coda waves</b>                | The <i>seismograms</i> of an earthquake usually show some vibrations long after the passage of relevant <i>body wave</i> and                                                                                                                                                                                                                                                                                                                                                                  |

*surface wave onsets* in the considered distance range. This portion of the *seismogram* to its end is called the *coda*. Coda waves are believed to be *back-scattering* waves due to *inhomogeneities* in the Earth. They have been extensively used to obtain source spectra, as well as to measure seismic *attenuation* and *site amplification* factors by the *coda normalization method*. See Sato et al. (2002; p. 198-201). Measurements of *rotations* in the *coda wave field* may help to separate P and S energy (see Pham et al., 2009).

|                                                     |                                                                                                                                                                                                                                                                                                       |
|-----------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>coefficient of friction (<math>\mu_f</math>)</b> | Coefficient of the linear relationship between <i>shear stress</i> ( $\tau$ ) and <i>normal stress</i> ( $\sigma_n$ ) on a <i>fault</i> or joint that is slipping $\tau = S_o + \mu_f \sigma_n$ . See Lockner and Beeler (2002, p. 506), and section 5 in Ben-Zion (2003).                            |
| <b>coherent</b>                                     | <i>Seismic signals</i> detected on various <i>seismic sensors</i> of an <i>array</i> or network of <i>seismic stations</i> are said to be coherent if they are related to each other in time, <i>amplitude</i> and <i>waveform</i> shape because they come from the same <i>seismic source</i> .      |
| <b>coherence or coherency</b>                       | The degree to which two <i>wave fields</i> are in <i>phase</i> and similar in shape. Mathematically it is the normalized cross-power spectrum of the two wave fields. It is the <i>frequency-domain</i> counterpart of the “ <i>correlation</i> ” in the <i>time domain</i> . See Sato et al. (2002). |
| <b>cohesion (<math>S_o</math>)</b>                  | The inherent shear strength of a <i>fault</i> or joint surface; shear strength at zero <i>normal stress</i> in the equation $\tau = S_o + \mu_i \sigma_n$ . See Lockner and Beeler (2002, p. 506).                                                                                                    |
| <b>cohesionless</b>                                 | Referring to the condition of a sediment whose shear strength depends only on <i>friction</i> because there is no bonding between the grains. This condition is typical of clay-free sandy deposits. See <i>cohesion</i> .                                                                            |
| <b>cohesive force</b>                               | The force that acts in the cohesive zone of a crack located between the freely slipping crack surface and the intact elastic body ahead of the crack tip. The existence of a cohesive zone removes the <i>stress singularity</i> at the crack tip. See Aki and Richards (2002, p. 548-552).           |
| <b>collision zone</b>                               | A zone where two continents collide. Continental collision is the cause of mountain building or <i>orogeny</i> . See Uyeda (2002), and Duff (1993).                                                                                                                                                   |
| <b>colluvial wedge</b>                              | A prism-shaped deposit of fallen and washed material at the base of (and formed by erosion from) a <i>fault</i> scarp or other slope, commonly taken as evidence in outcrop of a scarp-forming event.                                                                                                 |

|                                         |                                                                                                                                                                                                                                                                                                                         |
|-----------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>colluvium</b>                        | Loose soil or rock fragments on or at the base of gentle slopes or hillsides. Deposited by or moving under the influence of rain wash or downhill <i>creep</i> .                                                                                                                                                        |
| <b>column collapse</b>                  | A condition that occurs when an explosive <i>eruption column</i> , composed of volcanic gases, <i>tephra</i> , and air, becomes denser than the ambient atmosphere and collapses to ground level.                                                                                                                       |
| <b>compatibility condition</b>          | The ability of two or more fields to fulfill the required mutual relationship. For example, a condition imposed on the <i>strain</i> field so that it can be expressed by the derivatives of the <i>displacement</i> field.                                                                                             |
| <b>compensated linear-vector dipole</b> | The force system representing a crack opening or closing under the constraint of no volume change. The corresponding <i>moment tensor</i> has zero trace (purely deviatoric). Traditionally called “cone-type mechanism” in Japan. See Sipkin (2002).                                                                   |
| <b>complex site effect</b>              | Dynamic <i>amplification</i> effects, e.g., those arising from a <i>resonance</i> condition, generated by propagation of earthquake waves in 2D/3D near-surface geological configurations, such as sedimentary valleys, and topographic irregularities. See Chapter 14 of this Manual, and Faccioli and Pessina (2003). |
| <b>component</b>                        | (1) One dimension of a three-dimensional <i>signal</i> , (2) The vertically- or horizontally-oriented (north or east) sensor of a <i>seismic station</i> .                                                                                                                                                              |
| <b>composite volcano</b>                | Synonymous with <i>stratovolcano</i> .                                                                                                                                                                                                                                                                                  |
| <b>compressibility</b>                  | Reciprocal of <i>bulk modulus</i> .                                                                                                                                                                                                                                                                                     |
| <b>compressional tectonic setting</b>   | A region undergoing lateral contraction and for which the vertical <i>principal stress</i> is the minimum. See McGarr et al. (2002).                                                                                                                                                                                    |
| <b>compressional wave</b>               | See <i>P wave</i> .                                                                                                                                                                                                                                                                                                     |
| <b>compressional stress</b>             | The <i>stress</i> that squeezes something. It is the stress component normal to a given surface, such as a <i>fault plane</i> . It results from forces applied perpendicular to the surface or from remote forces transmitted through the surrounding rock.                                                             |
| <b>compressive strength</b>             | The maximum compressive stress that can be applied to a material before failure occurs. See <i>strength</i> .                                                                                                                                                                                                           |
| <b>confidence ellipsoid</b>             | In <i>hypocenter</i> location calculations, an ellipsoid surrounding the computed hypocenter within which the true hypocenter is (formally) located at some specified level of confidence (see IS 11.1 in this Manual).                                                                                                 |

## Glossary

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                |
|-------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>conical wave</b>                 | See <i>head wave</i> .                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>Conrad discontinuity</b>         | Seismic boundary between the upper and middle <i>crust</i> that is usually defined by an increase in seismic <i>velocity</i> from 6.2-6.4 km/sec to about 6.6-6.8 km/sec. The term has fallen into disuse in recent years due to the lack of universality of such a <i>discontinuity</i> . See Mooney et al. (2002), and the biography of Victor Conrad in Howell (2003).                                      |
| <b>constitutive equation</b>        | For continuous media, an equation relating basic physical quantities characterizing the state of the material - for instance, the relation between <i>strains</i> , temperature, the gradient, and <i>stresses</i> of a continuum. Constitutive equations define the behavior of the material, and they are different for different materials.                                                                 |
| <b>continental deformation</b>      | A term usually used to emphasize the contrast between deforming zones in the oceans and on the continents (see <i>continental tectonics</i> ).                                                                                                                                                                                                                                                                 |
| <b>continental drift</b>            | A theory that continents have displaced thousands of kilometers in the last few hundred million years, first proposed by the German meteorologist and geophysicist Alfred Wegener (1912 a and b). See Uyeda (2002), and biography of Alfred Wegener in Howell (2003).                                                                                                                                          |
| <b>continental tectonics</b>        | A term used to include the large-scale motions, interactions and deformation of the continental <i>lithosphere</i> . It is often used in contrast to <i>plate tectonics</i> . Whereas deforming zones in the oceanic <i>plates</i> are usually narrow and confined, on continents they are often spread over wide areas, requiring a different approach to their description and analysis. See Jackson (2002). |
| <b>contractional jog</b>            | A stepover between <i>en échelon fault</i> strands where <i>slip</i> transfer involves contraction and increased mean compressive <i>stress</i> within the <i>stepover</i> . See Sibson (2002).                                                                                                                                                                                                                |
| <b>contractive soils</b>            | Granular soils that decrease or tend to decrease in volume during large <i>shear deformation</i> . The tendency to contract increases pore water pressures in undrained saturated soils during shear. Slopes and embankments underlain by contractive soils may suffer catastrophic strength loss and flow failure during earthquake shaking. See Youd (2003).                                                 |
| <b>controlled-source seismology</b> | Seismic investigations that utilize man-made <i>seismic sources</i> , such as chemical explosions detonated in boreholes or water, vibrators at land or <i>airguns</i> in water. See Mooney et al. (2002).                                                                                                                                                                                                     |
| <b>convergent margin</b>            | <i>Plate boundary</i> zone where two <i>plates</i> are converging to one another. Either plate may subduct under, collide into, or obduct over the other plate. See Suyehiro and Mochizuki (2002).                                                                                                                                                                                                             |



- conversion relationships** Statistical relationships between parameter data of similar kind, e.g., different types of *magnitude* estimates. Such relationships are used for converting, with some uncertainty, one type of data into the other one, e.g., *body* and/or *surface-wave magnitude* values into the equivalent *moment-magnitude* values. This is common practice in efforts to create *homogeneous* - with respect to magnitude - *earthquake catalogs*. Usually, conversion relationships are derived via *least-square standard regression analysis*, although for data afflicted with comparable measurements and initial errors, as is the case with many types of magnitude data, *Chi-squared* or *orthogonal regression* relationships are preferable. See, e.g., Stromeyer et al. (2004), Castellaro et al. (2006), Castellaro and Bormann (2007), Gutdeutsch et al. (2011), Lolli and Gasperini (2012).
- converted waves** Conversion of *P* to *S* waves and *S* to *P* waves occurs at a *discontinuity* for non-normal *incidence angle* (see Figs. 2.34, 2.35 and 2.52 in Chapter 2 of this Manual). Converted waves sometimes show distinct *arrivals* on *seismograms* (e.g., Figs. 2.25 and 2.52 in Chapter 2; 11.65, 11.67 and 11.70 in Chapter 11) and may be used to determine the location of the discontinuity.
- convolution** A mathematically equivalent operation that describes the action of a linear (mechanical and/or electronic) system on a signal, such as that of a *filter* on a *seismic signal*. The convolution of two functions  $x(t)$  and  $h(t)$  is often written as  $x(t) * h(t)$ , and is defined mathematically as  $y(t) = x(t) * h(t) = \int x(\tau) h(t - \tau) d\tau$ . In signal processing,  $y(t)$  represents an output of a *stationary linear system* in terms of the input  $x(t)$  and impulse response  $h(t)$  of the system. See *linear system* and *impulse response*.
- core (Earth's)** Central part of the Earth's interior with upper boundary at a depth of about 2900 kilometers. It represents about 16% of the Earth's volume with about 33% of the Earth's mass. It is divided into an inner, solid core, and an outer, fluid core. The outer core extends from about 2900 to about 5120 km below the Earth's surface and consists in its main components of a mixture of liquid iron and nickel. The inner core is the central sphere of the Earth with a diameter of about 1250 km and consists of solid metal. For wave *velocity*, *density* and *attenuation* distribution in the core, as compared to the Earth's *mantle*, see Fig. 2.79 in Chapter 2 of this Manual. For further information see Jacobs (1987), Lay (2002), and Song (2002).
- core-mantle boundary (CMB)** Sharp *discontinuity* between the bottom of the Earth's *mantle* and the Earth's *core*, characterized by a decrease of the *P-wave velocity*  $v_p$  from 13.7 to 8.0 km/s and of the *S-wave velocity* from 3.44 to 0 km/s (see Fig. 2.79 in Chapter 2 and Tables in DS 2.1 of this Manual).

- core shadow zone** A gap in the emergence of *P* waves that have traveled directly through the Earth extending for short-period *P* waves from *epicentral distances* of about 103° to 114°. This gap or shadow of the core is caused by the sharp decrease in the *P* wave velocity at the *core-mantle boundary*. Note, however, that long-period mantle *P* waves may be diffracted around the curvature of that boundary, thus feeding mantle *P*-wave energy of strong earthquakes into this shadow beyond 140° (see *diffracted P* (*Pdif*), Figs. 2.55 of Chapter 2 and 11.73 and 11.74 of Chapter 11 in this Manual, and Kulhanek (2002, p. 340).
- Coriolis force** Inertia force named after the French engineer and physicist Gaspard Gustave Coriolis (1792-1843), which describes mechanical processes in rotating reference systems. The C.f. always acts perpendicular to the momentary direction of movement and the axis of *rotation* defined by the *rotation vector*  $\omega$ . It acts on each mass moving on the Earth surface and results in a rightward drift on the northern hemisphere and a leftward drift on the southern hemisphere. The C. f. needs to be taken into account, e.g., in the analysis of the spectral splitting of *free oscillations* when investigating lateral inhomogeneities of the Earth's structure (see Lay and Wallace, 1995, p. 162; Aki and Richards, 2002, p. 375).
- corner frequency** The *frequency* at which the curve representing the *Fourier amplitude spectrum* of a recorded *seismic signal* abruptly changes its slope and drops by -3 db from the spectral plateau. In earthquake source studies, a parameter characterizing the *far-field body-wave displacement* “source spectrum”, being related to *fault size*, *rupture velocity*, *source duration* and *stress drop* in the source. See *omega-squared model*, Fig. 3.5 in Chapter 3 of this Manual, and Figure 7 and Eq (6.1) in Ben-Zion (2003). In *seismometry* the frequency at which the *transfer function* (*magnification curve*) of a recording system changes its slope and drops by -3 dB from its plateau (maximum) level. See Figures in IS 5.2 in this Manual, also *bandwidth* and *band-pass filter*).
- corrected acceleration** An *acceleration* time history that has been “corrected” from the raw data recorded by an *accelerograph*. The correction typically involves removing *drift*, spikes, and any distorting effects created by the instrument or digitizing process. See Shakal et al. (2003).
- corrected penetration resistance** A measure of an in-situ property of soil, important for interpreting geotechnical and seismic *site response* data. It is based on a penetration test, such as standard penetration-test resistance, corrected for the influence of overburden *pressure*, hammer energy, rod length, borehole diameter, and sampler

|                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|--------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                | type; or cone penetration resistance, corrected for the influence of overburden pressure. See Youd (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>correlation</b>             | The degree of relative correspondence, as between two sets of data, expressed, e.g., by the <i>correlation coefficient</i> . See also <i>autocorrelation</i> and <i>crosscorrelation</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>correlation coefficient</b> | In statistics a descriptive index applied to two sets of numbers (x,y), the value of which serves to specify the overall dependence (in per cent) exhibited by the data between the variables x and y.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>correlation techniques</b>  | Different mathematical methods and numerical algorithms to estimate the <i>coherence</i> between data sets and time-series by investigating their <i>auto-</i> and <i>cross-correlation</i> relationships.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>coseismic</b>               | Occurring at the same time as an earthquake.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>Cosserat continuum</b>      | A continuum, whose point bodies (infinitesimal particles) possess both <i>translational</i> and <i>rotational</i> degrees of freedom. In the Cosserat continuum, the <i>stress tensor</i> is asymmetric, and there exists a <i>couple-stress tensor</i> . Apart from the full Cosserat continuum, there exists a Cosserat pseudo-continuum, in which <i>translations</i> and <i>rotations</i> are not independent (in the linear approximation, the <i>rotation vector</i> equals $\text{curl } \mathbf{u}/2$ , where $\mathbf{u}$ is the <i>translational displacement</i> ), and a reduced Cosserat continuum, in which rotations and translations are independent but the <i>couple-stress tensor</i> is zero. |
| <b>Cosserat elasticity</b>     | An elasticity theory introduced by the Cosserat brothers (Eugène and François Cosserat, 1909), in which each material point has six degrees of freedom, three of which correspond to <i>translation</i> , as in the classical theory, and the other three to <i>rotation</i> . See Pujol (2009).                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>Cosserat length</b>         | A characteristic length associated with a rotation independent field, and defined for the given type of continuum.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Coulomb friction</b>        | In a constitutive law (see <i>constitutive equation</i> ) governed by Coulomb friction, a <i>slip</i> occurs when <i>shear stress</i> exceeds strength defined by a coefficient of <i>friction</i> (a constant) times the <i>normal stress</i> . In contrast, the rate-and-state dependent law involves the coefficient of friction, which is not a constant but depends on the sliding speed (rate) and contacts (state) on the <i>fault plane</i> . See Harris (2003), and Eq. (5.1) in Ben-Zion (2003). See also the biography of Charles Augustin de Coulomb in Howell (2003).                                                                                                                                |
| <b>couple stress tensor</b>    | Let $\mathbf{n}$ be the unit vector corresponding to the normal to an elementary material surface. A tensor $\boldsymbol{\mu}$ of second rank is called the couple-stress tensor, if $\mathbf{n} \cdot \boldsymbol{\mu}$ gives the <i>torque</i> acting upon                                                                                                                                                                                                                                                                                                                                                                                                                                                      |

this surface. The couple-stress tensor plays the same role for *torques* as the *stress tensor* does for forces.

**Crary waves**

Crary waves are a train of sinusoidal waves with nearly constant *frequency* observed on a floating ice sheet (Crary, 1955). They are multi-reflected *SV waves* with horizontal *phase velocities* near the speed of compressional waves in ice.

**crater**

A bowl- or funnel-shaped depression, generally in the top of a volcanic cone, often the major vent for eruptions; the distinction between the terms crater and *caldera* is blurred, but caldera tends to be applied to depressions with diameters >5 km.

**creep**

Slow, more or less continuous movement occurring on *faults* due to ongoing *tectonic deformation*. Also applied to slow movement of *landslide* masses down a slope because of gravitational forces. Faults that are creeping release shear strain without significant radiated seismic energy, i.e., they do not tend to have large earthquakes. This fault condition is commonly referred to as *unlocked*. See Lockner and Beeler (2002).

**creep events**

Episodic *slip* across a *fault* trace observed at the surface over minutes to days. See Johnston and Linde (2002).

**creep meters**

Instruments for measuring *displacement* across a *fault* trace on the Earth's surface, usually with a baseline of length around 10 meters. See Johnston and Linde (2002) and IS 5.1 on *strainmeters* in this Manual.

**creep waves**

Regional strain waves propagating along *active faults* as suggested by Savage (1971). However, these slip waves have not yet been directly observed. See Johnston and Linde (2002).

**Cretaceous pulse of spreading**

During the Cretaceous period (124-83 Ma) when there was no *geomagnetic reversal*, *sea-floor spreading* was 50-75% faster than in other periods. This phenomenon is interpreted as caused by *mantle* superplume activity. See Uyeda (2002), and Larson (1991).

**critical damping**

Critical damping exists if the *damping* coefficient of the special damping device of an harmonic *oscillator* (nicknamed “dashpot”; see Fig. 5.5 in Chapter 5) is  $h = 1$ . This value marks the transition of free vibration of the oscillator from the under-damped case of exponentially decaying harmonic motion to a-periodic motion that decays without changing algebraic sign. Widely used classical *seismometers* had damping values ranging from about 0.5 (WWSSN short-period and the Russian Kirnos *seismographs* SKM-III and SKD), 0.7 (modern electrodynamic seismometers such as GS13, CMG-3T or STS-2 and the re-calibrated Wood-Anderson standard response

according to Uhrhammer and Collins, 1990), up to near critical 0.98 (WWSSN long-period). Systems with very low *attenuation* may experience damage when excited to *resonance*.

**critical facilities**

Man-made structures whose ongoing performance during an emergency is required or whose failure could threaten many lives. May include (1) structures such as nuclear-power reactors or large dams whose failure might be catastrophic, 2) major communication, utilities and transportation systems, (3) high-occupancy buildings such as schools or offices, and (4) emergency facilities such as hospitals, police and fire stations, and disaster-response centers.

**critically stressed crust**

A state of *stress* in the Earth's *brittle crust* balanced by the frictional strength of the Earth's crust.

**cross-axis sensitivity** (rotation and translation)

The susceptibility of a *sensor* to produce an additional *signal* in response to motion other than that for which the sensor is intended to be sensitive. It is often due to limitations in the design or manufacture of the mechanical suspension in the sensor, or misalignment of the *components* within the sensor. There are five cross-axis sources for every sensitive axis of an *inertial sensor* (Nigbor *et al.*, 2009). A translational *accelerometer* may be sensitive to both off-axis translational axes and to *rotations* about any axis. See Trifunac and Todorovska (2001).

**cross-correlation coefficient**

Parameter, that describes the correlation between the values of two functions or independent samples of a discrete time series, separated by time shifts in case of continuous functions or time lags in case of discrete time series. See also *autocorrelation* and *correlation*.

**crossover**

The distance from an event where two different *phases* arrive at the same time, allowing constructive interference that sometimes enhances the *signal amplitudes*.

**cross-talk**

The appearance of a *signal* on a *channel* due to electrical leakage from a signal on an adjacent channel. It is often due to electrical problems with shielding, grounding, etc. See Shakal *et al.* (2003).

**crust (Earth's)**

The outmost layer of the Earth above the *Mohorovičić discontinuity* ("Moho" for short). The crust in continental regions is about 25-75 km thick, and that in oceanic areas is about 5-10 km thick. (For global cross section see Fig. 2.10 in Chapter 2 of this Manual). It represents less than 0.1% of the Earth's volume, with rocks that are chemically different from those in the *mantle*. The uppermost 15-35 km of the crust is *brittle* enough to produce earthquakes. The seismogenic crust is separated from the lower crust by the *brittle-ductile boundary*.

The crust is usually characterized by *P-wave velocities* below 8 km/s (average velocity of about 6 km/s). See also *mantle (Earth's)*, Mooney et al. (2002), and Minshull (2002).

- cultural noise** Seismic *noise* generated by various human activities such as traffic, construction works, industries, etc. See Chapter 4, Sub-Chapter 7.2 and IS 7.3 of this Manual and Webb (2002).
- cumulative seismic moment** The sum of the *seismic moments* of earthquakes which have occurred in a region from a fixed starting time till time *t*, as a function of *t*. It can be related to the cumulative *tectonic displacement* across a *plate boundary* lying in the region, or to the cumulative tectonic *strain* within the region.
- curl** In mathematics, the curl is the vector operator describing the “*rotation*” of a vector field. Curl **A**, or rot **A**, is equal to  $\nabla \times \mathbf{A}$ , the vector product of the gradient operator and vector or tensor **A**. In seismic *wave fields* based on linear *isotropic elasticity*, the curl operator separates out the *S-wave* field because the curl of the *P-wave* field is zero.
- cyclic mobility** A condition in which *liquefaction* occurs and flow *deformation* commences in response to static or dynamic loads but deformation is arrested by dilation at moderate to large shear strains. See Youd (2003).
- Cyclic Resistance Ratio (CRR)** The capacity of a granular soil layer to resist *liquefaction*, expressed as a ratio of the dynamic *stress* required to initiate liquefaction to the static effective overburden *pressure*. See Youd (2003).
- Cyclic Stress Ratio (CSR)** The seismic load placed on a granular soil layer, expressed as the ratio of the average horizontal dynamic *stress* generated by an earthquake to the static effective overburden *pressure*. See Youd (2003).

## D

- D''** A ~ 200 km thick layer surrounding the fluid *core* of the Earth occupying the lowermost *mantle*. D'' had been defined by Keith Bullen (1950) and is believed to be a region of strong thermal and chemical *heterogeneity*. See Lay (2002).
- damage probability matrix** A matrix giving the probabilities of sustaining a range of *losses* for various levels of a ground-motion parameter.
- damage scenario** A representation of the possible damage caused by an earthquake to the built environment in an area, in terms of parameters useful for economical and engineering assessment

or post-earthquake emergency management. See Faccioli and Pessina (2003).

|                                              |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|----------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>damping</b>                               | The reduction in <i>amplitude</i> of a <i>seismic wave</i> or <i>oscillator</i> due to <i>friction</i> and/or the internal absorption of energy by matter. In seismometry, a specific damping device (symbolized as a “dashpot”; see Fig. 5.5 in Chapter 5 in this Manual) is added in the design of a <i>seismometer</i> to diminish the effects of <i>resonance</i> and of the transient <i>free oscillation</i> . Fig. 5.22 in Chapter 5 of this Manual shows the dependence of normalized resonance curves on the damping coefficient $h$ . In vibration analysis, a term that indicates the mechanism for the dissipation of the energy of motion. Viscous damping, which is proportional to the velocity of motion and is described by linear equations, is used to define different levels of <i>response spectra</i> and is commonly used to approximate the energy dissipation in the lower levels of earthquake response. See <i>critical damping</i> . |
| <b>damping constant</b>                      | A parameter quantifying the effect of damping.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>data</b>                                  | Series of observations, measurements or facts.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>data acquisition</b>                      | Process of acquiring and storing <i>data</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>database</b>                              | Systemized collection of <i>data</i> that can be manipulated by data processing systems for specific purposes.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>data logger</b>                           | A digital <i>data acquisition</i> unit, usually for multi-channel recordings.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>data processing</b>                       | Handling and manipulating of <i>data</i> by computer.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>debris avalanche</b>                      | A flowing mixture of debris, rock, and moisture that moves down-slope under the influence of gravity. Principally differs from debris flows in that they are not water saturated. Debris avalanches are almost always associated with sector collapse at volcanoes. See Vallance (2000).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>decibel (dB)</b>                          | A logarithmic measure of relative <i>signal</i> power, defined as $10 \log(P/P_o)$ , or $20 \log(A/A_o)$ , where $P_o$ is the power and $A_o$ the signal <i>amplitude</i> at some reference level, typically the minimum or maximum signal resolvable by the recording instrument ( $P_o = A_o^2$ ). The <i>dynamic range</i> of seismic recorders is expressed in units of dB.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>declustering (of earthquake catalogs)</b> | The procedure for removing dependent earthquakes such as <i>foreshocks</i> and <i>aftershocks</i> . A declustered <i>earthquake catalog</i> contains only <i>mainshocks</i> and isolated earthquakes. The largest shock in each earthquake sequence is considered as the mainshock and remains in the catalog. See Utsu (2002b).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |

## Glossary

|                                           |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|-------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>deconvolution</b>                      | A procedure that removes the unwanted effects of <i>convolution</i> on a <i>signal</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>delay-and-sum processing</b>           | A procedure for forming an <i>array beam</i> . Time delays are applied to the output of each of the <i>seismometers</i> of an array to bring the desired <i>signal</i> recorded at the different seismometer positions into phase, followed by summing of the outputs with appropriate weights. See Chapter 9 of this Manual.                                                                                                                                                                                                                                                                                                 |
| <b>defining (of an arrival)</b>           | An arrival <i>attribute</i> , such as <i>arrival time</i> , <i>azimuth</i> , <i>slowness</i> , or <i>amplitude</i> and <i>period</i> , which is used in the calculation of location or magnitude of the <i>seismic source</i> .                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>deformation</b>                        | A change in the original shape and/or volume of a material due to <i>stress</i> and <i>strain</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>density</b>                            | Either (1) the quantity of something per unit measure such as unit length, area, volume, or <i>frequency</i> (see, for example, <i>power spectral density</i> ), or (2) the mass per unit volume of a substance under specified conditions of <i>pressure</i> and temperature.                                                                                                                                                                                                                                                                                                                                                |
| <b>depth phases (pP, pwP, pS, sP, sS)</b> | Depth phases are called all <i>seismic waves</i> which propagated upward from the <i>hypocenter</i> , turned into downward propagating waves by <i>reflection</i> at the free surface. These phases are useful for an accurate determination of the focal depth and their <i>phase names</i> (codes) indicate their phase type before and after the reflection. For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). Further see section 2.6.3 of Chapter 2, Figure 7 in IS 11.1 of this Manual, Kulhanek (2002), Engdahl and Villasenor (2002). |
| <b>design earthquake</b>                  | The postulated earthquake (commonly including a specification of the <i>design ground motion</i> at a site) that is used for evaluating the earthquake resistance of a particular structure.                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>design ground motion</b>               | A level of ground motion used in structural design. It is usually specified by one or more specific <i>strong-motion</i> parameters or by one or more time series. The structure is designed to resist this motion at a specified level of response, for example, within a given <i>ductility</i> level. See Borchardt et al. (2003).                                                                                                                                                                                                                                                                                         |
| <b>design spectrum</b>                    | The specification of the required strength or capacity of the structure plotted as a function of the <i>natural period</i> or <i>frequency</i> of the structure and of the <i>damping</i> appropriate to earthquake response at the required level. Design spectra are often composed of straight line segments and/or simple curves, for example, as in most building codes, but they can also be constructed from statistics of <i>response spectra</i> of a suite of ground motions appropriate to the <i>design earthquake(s)</i> . To be implemented, the requirements of a design spectrum are                          |



|                                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
|------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                          | associated with allowable levels of <i>stresses</i> , <i>ductilities</i> , <i>displacements</i> or other measures of response. See <i>response spectra</i> ; also Housner (1970).                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>detection</b>                         | Identification of an <i>arrival</i> of a <i>seismic signal</i> with <i>amplitudes</i> above and/or waveform and spectral content different from <i>seismic noise</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>deterministic earthquake scenario</b> | A representation, in terms of useful descriptive parameters, of an <i>earthquake</i> of specified size postulated to occur at a specified location (typically an <i>active fault</i> ), and of its effects. See Faccioli and Pessina (2003).                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>deterministic methods</b>             | Refers to methods of calculating <i>ground motions</i> for hypothetical earthquakes based on earthquake-source models and wave-propagation algorithms that exclude random effects.                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>deterministic system</b>              | A dynamic system whose equations and initial conditions are fully specified and are not <i>stochastic</i> or random. See Turcotte and Malamud (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>deviatoric moment tensor</b>          | A <i>moment tensor</i> with no <i>isotropic</i> (explosive or implosive) component. The sum of its eigenvalues is zero (see IS 3.9 of this Manual).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>dextral fault</b>                     | See <i>right-lateral fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>differential pressure gauge</b>       | A device for measuring long period (0.1 to 2000 seconds) <i>pressure</i> fluctuations on the seafloor. See Webb (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>differential travel time</b>          | The difference between the <i>arrival time</i> of one <i>seismic phase</i> and that of another phase recorded at the same station from the same source. Differential travel times are often used to eliminate the uncertainty in the event <i>origin time</i> and to reduce the influence of the <i>velocity</i> structure near the source or receiver.                                                                                                                                                                                                                                                                                            |
| <b>diffracted P (Pdif)</b>               | The <i>P wave</i> that follows a ray path from the <i>seismic source</i> that grazes the Earth's <i>core</i> , is diffracted along the <i>core-mantle boundary</i> , and emerges at an <i>epicentral distance</i> of about 97° on. Diffracted P waves are observed, especially longperiod ones, in the <i>shadow zone</i> beyond distances of 100° up to about 140° (see Figs. 2.60 and 2.61 of Chapter 2 and Figs. 11.74 and 11.77 of Chapter 11 in this Manual as well as Aki and Richards (2002, p. 456-457). For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). |
| <b>diffracted S (Sdif)</b>               | The <i>S wave</i> that follows a ray path from the <i>seismic source</i> that grazes the Earth's <i>core</i> , is diffracted along the <i>core-mantle boundary</i> . See the analogues phase <i>diffracted P (Pdif)</i> and for                                                                                                                                                                                                                                                                                                                                                                                                                    |

this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**diffraction**

A wave propagation phenomenon, which can be explained by the principle of Huygens-Fresnel but not by the high-frequency approximation of geometrical optics theory. Diffraction commonly occurs when the size of an obstacle becomes comparable with (and thus can be “sensed” by) the *wavelengths* of the *incident wave*.

**diffusion**

A process describing transport of mass or heat caused by differences in chemical potential, pressure or temperature. It can also describe *seismic energy* transport in randomly *heterogeneous* elastic media. See Sato et al. (2002), and Johnston and Linde (2002).

**digital accelerograph**

See *digital seismograph*.

**digital filter**

A mathematical tool designed to modulate the *frequency* characteristics of a discrete time series by means of numerical calculations on a computer. See Scherbaum (2002), and Kinoshita (2003).

**digital seismograph**

A *seismograph* in which the analog output *signal* from a *seismometer* is first converted to digital samples and then stored as time-series data in digital form. This contrasts with an analog seismograph, which directly records the analog output signal on paper, film, or magnetic tape.

**digital signal processor (DSP)** The processor implements mathematical operations, such as a *Fast Fourier transform*, numerically. Recent electronic *analog-to-digital converters* use DSP’s to attain high resolution and to remove *aliasing*. See Chapter 6 of this Manual.

**digitization**

The conversion of an analog waveform, e.g., a *seismogram* recorded on film, to a series of discrete x-y values corresponding to time-acceleration pairs, for use in subsequent computer processing. See Shakal et al. (2003).

**digitization noise**

See *noise* and *least significant bit*.

**dilatancy**

Inelastic increase of a volume of rock caused by the formation of cracks. Dilatancy may occur in a rock mass prior to an earthquake with some observable effects, such as changes in elevation, local wave propagation *velocity*, electrical conductivity, changes in well or ground water levels, spring discharges and effects of temporary hardening against *fault* rupture.

**dilative**

Granular soils that increase or tend to increase in volume during shear.

|                                   |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|-----------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>dilatometers</b>               | Borehole <i>strainmeters</i> measuring volumetric <i>strain</i> . See Johnston and Linde (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>dip</b>                        | Inclination of a planar geologic surface from the horizontal (measured in degrees).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>dip slip</b>                   | The component of the <i>slip</i> parallel with the <i>dip</i> of the fault. See <i>fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>dip-slip fault</b>             | A <i>fault</i> in which the <i>slip</i> is predominantly in the direction of the <i>dip</i> of the fault.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>Dirac tensors</b>              | The invariant tensors defined in four dimensions with use of the numbers: 0, 1, -1 and the imaginary unit <i>i</i> or - <i>i</i> . These tensors remain unchanged under any transformation in 4-D and help to define the invariant fields. These tensors are symmetric, anti-symmetric or asymmetric.                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>directivity</b>                | A concept in <i>seismic source</i> studies that was introduced by Ben-Menahem (1959). An effect of a propagating <i>fault</i> rupture whereby the <i>amplitudes</i> and <i>frequency</i> of the generated <i>ground motions</i> depend on the direction of wave propagation with respect to fault orientation, <i>slip</i> direction(s) and rupture velocity. The directivity and thus the radiation pattern is different for <i>P</i> and <i>S</i> waves. See also <i>radiation pattern</i> , <i>source directivity effect</i> and <i>Doppler effect</i> , Eq. (3.8) and Figure 6 in Ben-Zion (2003), section 9.1.1 in Lay and Wallace (1995), and the animations in Chapter 3 of this Manual. |
| <b>directivity focusing</b>       | The variation in wave <i>amplitude</i> as a function of azimuth relative to <i>strike</i> or <i>dip</i> of the <i>fault</i> source caused by the propagation of the <i>rupture front</i> . See animations in Chapter 3 of this Manual and <i>directivity</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>directivity pulse</b>          | A concentrated pulse-like <i>ground motion</i> generated by constructive interference of <i>S</i> waves traveling ahead of the tip of a rupturing <i>fault</i> . See <i>directivity</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>direct wave</b>                | A <i>seismic wave</i> that travels directly from the <i>seismic source</i> to the recording <i>station</i> , without being reflected or refracted at <i>interfaces</i> within the Earth. Observation of local, direct <i>P</i> and <i>S</i> waves are of particular interest for <i>focal depth</i> determination.                                                                                                                                                                                                                                                                                                                                                                              |
| <b>disaster</b>                   | Accidental or uncontrollable event, actual or threatened, in which individuals or society undergo severe danger or damage that disrupts the social structure of a society.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>discrete wavenumber method</b> | A simple and accurate method for calculating the complete <i>Green's function</i> for a variety of problems in elastodynamics. See Bouchon (2003) for a review.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
|-------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>discontinuity</b>                | A boundary ( <i>interface</i> ) between adjacent media of different physical properties, such as <i>density</i> , wave propagation <i>velocity</i> and/or others (e.g., the <i>Mohorovičić discontinuity</i> ). One may differentiate between sharp first order discontinuity (with rapid or step-like changes) and second order discontinuities with gradual changes of such properties. It depends on the ratio between <i>wavelength</i> $\lambda$ and the thickness $d$ of the transition zone between the two media whether seismic waves “sense” a discontinuity as being “sharp” ( $\lambda \gg d$ ) or gradual ( $d \gg \lambda$ ). |
| <b>discriminant</b>                 | A characteristic feature of a seismic or other <i>signal</i> that can be used for <i>discrimination</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>discrimination</b>               | The work of identifying different types of <i>seismic events</i> , and specifically of distinguishing between earthquakes, mining or quarry blasts, and underground nuclear explosions, and/or between underground nuclear and chemical explosions. See Chapter 17 and IS 11.2 in this Manual, Richards (2002) and Richards and Wu (2011).                                                                                                                                                                                                                                                                                                  |
| <b>disk loop</b>                    | A storage device that continuously stores new <i>waveform data</i> while simultaneously deleting the oldest data on the device.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>dislocation</b>                  | In seismology, the <i>displacement discontinuity (slip)</i> associated with faulting on an internal surface. In physics, dislocation generally refers to the line bounding of an internal surface subject to uniform slip, in which case the segments of the dislocation may be described as edge, screw, or mixed depending upon whether the slip is perpendicular, parallel, or oblique to the line. See Agnew (2002), Madariaga and Olsen (2002), Teisseyre and Majewski (2002), and Johnston and Linde (2002), and Savage (1980). See also <i>dislocation modeling</i> .                                                                |
| <b>dislocation modeling</b>         | Determining the dislocation distribution that would produce the observed surface <i>deformation</i> of an earthquake. See <i>dislocation</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>dispersion</b>                   | Spreading of wave duration with propagation distance due to <i>frequency-dependent velocity</i> . See Chapter 2 in this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>dispersion relations</b>         | See <i>Kramers-Kronig relations</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>displacement (rotational)</b>    | It is often used as a synonym for <i>rotation</i> vector.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>displacement (translational)</b> | A vector from the point where a material particle initially was located, to its location after an increment of motion. Unless otherwise specified, it means <i>translational</i> displacement in seismology. In earthquake geology, a general term for the movement of one side of a <i>fault</i> relative to the other; the amount of displacement may be measured in any chosen direction,                                                                                                                                                                                                                                                |

usually along the *strike* and the *dip* of the fault. In *seismometry* displacement is the *ground motion* commonly inferred from a *seismogram*. For example it may be calculated by double integration of an *accelerogram* or a single integration of a velocity-proportional recording with respect to time. It is expressed in units of length, such as nanometer, micrometer or millimeter. In geology, displacement is the (quasi-) permanent offset of a geological or man-made reference point along a fault or a *landslide*. In geodesy, movement of a point on the Earth's surface; traditionally defined in a local (east, north, up) coordinate system centered at a reference point fixed to the Earth. In space geodesy, the reference point can be attached to the inertial frame. In *strong-motion* seismology and earthquake engineering, it is the time-dependent position of a material particle during earthquake shaking relative its position at rest, typically obtained by doubly integrating the acceleration records. Recovering the long-period part of the displacement history accurately from velocity or acceleration sensors is difficult because of the sensitivity of translation sensors to rotational ground motions (Trifunac and Todorovska, 2001).

**dome**

See *lava dome*.

**Doppler effect**

A shift in the *frequency* of a *signal* due to relative motion between the signal source and the sensor measuring the signal. The observed frequency is shifted from the source frequency depending on the direction with respect to the moving direction.

**dormant volcano**

By a widely used but inadequate definition, a dormant volcano is one that is not presently erupting but is considered likely to do so in the future. Volcanoes can remain dormant for hundreds, thousands, and, in rare cases, millions of years before reactivating to erupt again. In general, the longer a volcano remains dormant, the greater the likelihood that its next eruption will be large (Simkin and Siebert, 1994).

**double couple**

A force system consisting of two opposing (or orthogonal) couples with the same *scalar moment* and opposite direction (or sign). Its equivalence to a dynamic *fault slip* was first shown by Maruyama (1963) for an *isotropic* elastic medium. The force system equivalent to fault slip for an anisotropic medium was obtained by Burridge and Knopoff (1964). See Aki and Richards (2002, p. 42-48).

**double-difference**

In InSAR (see *satellite interferometry*), this term denotes the difference of two *interferograms*, each of which is the difference of two radar images. In GPS, this term describes a linear combination of four ray paths involving two satellites and two receivers. An equivalent is the so-called double-difference relative precision location algorithm in seismology (Waldhauser and Ellsworth, 2000).

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
|-------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>double frequency microseisms</b> | Describes microseismic noise on the Earth generated by ocean waves with the main spectral peak at half the wave period (hence the seismic noise is at double frequency) as explained by Longuet-Higgins (1950). See Chapter 4 of this Manual, and Webb (2002).                                                                                                                                                                                                                                                                                                                                                                      |
| <b>drift</b>                        | For a structural engineer, “inter-story drift” means cord rotation defined by the difference in horizontal offsets between two elevations (floors) of a given structure. To instrumentalists, it means a change in the sensor offset by time caused by e.g., a fluctuation of temperature, tilting of the base, and instability of mechanical or electronic elements of the instrument. Finally, to a geologist it means a type of glacial deposit.                                                                                                                                                                                 |
| <b>ductile</b>                      | Able to accommodate inelastic <i>strain</i> without loss of strength.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>ductility</b>                    | The property of an object, a structure or a structural component that allows it to continue to have significant strength after it has yielded or begun to fail. Typically, a well-designed <i>ductile</i> structure or component will show, up to a point, increasing strength as its deflection increases beyond yielding or cracking in the cases of reinforced concrete or masonry. See Jennings (2003).                                                                                                                                                                                                                         |
| <b>ductility ratio</b>              | The ratio of the maximum deflection (or rotation) of a structure or structural component to the deflection (or rotation) at first yielding or cracking.                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>duration magnitude</b>           | The logarithm of the duration of a local earthquake <i>seismogram</i> is generally proportional to earthquake <i>magnitude</i> calculated from <i>direct-wave amplitudes</i> . Its physical basis is given by the <i>back-scattering</i> theory of <i>coda waves</i> , after which the duration must be measured from the earthquake origin time. More often it is measured from the <i>onset</i> of <i>P waves</i> , requiring an additional minor, usually neglected correction for the <i>epicentral distance</i> . See Chapter 3 of this Manual, Aki (1987), Lee and Stewart (1981, p. 155-157), Havskov and Ottemöller (2010). |
| <b>dynamic range</b>                | The squared <i>amplitude</i> ratio between the amplitudes of the output and input <i>signal</i> , e.g., of an amplifier, or between the largest and the smallest signal that can be faithfully recorded by a sensor-recording system such as a <i>seismograph</i> or of the range of amplitude variations with respect to a reference amplitude; usually expressed in <i>decibel</i> (see Chapters 4 to 6).                                                                                                                                                                                                                         |
| <b>dynamic ray equations</b>        | The dynamic ray equations govern the properties of a <i>paraxial ray</i> , which is defined as a ray generated by perturbing the source position or take-off direction of a central ray. They are expressed as ordinary differential equations for certain characteristics of a <i>wave field</i> and are related to the amplitude                                                                                                                                                                                                                                                                                                  |

and curvature of wave front in the vicinity of the central ray as a function of the *travel time* or the ray path length. See Chapman (2002) and Červený (2001).

**dynamics (of seismic source)** A term concerned with the *stresses* or forces responsible for generating seismic motion. See *kinematics (of seismic source)*; also Aki and Richards (2002, Box 5.3, p. 129).

**dynamic stress drop** A term in earthquake source physics; the space-time history of *stress drop* while the *fault* is slipping, affected by the detailed processes of *friction* sliding and *seismic wave* radiation. It is more difficult to estimate from observations than *static stress drop*. See Madariaga and Olsen (2002), Ruff (2002, p. 550-551), and Figure 14 in Ben-Zion (2003).

## E

**early warning system (EWS)** Any installation, which can inform with very short time delay public and/or authorities or trigger automatically certain safety or mitigation operations after a potentially catastrophic event. Examples are: Earthquake Early Warning Systems (EEWS), with response times typically within a few seconds after *origin time* (OT) and thus triggering only automatic operations; Tsunami Early Warning Systems (TEWS) with issued warnings or alarms within a few to 10 minutes after OT and lead times before the arriving tsunami waves of minutes to hours, such as the Pacific Tsunami Warning Center (PTWC) in Honolulu or the German Indonesian Tsunami Early Warning System (GITEWS) in Jakarta; Volcano Eruption Early Warning Systems with sometimes even longer lead times. For a review of existing or planned EEWS systems and concepts see Allen et al. (2009) and Gasperini et al (2007) and for TEWS see <http://www.gitews.org> and <http://www.tsunami.gov>.

**Earth** The third planet from the Sun in the solar system. It has a mean radius of 6371 kilometers (km), a surface area of  $5.1 \times 10^8 \text{ km}^2$ , a volume of  $1.08 \times 10^{12} \text{ km}^3$ , and a mass of  $5.98 \times 10^{24}$  kilograms. Its internal structure consists of the *crust*, the *mantle*, and the *core*.

**Earth eigen modes** See Earth *normal modes*.

**Earth normal modes** Comprise the fundamental mode (with lowest frequency) and higher (frequency) modes (or “overtones”) of the *free oscillations* of the Earth as being measurably excited by very strong earthquakes. They comprise *spheroidal* modes and *toroidal* modes. The longest *period* (about 54 minutes) has the spheroidal mode  ${}_0S_2$ , nicknamed “rugby” mode. The toroidal mode  ${}_0T_2$  corresponds to a purely horizontal twisting motion between the northern and southern hemisphere and has a period

|                                   |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|-----------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                   | of about 44 min. See section 2.4 in Chapter 2 of this Manual and <i>normal modes</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>earthquake</b>                 | A shaking of the Earth that is either <i>tectonic</i> or volcanic in origin or caused by collapse of cavities in the Earth. A tectonic earthquake is caused by <i>fault slip</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>earthquake catalog</b>         | See <i>catalog</i> (catalogue) of earthquakes.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>earthquake energy change</b>   | The total change in energy associated with an earthquake. This includes changes in elastic energy, gravitational potential, energy radiated in <i>seismic waves</i> , frictional dissipation (heat), and surface energy of newly created surfaces. See Brune and Thatcher (2002, p. 571-572), and Eq. (6.3) in Ben-Zion (2003). See also Scholz (2002).                                                                                                                                                                                                                                                                                                                                        |
| <b>earthquake engineering</b>     | The field of earthquake engineering is defined as encompassing man's efforts to cope with the harmful effects of earthquakes. It may be subdivided into three parts; (1) a study of those aspects of seismology and geology that are pertinent to the problem, (2) an analysis of the dynamic behavior of structures under the action of earthquakes, and (3) the development and application of appropriate methods of planning, designing and constructing of man-made structures, installations or systems to resist earthquakes. It overlaps with Earth sciences on one hand and with social scientists, civil engineers and planners, and with industry and government on the other hand. |
| <b>earthquake faulting</b>        | The disruption and displacement of rocks along a <i>fault</i> in conjunction with an <i>earthquake</i> rupture. For basic types of earthquake faulting see Figures and Animations in Chapter 3 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>earthquake focal mechanism</b> | A description of the orientation and sense of <i>slip</i> on the causative <i>fault plane</i> derived seismologically from the <i>radiation pattern</i> of <i>seismic waves</i> (traditionally by the sense of the <i>first P-wave motion</i> and <i>polarization</i> of <i>S-wave motion</i> , and now the <i>moment tensor inversion</i> of <i>waveforms</i> ). From the seismic radiation pattern alone it is not possible to distinguish between the fault plane and the auxiliary plane orthogonal to the slip direction. See Sub-chapters 3.4 and 3.5 and Exercises 3.2 and 3.5 of this Manual. See also <i>fault-plane solution</i> and <i>focal mechanism</i> .                        |
| <b>earthquake focus</b>           | See <i>hypocenter</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>earthquake hazard</b>          | Any physical phenomena associated with an earthquake (e.g., <i>ground motion</i> , ground failure, <i>liquefaction</i> and <i>tsunami</i> ) and their effects on land use, man-made structure and socio-economic systems that have the potential to produce a <i>loss</i> . More correct and less ambiguous is the probabilistic definition                                                                                                                                                                                                                                                                                                                                                    |



now generally accepted and used in the hazard community: Earthquake hazard  $H$  is the probability of occurrence of an event with a specified type and level of ground shaking and thus disaster potential within a defined area and interval of time. This definition does not relate to specific effects and loss potential (see *earthquake risk*). They come into play only when  $H$  is convolved with the *exposure*  $E$  to *earthquake hazard* and the *vulnerability*  $V$  of the objects and persons exposed. See also Somerville and Moriwaki (2003), and Giardini et al. (2003).

|                              |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>earthquake intensity</b>  | See <i>seismic intensity</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>earthquake, local</b>     | Although there exists no standard definition for the characterization of earthquakes according to distance, in this Manual (see Chapter 11) an earthquake is considered as local if the <i>direct</i> crustal <i>phases</i> $P_g$ and $S_g$ arrive as first $P$ - and $S$ -wave <i>onsets</i> . The corresponding distance range varies from region to region and ranges between about 100 and 250 km.                                                                                                                                                   |
| <b>earthquake precursor</b>  | Anomalous phenomenon preceding an earthquake. For overviews see Kanamori (2003) and Bormann (2010).                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>earthquake prediction</b> | The statement, in advance of the event, of the time, location, and magnitude of a future earthquake. Earthquake prediction programs have been promoted in Japan, China, the United States, the former Soviet Union, and other countries. See, e.g., Kanamori (2003), and Bormann (2010).                                                                                                                                                                                                                                                                 |
| <b>earthquake, regional</b>  | Usually, the distance range in which phases $P_n$ and $S_n$ appear as first $P$ - and $S$ -wave <i>arrivals</i> on seismic records. The actual range varies from region to region, as in the case of local earthquakes, between about 100 and 1000 km. The distance range between about 1000 and 2000 km is often called far-regional. See <i>earthquake, local</i> , and Chapter 11 of this Manual.                                                                                                                                                     |
| <b>earthquake risk</b>       | <p>Earthquake risk <math>R</math> is the expected (or probable) total loss of life, injury, building and infrastructure damage and their socio-economic consequences in the case of an earthquake and thus, the cumulative product of <math>H</math> with the <i>elements at risk</i> <math>RE</math> and their <i>vulnerability</i> <math>V</math>:</p> $R_i = H_i \sum_{j=1}^m RE_j * V_j .$ <p>However, in common language, <i>earthquake risk</i> and <i>earthquake hazard</i> are occasionally used interchangeably, which might be misleading.</p> |
| <b>earthquake sequence</b>   | A series of earthquakes originating in the same (wider) locality, such as <i>aftershocks</i> , <i>earthquake swarms</i> , but sometimes also a sequence of consecutive <i>mainshocks</i> . See Utsu (2002b).                                                                                                                                                                                                                                                                                                                                             |

- earthquake source parameters** The parameters specified for an earthquake source depend on the applied model. They are *origin time*, *hypocenter location*, *magnitude*, *focal mechanism* and *moment tensor* for a *point source* model. They include *fault geometry*, *rupture velocity*, *rupture duration*, *stress drop*, *slip distribution*, etc. for a finite fault model. See Section 6 in Ben-Zion (2003).
- earthquake source spectrum** In the far-field measured *spectra* of seismic *P* or *S* waves that have been corrected for propagation *path* and source *radiation pattern effects* so as to approximate best the primary spectrum radiated by the *seismic source*. The shape of the source spectrum is mainly characterized by its plateau value, *corner frequency* and high-frequency roll-off, which are controlled by the geometry and size (*seismic moment*), kinematics (*slip* and *rupture velocity*) and dynamics (*stress-drop*, radiated *seismic energy*) of the *earthquake rupture*. Values of these source parameters can therefore, based on reasonable model assumptions, be roughly inferred from the parameters of the source spectrum.
- earthquake storms** A sequence of large earthquakes, occurring over a period of a few years to tens of years over distances of hundreds of kilometers. For example, along the North Anatolian fault between 1939 and 1999; the Mojave region between 1950 and the present, or the eastern Mediterranean between about 340 AD and 380AD. See Nur (2002).
- earthquake stress drop** See *static stress drop* and *dynamic stress drop*, and Figure 14 in Ben-Zion (2003).
- earthquake swarm** A series of earthquakes occurring in a limited area and time period, in which there is not a clearly identified main shock of a much larger magnitude than the rest. See Utsu (2002b) and *swarm (of earthquakes)*.
- earthquake thermodynamics** A branch of physics of *earthquakes* that deals with consequences of the second law of thermodynamics on rock fracture and *earthquake faulting*. See Teisseyre and Majewski (2002).
- earth tides** Periodic *strains* primarily at diurnal and semi-diurnal periods generated in the solid Earth by the gravitational attraction of the Sun and the Moon, firstly observed by Ernst von Rebeur-Paschwitz (1892, 1893).
- edge dislocation** See *dislocation*.
- edge effect (on ground motion in a basin)** A special amplification effect found near the edge of a basin. In a two- or three-dimensional model of basin, basin-induced *diffracted waves* and *surface waves* generated at the edge by an incident *body wave* meet together in phase at

|                                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                          | some point causing constructive interference. As a result the <i>amplitude of ground motion</i> there becomes much larger than in the case of a simple one-dimensional model. The edge effect contributed importantly to the damage concentration in Kobe during the Hyogo-ken Nanbu earthquake of 1995. See also Chapter 14 of this Manual and Kawase (2003).                                                                                                                                                                                   |
| <b>effective normal stress</b>           | The <i>normal stress</i> minus the <i>pore pressure</i> . See Nur and Byerlee (1971), and McGarr et al. (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>effective stress</b>                  | In soil mechanics, the sum of the inter-particle contact forces acting across a plane through a soil element divided by the area of the plane. On a horizontal plane, the effective stress is defined as the total overburden pressure minus the instantaneous pore water pressure. In rock mechanics, it also means the <i>stress</i> above the <i>pore pressure</i> . See Lockner and Beeler (2002), and Youd (2003).                                                                                                                          |
| <b>effusive eruption</b>                 | Non-explosive extrusion of <i>lava</i> at the surface; typically produces <i>lava flows</i> and <i>lava domes</i> .                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>eigenperiod</b>                       | From the German word “Eigenperiode”; see <i>fundamental period</i> and <i>natural period</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>eigenfrequency</b>                    | The inverse of <i>eigenperiod</i> or <i>natural period</i> or <i>fundamental period</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>eikonal</b>                           | From the Greek word “eikon” which means “image”. It was introduced by Heinrich Bruns (1848-1919) to describe the phase function (or <i>travel time</i> ) in wave solutions that can usefully be written $A(x) \exp[i \psi(x)]$ . Constant values of the eikonal represent surfaces of constant phase, or <i>wavefronts</i> . The <i>eikonal equation</i> is a partial differential equation for the eikonal.                                                                                                                                     |
| <b>eikonal equation</b>                  | Consider a <i>wavefront</i> originating from an impulsive source at a point, and define the <i>travel time</i> in space as its <i>time of arrival</i> . The eikonal equation in this case equates the magnitude of <i>gradient</i> vector of the travel time to the reciprocal of the local wave speed. The term was also used in relation to Hamilton’s equations in analytical dynamics, which has some similarity with equations in the <i>ray theory</i> . See Chapman (2002), and the biography of William Rowan Hamilton in Howell (2003). |
| <b>elastic dislocation theory</b>        | In <i>seismology</i> , the theoretical description of how the elastic Earth responds to <i>fault slip</i> , as represented by a distribution of <i>displacement discontinuities</i> .                                                                                                                                                                                                                                                                                                                                                            |
| <b>elastic energy (or strain energy)</b> | The internal energy contained in an elastic material and caused by its <i>translational</i> and/or <i>rotational strain</i> . Elastic energy is a function of materially objective <i>strain tensors</i> (rotating together with the material and independent of its rigid                                                                                                                                                                                                                                                                       |

translations). Knowing this function, one may obtain the *constitutive equations* of the elastic material. In the linear material, elastic energy is a quadratic form of the linear strain tensors (materially objective in the linear approximation). See *principle of material objectivity*.

|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|---------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>elasticity</b>               | The property of a body to recover its shape and size after being deformed.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>elastic rebound model</b>    | A model of the overall earthquake genesis process first put forth by H. F. Reid in 1910. It says that the energy that is released in a <i>tectonic earthquake</i> accumulates slowly in the vicinity of the <i>fault</i> in the form of elastic <i>strain</i> energy in the rocks, until the <i>stresses</i> on the fault exceed its strength and sudden rupture occurs. The rock masses on the two sides of the fault move in a direction to reduce the strain in them. At the time this model was proposed, the source of the accumulating strain was not known.                                                                                                                                                                                                        |
| <b>elastic wave</b>             | A wave that is propagated by some kind of elastic <i>deformation</i> , that is, a deformation that disappears when the deforming forces are removed. A <i>seismic wave</i> is a type of an elastic wave.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>elastodynamic equation</b>   | The basic equation governing the dynamic deformation of a solid body. It is derived from Newton's law on force and <i>acceleration</i> , generalized Hooke's law on <i>stress</i> and <i>strain</i> , and the assumption of the existence of the strain energy function. For an infinitesimally small <i>displacement</i> , it can be reduced to a form equating <i>density</i> times the second time-derivative of the displacement to the sum of the body force and the divergence of the stress tensor (written in terms of the space-derivatives of the displacement). See Aki and Richards (2002, p. 11-36), Teisseyre and Majewski (2002), and Eqs. (1.7) and (1.8) in Ben-Zion (2003). See also the biographies of Isaac Newton and Robert Hooke in Howell (2003). |
| <b>electromagnetic</b>          | Coupled electric and magnetic field effects. For electromagnetic <i>transducers</i> and <i>seismographs</i> see Chapter 5 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>elements exposed to risk</b> | Endangered elements $RE_j$ ( $j = 1, \dots, m$ ) within a defined area, e.g., number of persons, value of property, level of economic activity etc., which constitute the elements of an " <i>exposure model</i> ".                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>elliptic crack</b>           | <i>Fault plane</i> geometry in which the <i>rupture front</i> starts at a point on the fault and spreads as an ellipse. See also Aki and Richards (2002, p. 509-510, 552-560).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>elliptic fault</b>           | See <i>elliptic crack</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |

|                                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>empirical Green's function method</b> | A method for calculating <i>ground motions</i> for a large earthquake using actual records of small earthquakes originating on or near the same <i>fault plane</i> as <i>Green's function</i> representation for the propagation <i>path effects</i> .                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>EMS-98</b>                            | See <i>European Macroseismic Scale</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>en echelon</b>                        | An overlapping or staggered arrangement, in a zone, of geologic features (such as <i>faults</i> ) that are oriented obliquely to the orientation of the zone as a whole. The individual features are short relative to the zone. See <i>en echelon faulting</i> .                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>en echelon faulting</b>               | Faulting type which produces typical <i>en echelon</i> features with repeated oblique <i>fault</i> offsets. Often observed along the surface rupture trace of large <i>strike-slip</i> faults (see <i>faulting</i> ).                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>energy magnitude (Me)</b>             | The radiated <i>seismic energy</i> $E_s$ by an earthquake can nowadays be estimated routinely by integrating over a wide range of <i>periods</i> the squared <i>velocity amplitudes</i> recorded by <i>broadband seismographs</i> and correcting them for wave propagation effects. See Chapter 3 in this Manual and Bormann and Di Giacomo (2011), also for the relationship between $M_e$ and <i>moment magnitude</i> $M_w$ .                                                                                                                                                                                                                        |
| <b>engineering seismology</b>            | That part of seismology which aims primarily at providing seismological data for <i>earthquake engineering</i> , <i>earthquake hazard</i> and <i>earthquake risk</i> applications. Such data are <i>ground-motion</i> parameter data ( <i>intensity</i> , peak values of ground <i>displacement</i> , <i>velocity</i> and <i>acceleration</i> ), <i>time series</i> and <i>spectra</i> of <i>ground motions</i> for <i>response</i> modeling of structures, for assessing spectral <i>site-amplification</i> effects, <i>ground shaking scenario modeling</i> , <i>liquefaction</i> potential, etc. See Chapter 15 of this Manual and Campbell (2003). |
| <b>envelope broadening</b>               | Phenomenon where a wave envelope which has been impulsive at the time of radiation from the source is broadened with increasing travel distance because of <i>diffraction</i> and <i>scattering</i> by distributed <i>heterogeneities</i> . See Sato et al. (2002).                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>epicenter</b>                         | The point on the Earth's surface vertically above the earthquake-rupture initiation point (the <i>focus</i> or <i>hypocenter</i> ).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>epicentral area</b>                   | The area around the epicenter of an earthquake, often characterized by the largest <i>macroseismic intensity</i> $I$ of <i>ground shaking</i> . See <i>epicentral intensity</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>epicentral distance</b>               | Distance from a site (usually a recording <i>seismograph</i> station) to the <i>epicenter</i> of an earthquake. It is commonly given in kilometers for local earthquakes (see <i>earthquake, local</i> ), and in degrees (1 degree is about 111 km) for <i>teleseismic</i> events.                                                                                                                                                                                                                                                                                                                                                                     |

## Glossary

|                                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|---------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>epicentral intensity</b>           | The <i>intensity</i> of <i>macroseismic</i> shaking observed at (or inferred to be) the <i>epicenter</i> of a strong earthquake, usually abbreviated as $I_0$ . See Chapter 12 of this Manual.                                                                                                                                                                                                                                                                                                                                                  |
| <b>episodic tremor and slip (ETS)</b> | The coupled phenomena of <i>non-volcanic tremor</i> and slow volcanic tremor (Rogers and Dragert, 2003). See also <i>non-volcanic tremor</i> , <i>slow slip-events</i> .                                                                                                                                                                                                                                                                                                                                                                        |
| <b>epistemic uncertainty</b>          | The analysis uncertainty due to imperfect knowledge in model parameterization and other limitations of the methods employed. Epistemic uncertainty can be reduced by improvements in the modeling and analysis. See <i>aleatory uncertainty</i> and Anderson (2003).                                                                                                                                                                                                                                                                            |
| <b>equivalent linearization</b>       | The process of selecting a linear system and its parameters so that it approximates the response of a nonlinear system.                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>erg</b>                            | A cgs unit for energy; $1 \text{ erg} = 10^{-7} \text{ joule}$ . Now an obsolete, non-standard unit but commonly used in older seismological literature. See <i>Joule</i> .                                                                                                                                                                                                                                                                                                                                                                     |
| <b>eruption</b>                       | A general term applied loosely to any sudden expulsion — explosive or non-explosive — of material (solid, liquid, or gas) from <i>volcanic vents</i> . Some volcanologists, however, prefer that the term be restricted only to ejections of <i>lava</i> and (or) fragmental solid debris, thereby distinguishing them from aqueous or gaseous emissions devoid of particulate matter. Such a distinction is important in the assessment of volcano hazard and in the communications of volcano-hazard information during a volcanic emergency. |
| <b>eruption column</b>                | A vertical plume of <i>tephra</i> and gases mixed with air that forms above a <i>volcanic vent</i> during an explosive volcanic <i>eruption</i> ; may rise to stratospheric heights (tens of kilometers) and can last for many hours.                                                                                                                                                                                                                                                                                                           |
| <b>ESC</b>                            | See <i>European Seismological Commission</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>Euler pole</b>                     | See <i>Euler vector</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>Euler vector</b>                   | Rotation vector describing the relative motion between two <i>tectonic plates</i> along the Earth's surface. The magnitude of the Euler vector is the rotation rate, and its intersection with the Earth's surface is the Euler pole. The linear velocity at any point on the <i>plate boundary</i> is the vector product of the Euler vector and the radius vector to that point.                                                                                                                                                              |
| <b>European Macroseismic Scale</b>    | A twelve-degree <i>macroseismic intensity</i> scale with the acronym EMS. In 1988 the European Seismological Commission initiated a major revision and elaboration of the <i>MSK Scale</i> , which resulted finally in the publication of the                                                                                                                                                                                                                                                                                                   |

*European Macroseismic Scale 98 (EMS-98; Grünthal, 1998).* Although this new scale is largely based and compatible with the *MSK Scale*, its organization is very different. Most importantly, its guidelines for practical use, supported by illustrations, tables, vulnerability classification of building types and worked examples, are very elaborate and unique in comparison with all other widely used *macroseismic intensity* scales. Since its publication the EMS-98 has been widely adopted inside and also outside Europe.

**European Seismological Commission (ESC)** A regional commission of *IASPEI*. The ESC was founded after World War II and had its first General Assembly in 1952, has more than 30 member states in the region covering an area from the Mid-Atlantic Ridge to the Urals and the Arctic Ocean to northern Africa. It holds ever since its foundation General Assembly meetings every second year. See <http://www.esc-web.org>.

**event association** See *signal association*.

**event, seismic** See *seismic event*.

**event horizon** A term in *paleoseismology* indicating the ground surface at the time of an earthquake in the geologic past. See Grant (2002).

**exceedance probability** Probability that a specified value of a *strong-motion* parameter, e.g., the *intensity* of shaking or the *peak ground acceleration*, is exceeded by a future earthquake occurrence within a specified period of time. See Campbell (2003).

**exploration seismology** Application of seismological methods by using mostly artificial *seismic sources* (e.g., explosions, vibrators) and thus waves of higher frequencies within shorter distance ranges for a more detailed imaging of structures in the upper Earth's *crust* with exploratory interest. See Sheriff and Geldart (1995) and *seismic exploration*.

**explosion earthquake** A term in volcanology; *seismic events* associated with explosive volcanic activities. In addition to the usual *seismic waves* such as *P* and *S* waves, they often also generate acoustic waves travelling through the air (see *infrasound*) and are transmitted back into the ground in the vicinity of the seismometer. See Ben-Menahem and Singh (1981) and McNutt (2002).

**exposure model** A more or less quantified model of the number and distribution of persons and objects of different degree of *vulnerability* in a given area which are exposed to hazard (such as *earthquake hazard*).

**exposure time** The time period used in the specification of probabilistic *seismic hazard* and *seismic risk*. It is usually chosen to represent the design or economic life of a structure.

**extensional tectonic setting** A region in which the *crust* is undergoing lateral expansion and for which the maximum *principal stress* is vertical. See McGarr et al. (2002).

**extinct volcano** By a widely used, but inadequate definition, an extinct volcano is one that is not erupting and is not expected to do so in the future (see Decker and Decker, 1998).

## F

**failure envelope** Boundary of *stress* conditions associated with rock failure, as defined in *shear stress* versus *normal stress* space. Typically the slope of the failure envelope  $d\sigma/d\sigma_n$  is steepest at low normal stress ( $\sigma_n$ ). The slope is also referred to as the “pressure dependence” of failure strength. See Lockner and Beeler (2002, p. 508).

**fallout** The settling and deposition of *tephra* from the atmospheric plume of an explosive volcano *eruption*.

**far field** See *near field* and *far field*.

**Fast Fourier transform (FFT)** A computer algorithm that transforms a time domain sample sequence to a frequency-domain sequence which describes the spectral content of the *signal*. This efficient algorithm for computing the Fourier transform was first introduced by Cooley and Tukey (1965). See Brigham (1988), Hutt et al. (2002), and Scherbaum (2002).

**fault** A fracture or fracture zone in the Earth along which the two sides have been displaced relative to one another parallel to the fracture. The accumulated *displacement* across a fault may range from a fraction of a meter to many tens or hundreds of kilometers. The type of fault is specified according to the orientation and sense of *slip* and the inclination (*dip*) of the *fault plane* (Aki and Richards, 1980; Yeats et al., 1997). If the block opposite an observer looking across the fault moves to the right, the slip style is termed *right lateral (dextral)*, if the block moves to the left, the motion is termed *left lateral (sinistral)*. *Dip-slip* faults are inclined fractures along which rock masses have mostly shifted vertically. If the rock mass above an inclined fault moves down (due to lateral extension) the fault is termed *normal*, whereas if the rock above the fault moves up (due to lateral compression), the fault is termed *reverse* (or *thrust*). *Oblique-slip* faults have significant components of both slip styles (i.e., *strike slip* and *thrust* or *dip slip*). See Figs. 3.105 and



|                                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|---------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                       | 3.106 and EX 3.2 in this Manual, and for more specific explanations and definitions Aki and Richards (1980, 2002, 2009) and the entries <i>normal fault</i> , <i>reverse fault</i> , <i>thrust fault</i> and <i>strike-slip fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>fault, active:</b>                 | See <i>active fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>fault breccia</b>                  | A fault-rock made up of angular rock fragments derived from <i>brittle</i> fragmentation in a fine-grained or hydrothermal matrix. See Sibson (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>fault creep</b>                    | Quasi-static <i>slip</i> on a <i>fault</i> ; slip that occurs so slowly that radiation of <i>seismic waves</i> is negligible. See <i>creep</i> and <i>creep events</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>fault dip</b>                      | The inclination of a <i>fault plane</i> relative to the horizontal in degrees. A dip of 0 degree is horizontal, 90 degrees is vertical. See Figs. 3.104 to 3.107 and 3.111 in Chapter 3 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>fault-fracture mesh</b>            | A mesh structure of interlinked minor <i>faults</i> (shear fractures) and extension fractures occupying a substantial volume of the rock-mass. See Sibson (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>fault gouge</b>                    | A clayey, soft material formed when the rocks in a <i>fault zone</i> are pulverized during slippage. Textures range from essentially random-fabric to foliated.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>fault length</b>                   | See <i>rectangular fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>fault model (of an earthquake)</b> | A model of an earthquake source in which a slip across a <i>fault plane</i> causes the <i>ground motion</i> . See Aki (2002), Madariaga and Olsen (2002), Das and Kostrov (1988), and Scholz (2002). See also <i>Brune model</i> , <i>Burridge and Knopoff model</i> , <i>Haskell model</i> , <i>Knopoff model</i> , <i>Kostrov model</i> , <i>omega-squared model</i> and <i>specific barrier model</i> .                                                                                                                                                                                                                                                                                                                                    |
| <b>fault plane</b>                    | The surface along which there is <i>slip</i> during an earthquake. In simplified mathematical approximations and <i>fault-plane solutions</i> assumed to be planar (flat).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>fault-plane solution</b>           | A simplified geometric-mathematical way of presenting an earthquake <i>fault</i> and the direction of <i>slip</i> on it, using circles with two orthogonally intersecting curves for the two potential <i>fault planes</i> that look like “beach balls” (see Figs. 3.105 and 3.106 of Chapter 3 in this Manual). A fault-plane solution is found by an analysis using stereographic projection or its mathematical equivalent to determine the attitude of the causative fault and its direction of slip from the <i>radiation pattern</i> of <i>seismic waves</i> using earthquake records at many stations. The most common analysis uses the direction of <i>first motion</i> of <i>P-wave</i> onsets and yields two possible orientations |

for the fault rupture and the direction of seismic slip. Other techniques use in addition the *polarization* of teleseismic S waves and/or *amplitude* ratios between P and S waves. From these data, approximate inferences can be made concerning the principal axes of stress in the region of the earthquake. The *principal stress* axes determined by this method are the *compressional* axis (also called the P-axis, i.e., the axis of greatest compression, or  $s_1$ ), the *tensional* axis (also known as the T-axis, i.e., the axis of least compression, or  $s_3$ ), and the intermediate axis ( $s_2$ ). See also *focal mechanism* and *earthquake focal mechanism*.

|                        |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>fault rake</b>      | A parameter describing the <i>slip</i> direction on a <i>fault</i> , measured as an angle in the <i>fault plane</i> between the <i>fault strike</i> and the slip vector. Slip in the direction of the maximum slope (i.e., the <i>dip</i> ) of the fault plane is a rake of 90°, positive for <i>reverse faults</i> , negative for <i>normal faults</i> . See Fig. 3.111 in Chapter 3 of this Manual; also Aki and Richards (2002, Fig. 4.13, p. 101), and Figure 13 in Ben-Zion (2003). |
| <b>fault-rock</b>      | A rock whose textural/micro-structural characteristics (usually involving grain-size reduction from the <i>protolith</i> ) developed through deformation within a <i>fault zone</i> at depth. See Sibson (2002).                                                                                                                                                                                                                                                                         |
| <b>fault sag</b>       | A narrow <i>tectonic</i> depression common in <i>strike-slip fault zones</i> . Fault sags are generally closed depressions in the order of a hundred meters wide and approximately parallel to the fault zone.                                                                                                                                                                                                                                                                           |
| <b>fault scarp</b>     | Step-like linear landform coincident with a <i>fault trace</i> and caused by geologically recent <i>slip</i> on the fault.                                                                                                                                                                                                                                                                                                                                                               |
| <b>fault slip</b>      | The relative <i>displacement</i> of points on opposite sides of a <i>fault</i> , measured on the fault surface. See Figs. 3.100 to 3.102, 3.107, and 3.111 of Chapter 3 in this Manual; also Aki and Richards (2002, Fig. 4.13, p. 101), and Figure 13 in Ben-Zion (2003).                                                                                                                                                                                                               |
| <b>fault slip rate</b> | The rate of <i>displacement</i> on a <i>fault</i> averaged over a time period encompassing several earthquakes.                                                                                                                                                                                                                                                                                                                                                                          |
| <b>fault strike</b>    | The azimuthal direction of the horizontal in the <i>fault plane</i> . See Figs. 3.30 and 3.31 in Chapter 3 of this Manual; also Aki and Richards (2002, Fig. 4.13, p. 101), and Figure 13 in Ben-Zion (2003).                                                                                                                                                                                                                                                                            |
| <b>fault trace</b>     | The intersection of a <i>fault</i> with the Earth's surface or with a horizontal plane; indirectly also the line commonly plotted on geologic maps to represent the surface exposure of a fault.                                                                                                                                                                                                                                                                                         |
| <b>fault, unlocked</b> | See <i>creep</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |

|                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|--------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>fault width</b>             | See <i>rectangular fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>fault zone</b>              | The <i>fault plane</i> of the geometric-mathematical model of an earthquake is usually a planar surface in an elastic medium. The actual fault is a zone of finite width, comprising also secondary fault branches (see Fig. 3.8 in Chapter 3 of this Manual), in which non-elastic processes occur during the earthquake, sometimes also called a break-down zone or cohesive zone. The fault zone includes the <i>fault gouge</i> .                                  |
| <b>fault-zone head wave</b>    | A seismic <i>P</i> or <i>S</i> wave propagating along a material interface in a <i>fault zone</i> structure with the <i>velocity</i> and motion polarity of the corresponding <i>body waves</i> on the faster side of the fault. See <i>head waves</i> .                                                                                                                                                                                                               |
| <b>fault-zone trapped mode</b> | Seismic <i>guided waves</i> trapped in the low-velocity <i>fault zone</i> . They are useful for studying the three-dimensional structure of the fault zone. They have been also used to find temporal changes in the physical characteristics of fault zones.                                                                                                                                                                                                          |
| <b>feedback</b>                | The addition of part of the output signal of a linear system like an amplifier to the input signal in order to modify the response. Mostly used to stabilize the system (negative feedback). See Chapter 5 of this Manual.                                                                                                                                                                                                                                             |
| <b>felt area</b>               | The area of the Earth's surface over which the effects of an earthquake were humanly perceptible. See Chapter 12 of this Manual.                                                                                                                                                                                                                                                                                                                                       |
| <b>filter</b>                  | Electronic device or numerical algorithm which acts on an input signal to produce an output signal with a desired modification in spectral characteristics. See <i>digital filter</i> ; also Chapters 5 and 6 as well as IS 5.2 in this Manual, and Scherbaum (2002).                                                                                                                                                                                                  |
| <b>filtering</b>               | <i>Attenuation</i> or downgrading of certain frequency components of a ( <i>seismic</i> ) <i>signal</i> as compared to others. For a recorded signal, the process can be accomplished mechanically, electronically or numerically in a computer. Filtering also occurs naturally as <i>seismic waves</i> are attenuated in a frequency-dependent way during their passage through the Earth.                                                                           |
| <b>final prediction error</b>  | Criterion proposed by Akaike (1970) for selection of the order of an <i>autoregressive process</i> . The final prediction error (FPE) is defined as: $FPE = \frac{n + p + 1}{n - p - 1} E_s^2$ where $n$ is the number of observations, $p$ the corresponding chosen model order, and $E_s^2$ the sum of squares of prediction errors. FPE and the <i>Akaike Information Criterion</i> (AIC) are related, as, with regard to <i>autoregressive models</i> , a model is |

obtained that approximately minimizes the FPE by minimizing the AIC. Another method for selection of the order of an *autoregressive process* is the use of the *partial autocorrelation function* (PACF).

**finite element**

A geometrical subdivision, generally small, of a larger structure or medium. Finite elements are commonly used in numerical studies of the response of complex media or structures to earthquake excitation. Typical finite elements include triangles and rectangles for analysis of two-dimensional problems, and tetrahedrons for three-dimensional problems. In applications, a structure is divided into many finite elements and field quantities such as *density* and *displacement*, for example, are assumed to vary over each element in a prescribed manner (e.g., linearly varying displacement). Thus the displacements and associated *stresses* and *strains* within each element are determined by the displacements at its corners or nodes. The displacements of the nodes of the finite elements then become the unknown vector in a matrix equation that is solved numerically. See Hall (2003).

**FIR filter**

Finite Impulse Response filter, also named *acausal*- or, in their symmetrical realization, *zero-phase filter*.

**first motion**

The first half-cycle of a *body wave*, usually the *direct P wave*. On a *seismogram*, the first discernible displacement of the record trace caused by the *arrival* of a P wave at the *seismometer*. The direction (*polarity*) of first motion, which for the P wave is either away from source (called compression) or towards it (called dilatation), is assumed to be conserved along the ray path from the focus to a given station, given that corrections have been made for any polarity changes due to *reflection* of the wave at any boundary between source and receiver. First motions of P arrivals are often used for determining the *earthquake focal mechanism*. However, when the *seismic signal* arrives in the presence of *seismic noise* the proper polarity of the first motion may be distorted and difficult to recognize.

**fissure eruption**

A volcanic eruption fed by one or more fissure vents, rather than by a central (point-source) vent or a tight cluster of central vents. See *volcanic vent*.

**f-k analysis**

*Frequency* (f) versus *wavenumber* (k) analysis is a key *array processing* technique that maps the power of seismic waves observed at an *array* as a function of the wave propagation *azimuth* (or *backazimuth*) and the horizontal *slowness* component for a given frequency band. F-k analysis is also used to identify *seismic phases* on the basis of their distance-dependent *slowness* differences and to locate seismic events.

|                           |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|---------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                           | See <i>wavenumber filtering</i> , <i>wavenumber vector</i> , <i>slowness</i> , <i>array beam</i> , and <i>slowness vector</i> and Chapter 9 in this Manual.                                                                                                                                                                                                                                                                                                                                                                     |
| <b>flank eruption</b>     | An eruption that takes place away from the summit area of a volcano (e.g., on rift zones or flanks of the volcano); synonymous with lateral eruption.                                                                                                                                                                                                                                                                                                                                                                           |
| <b>fling step</b>         | The permanent <i>displacement</i> of the ground at or near a ruptured <i>fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>flow deformation</b>   | Shear <i>deformations</i> in soil that occur with little resistance within a liquefied layer or element. Deformations may be of limited excursion, as occurs in <i>dilative</i> soils, or unlimited excursion, as occurs in <i>contractive</i> soils. See Youd (2003).                                                                                                                                                                                                                                                          |
| <b>flow failure</b>       | <i>Liquefaction</i> -induced failure of a slope or embankment underlain by contractive soils leading to large, rapid ground <i>displacements</i> . These failures are characterized by large loss of shear strength, large lateral displacements (several meters or more), and severe disturbance to the liquefied and overlying soil layers. See Youd (2003).                                                                                                                                                                  |
| <b>fluid overpressure</b> | Fluid-pressure exceeding the <i>hydrostatic pressure</i> value appropriate to the depth.                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>fmax</b>               | Above a limiting <i>frequency</i> , $f_{\max}$ , the observed spectrum of a typical local earthquake decays more steeply with increasing frequency than is expected from the standard <i>omega-squared model</i> . This was first recognized by Hanks (1982) and has been attributed to the earthquake <i>source effect</i> (e.g., Kinoshita, 1992), as well as to attenuation of the shaking by unconsolidated sediments underlying the recording site (e.g., Anderson and Hough, 1984). See also <i>omega-squared model</i> . |
| <b>focal depth</b>        | The conceptual “depth” of an <i>earthquake focus</i> . If determined from the <i>first motion arrival-time</i> data, this represents the depth of rupture initiation (the <i>hypocenter depth</i> ). If determined from waveforms that are long compared to the <i>fault</i> dimension, the depth represents some weighted average of the moment release in the earthquake (the “centroid” depth. See Romanowicz (2002, p. 153), and Lee and Stewart (1981, p. 130-139).                                                        |
| <b>focal mechanism</b>    | See <i>fault-plane solution</i> and <i>earthquake focal mechanism</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>focal sphere</b>       | Hypothetical (conceptual) sphere around a seismic <i>point source</i> (i.e., a source with linear dimensions smaller than the diameter of the focal sphere). It is used for calculating the <i>take-off angles</i> under which the <i>seismic rays</i> leave the source through the surface of this sphere towards the seismic stations and to reconstruct the <i>fault-plane solution</i> (i.e., the type and spatial orientation of the earthquake rupture) from the stereographic                                            |

projection of these ray “penetration points” through the focal sphere and their associated *first-motion polarities*, *polarisation* and *amplitude* ratios.

|                                                      |                                                                                                                                                                                                                                                                                                                                                                                                      |
|------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>focal zone</b>                                    | The rupture zone of an earthquake. In the case of a <i>great earthquake</i> , the focal zone may extend several hundreds of kilometers in length and several tens of kilometers in width.                                                                                                                                                                                                            |
| <b>focus</b>                                         | That point within the Earth from which the first <i>displacement</i> of an earthquake and radiation of its <i>elastic waves</i> originates. See <i>hypocenter</i> .                                                                                                                                                                                                                                  |
| <b>focusing effect (in local site amplification)</b> | Special amplification of <i>ground motion</i> found just above the corner of the basin bottom or layer interfaces. In a two- or three-dimensional basin with strong lateral variation, rays of an incident <i>body wave</i> are warped at these interfaces and constructive (or destructive) interference takes place at some point on the surface. See Chapter 14 of this Manual and Kawase (2003). |
| <b>footwall</b>                                      | The underlying side of a non-vertical <i>fault</i> surface. See <i>hanging wall</i> , Figure 3.107 in Chapter 3 of this Manual, Grant (2002), and Figure 13 in Ben-Zion (2003).                                                                                                                                                                                                                      |
| <b>force-balance accelerometer</b>                   | A <i>seismometer</i> that compensates the inertial force acting on its suspended mass with a negative <i>feedback</i> force, and thereby keeps the mass at rest relative to the ground. Modern broadband seismometers make use of this feedback system. See Chapter 5 in this Manual.                                                                                                                |
| <b>fore arc</b>                                      | The region between the <i>subduction thrust fault</i> and the <i>volcanic arc</i> .                                                                                                                                                                                                                                                                                                                  |
| <b>forensic seismology</b>                           | Seismology applied to problems involving legal issues, such as, e.g., the verification of compliance with any treaty limiting the testing of nuclear explosions. See Chapter 17 in this Manual and Richards (2002).                                                                                                                                                                                  |
| <b>foreshock</b>                                     | Relatively smaller earthquake that precedes the largest earthquake (which is termed the <i>mainshock</i> ) in an <i>earthquake sequence</i> . Foreshocks usually originate at or near the <i>focus</i> of the mainshock and may precede the latter by seconds to weeks. Not all mainshocks have foreshocks. See also <i>aftershock</i> and Utsu (2002b).                                             |
| <b>forward problem</b>                               | A typical problem in physical science is that of the modeling of natural phenomena. Through such a modeling, and given some parameters describing the system under study, one may predict the outcome of some possible observables. Solving a forward (also called “direct”) problem requires model parameters → model → prediction of observed data. See <i>inverse problem</i> and                 |

Fig. 1.1 of Chapter 1 in this Manual; also Mosegaard and Tarantola (2002), and Menke (1984).

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|-------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>forward scattering</b>           | See <i>scattering</i> and <i>scattering theory</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>Fourier amplitude spectrum</b>   | See <i>Fourier transform (spectrum)</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>Fourier phase spectrum</b>       | See <i>Fourier transform (spectrum)</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>Fourier transform (spectrum)</b> | An integral transform mapping <i>signals</i> defined in the time domain $f(t)$ into the complex frequency domain $F(\omega)$ by multiplying $\exp(i\omega t)$ and integrating over $t$ from minus infinity to plus infinity. The Fourier spectrum consists of two complementary parts, the <i>amplitude-frequency</i> and the <i>phase-frequency spectrum</i> . See equation (4.2) in Chapter 4 of this Manual as well as Scherbaum (2002), and the biography of Jean Baptiste Joseph Fourier in Howell (2003).                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>Fourier analysis</b>             | The mathematical operation that resolves a time series (for example, a recording of <i>ground motion</i> ) into a series of numbers that characterize the relative <i>amplitude</i> and <i>phase</i> components of the <i>signal</i> as a function of <i>frequency</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>fractal</b>                      | A geometrical structure or object which is self-similar on all scales (i.e., <i>scale invariance</i> ), and is quantified by a fractional (instead of an integer) dimension. See Mandelbrot and Frame (2002) for a general review of fractals.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>fractal dimension</b>            | A statistical quantity in <i>fractal</i> geometry that indicates how completely a fractal appears to fill space, as one zooms down to smaller and smaller scales. Although there exist many specific definitions, the fractal dimension may be estimated from the exponent of a power law distribution such as the <i>Gutenberg-Richter relation</i> . Its <i>b-value</i> is about half of the fractal dimension of the considered complete random ensemble of earthquakes of different <i>magnitude</i> . Values of $b = 1.5$ , $1$ or $0.5$ , respectively, would indicate that the governing earthquake rupture process would be 3D, 2D or 1D (linear). <i>b-values</i> may vary with time and tend to be larger for <i>aftershocks</i> and <i>swarm</i> earthquakes than for <i>mainshocks</i> of larger magnitudes. Values $0.7 < b < 1.4$ have been reported by Neunhöfer and Güth (1989) within a major earthquake swarm. |
| <b>fractal statistics</b>           | A statistical distribution in which the number of objects has a power-law dependence on their size. See Turcotte and Malamud (2002), and Vere-Jones and Ogata (2003). See also <i>fractal</i> and <i>fractal dimension</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>fracture</b>                     | A general term for <i>discontinuities</i> in rock; includes <i>faults</i> , <i>joints</i> , and other breaks. See Section 4 in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |

## Glossary

|                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|--------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>fracture toughness</b>      | A measure of the energy expended in an increment of <i>fracture</i> . It is more difficult to propagate a fracture in a material with high fracture toughness than with low fracture toughness. See Teisseyre and Majewski (2002), and Lockner and Beeler (2002).                                                                                                                                                                                                                                                                                                                         |
| <b>fracture zone</b>           | An elongated zone of unusually irregular topography that separates regions of different water depths in the ocean floor. It is commonly found to displace the magnetic lineations and the axes of mid-ocean ridges, and it is interpreted as a <i>transform fault</i> . See Wilson (1965), and Uyeda (1978).                                                                                                                                                                                                                                                                              |
| <b>fragmentation</b>           | <i>Deformations</i> and failures in which the structures in the material undergo micro-crushing or crushing under high compressive load.                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>free-field motion</b>       | Strong <i>ground motion</i> that is not modified by the earthquake caused motions of nearby buildings or other structures or geologic features. In analyses, often defined as the motion that would occur at the interface of the structure and the foundation if the structure were not present. A fully instrumented building site typically includes one or more <i>accelerographs</i> located some distance from the structure to obtain a better approximation of the free-field motion. See Jennings (2003).                                                                        |
| <b>free oscillations</b>       | See <i>normal modes</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>frequency</b>               | The frequency is the number of cycles of a periodic motion per unit time, or per unit <i>period</i> (as the reciprocal time). In the context of the <i>Fourier transform</i> , “frequency” is also used for the angular frequency, which is $2\pi$ times the frequency defined above. The frequency unit is Hertz (Hz).                                                                                                                                                                                                                                                                   |
| <b>frequency domain</b>        | A <i>seismic signal</i> that has been recorded in the <i>time domain</i> (as a <i>seismogram</i> ) can be decomposed by means of <i>Fourier analysis</i> into its <i>amplitude</i> and <i>phase</i> components as a function of frequency (see <i>spectrum</i> ). The representations of a seismic signal in the <i>time domain</i> and in the frequency domain are mathematically equivalent. For some procedures of data analysis the time-domain representation of seismic records is more suitable, while for others the frequency-domain approach is more appropriate and efficient. |
| <b>frequency of occurrence</b> | The number of times something happens in a certain <i>period</i> of time, e.g., the (average) number of <i>earthquakes</i> in a given <i>magnitude</i> range per year or per century, also termed <i>occurrence rate</i> .                                                                                                                                                                                                                                                                                                                                                                |
| <b>frequency response</b>      | <i>Fourier transform</i> of the output signal of a linear system divided by the Fourier transform of the input signal. The frequency response of a linear system equals the Fourier                                                                                                                                                                                                                                                                                                                                                                                                       |



transform of its *impulse response*. For a typical inertial *seismometer* (e.g., an *accelerometer* or *rotational-motion sensor*), the *frequency response* is the complex *transfer function* relating the input ground motion (translational acceleration or *rotational velocity*) to the output signal (usually voltage). This transfer function is typically measured in terms of *amplitude* and *phase spectra*. For analog-output sensors, the measured frequency response is generally modeled in the Laplace domain with a best-fit pole-zero model (Nigbor et al., 2009).

**frequency-wavenumber analysis** See *f-k analysis*.

**friction** The ratio of shear *stress* to *normal stress* on a planar *discontinuity* such as *fault* or joint surface. See Douglas (2002, p. 506), and section 5 in Ben-Zion (2003).

**frictional instability** A condition that arises during quasi-static *fault slip* when the reduction of fault strength with *displacement* exceeds the elastic stiffness of the region surrounding the fault; under these conditions the *slip rate* accelerates rapidly. See Lockner and Beeler (2002, p. 517), and material below Eq. (5.4f) in Ben-Zion (2003).

**frictional (FR) regime** That portion of the *crust* or *lithosphere* where *fault* motion is dominated by unstable (velocity weakening) pressure-sensitive frictional sliding. See Sibson (2002).

**friction-melting** Melting as a consequence of high temperature generated by *friction* during rapid localized *fault slip*. See Sibson (2002).

**frozen waves** In the *epicentral area* of a *great earthquake*, walls, embankments, and the like are sometimes left in the form of a wave. These “frozen waves” are attributed to the ground cracking open at the crests of the waves, sometimes with the emission of sand and water. Frozen waves are also seen on the surface of the Moon, concentric with very large impact craters.

**fully-decoupled explosion** An underground explosion, carried out within a cavity large enough that the walls of the cavity are not stressed beyond their elastic limit. See Richards (2002).

**fundamental period** The longest *period* for which an object, e.g., a *seismometer*, a structure, the sub-surface underground or the whole planet Earth shows a maximum response. Also termed *eigenperiod* or *natural period*. The reciprocal is the *natural frequency* or *eigenfrequency*.

## G

|                                                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|------------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>g</b>                                             | A commonly used unit for the <i>acceleration</i> of a free falling mass due to the <i>gravity</i> of the Earth (9.81... m/s <sup>2</sup> ). When there is an <i>earthquake</i> , the accelerations caused by the shaking can be measured as a fraction or percentage of gravity (%g).                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Ga</b>                                            | An abbreviation for billion (10 <sup>9</sup> ) years ago.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>gain</b>                                          | Commonly used to describe an amplifier or <i>filter</i> gain, not the sensitivity of a sensor.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>gal</b>                                           | A unit of <i>acceleration</i> (1 gal = 1 cm/s <sup>2</sup> , or approximately one thousandth of 1 g).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Gaussian distribution (of measurement values)</b> | According to Carl Friedrich Gauss (1777-1855), a symmetric, bell-shaped distribution of randomly sampled measurement values about their mean value due to random measurement errors/fluctuations with a specified <i>standard deviation</i> SD. About 67% of the values deviate less than $\pm 1 \times \text{SD}$ from the mean value, and about 95% less than $2 \times \text{SD}$ .                                                                                                                                                                                                                                                                                                                                               |
| <b>Gaussian noise spectrum</b>                       | The spectrum of a time history whose sample values are generated by random selection from a statistical population that has a specified mean and a standard deviation. Accordingly, the ordinate values follow a <i>Gaussian distribution</i> about their mean. In earthquake studies, this type of spectrum is commonly multiplied by a theoretical <i>earthquake source spectrum</i> to obtain predicted <i>ground-motion spectra</i> for hypothetical earthquakes.                                                                                                                                                                                                                                                                |
| <b>Geiger's method</b>                               | A commonly used method to determine the <i>hypocenter</i> and the <i>origin time</i> of an earthquake or any other <i>seismic source</i> by using <i>arrival times</i> of <i>seismic waves</i> , assuming the knowledge of their <i>travel-times</i> within the Earth. Geiger (1910) applied the Gauss-Newton method (a linearized <i>least-squares fit</i> algorithm) to solve the earthquake location problem, but his method was not practical until the advance of computers in the late 1950s. See <i>linearize</i> , IS 11.1 in this Manual, Lahr (2003), Klein (2003), Lee et al. (2003b), Lee and Stewart (1981, p. 132-139), and the biographies of Ludwig Geiger, Karl Friedrich Gauss, and Isaac Newton in Howell (2003). |
| <b>genetic algorithm (GA)</b>                        | A learning algorithm used to maximize the adaptability of a system to its environment. It is often used as a computer algorithm in function optimization or object selection problems. See McClean (2002) and Sambridge (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>geochronology</b>                                 | Science of dating earth materials, geologic surfaces and processes. See Grant (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |

|                                   |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
|-----------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>geodesy</b>                    | The science of determining the size and shape of the Earth and the precise location of points on its surface as well as of the dynamical processes and forces leading to the observed shape.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>geodetic</b>                   | Referring to the determination of the size and shape of the Earth and the precise location of points on its surface.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>geodetic moment</b>            | Geodetic estimate of the <i>seismic moment</i> . See Feigl (2002, p. 616-617).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>geodynamics</b>                | Study of the dynamics of the Earth's interior including heat transfer and convection currents.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>geodynamo</b>                  | Magneto-hydrodynamic convection in the Earth's <i>core</i> that is responsible for the generation of the Earth's main magnetic field. See Song (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>geoid</b>                      | Surface of constant potential energy, representing the balance between gravitational and rotational forces; corresponds to an idealization of the Earth's actual surface. In practice, deviations of the geoid from a reference ellipsoid are presented as the geoid.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>geology</b>                    | The study of the planet Earth - the materials it is made of, the processes that act on those materials, the products formed, and the history of the planet and its life forms since its origin.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>geologic time</b>              | The time since the Earth was formed 4.6 billion years ago (Ga), and encompasses the entire geologic history of the Earth. It is usually divided chronologically into the following hierarchies of time units: eon, era, period and epoch. The eons are: (1) <i>Hadean</i> (4.6 – 4 Ga), (2) <i>Archaean</i> (4-2.5 Ga), (3) <i>Proterozoic</i> (2.5- 0.545 Ga), and (4) <i>Phanerozoic</i> (545 million years ago to the present). The Phanerozoic eon is divided into the following eras: (1) <i>Paleozoic</i> (545-245 Ma), (2) <i>Mesozoic</i> (2.45-66.4 Ma), and (3) <i>Cenozoic</i> (66.4-0 Ma). The Cenozoic era is divided into following periods: (1) <i>Tertiary</i> (66.4-1.6 Ma), and (2) <i>Quaternary</i> (1.6 Ma-present). The Quaternary period is divided into the following epochs: (1) <i>Pleistocene</i> (1.6 Ma – 10,000 years ago), and (2) <i>Holocene</i> (10,000 years ago to the present). For a detailed discussion of geologic time, see Eicher (2002), and Hancock and Skinner (2000). |
| <b>geomagnetic polarity epoch</b> | A period of <i>geologic time</i> during which the Earth's magnetic field was essentially or entirely of one polarity. Short period of opposite polarity within a polarity epoch is called "event".                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>geomagnetic pole</b>           | The point where the axis of the calculated best-fit dipole cuts the Earth's surface in the northern or southern hemisphere.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |

|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|---------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                 | These poles lie opposite to each other and wander with time. See <i>magnetic poles (of the Earth)</i> .                                                                                                                                                                                                                                                                                                                                                                   |
| <b>geomagnetic reversal</b>     | A change of the Earth's magnetic field between normal polarity and reversed polarity. These reversals occur at different time intervals.                                                                                                                                                                                                                                                                                                                                  |
| <b>geomagnetism</b>             | The study of the properties, history, and origin of the Earth's magnetic field. See Merrill and McFadden (2002).                                                                                                                                                                                                                                                                                                                                                          |
| <b>geometrical spreading</b>    | Decrease of wave energy per unit area of <i>wavefront</i> , and thus of the <i>amplitudes</i> of waves, due to the expansion of the wavefront with increasing distance.                                                                                                                                                                                                                                                                                                   |
| <b>geomorphology</b>            | The study of the character and origin of landforms, such as mountains, valleys, etc.                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>geophone</b>                 | A term commonly used for a relatively inexpensive, robust, light weight and small size <i>seismometer</i> , usually a mass-spring system with electrodynamic <i>transducer</i> , that responds to high-frequency (usually > 1 Hz) ground motions (see Fig. 5.3 and section 5.3.8 in Chapter 5 of this Manual). Even smaller and cheaper versions with <i>natural frequencies</i> typically between about 5 and 25 Hz are commonly used in exploration <i>geophysics</i> . |
| <b>geophysics</b>               | The study of the Earth and its sub-systems by physical methods (seismic, gravimetric, magnetic, electric, thermic, radioactive and others).                                                                                                                                                                                                                                                                                                                               |
| <b>geosyncline</b>              | A crustal down-warp that is several hundreds or thousands of kilometers long and contains an exceptional thickness of accumulated sediments, on the order of 10 to 15 km.                                                                                                                                                                                                                                                                                                 |
| <b>geotechnical</b>             | Referring to the use of scientific methods and engineering principles to acquire, interpret, and apply knowledge of Earth materials for solving engineering problems.                                                                                                                                                                                                                                                                                                     |
| <b>geotechnical engineering</b> | The branch of engineering that deals with the analyses, design and construction of foundations and retaining structures on, within, and with soil and rock. Common examples include structural foundations and earth or rockfill dams.                                                                                                                                                                                                                                    |
| <b>geotechnical zonation</b>    | A mapping that portrays specified ground characteristics or mechanical properties in an area, typically showing zones that correspond to pre-defined geotechnical units. For example, a map of <i>shear wave velocities</i> derived from soil profile classifications. See Faccioli and Pessina (2003).                                                                                                                                                                   |
| <b>geotherm</b>                 | A profile of temperature as a function of depth inside the Earth. See <i>geothermal gradient</i> .                                                                                                                                                                                                                                                                                                                                                                        |

|                                   |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|-----------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>geothermal gradient</b>        | The rate of increase in temperature with depth in the Earth. By convention, this quantity is positive when increasing downward.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>geyser</b>                     | A fountain of hot water and steam ejected intermittently from a source heated by <i>magma</i> or hot rocks. See King and Igarashi (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>global plate motion model</b>  | A set of <i>Euler vectors</i> specifying relative <i>plate</i> motions. Such models can be derived for time spans of millions of years using rates estimated from sea-floor <i>magnetic anomalies</i> , directions of motion from the orientations of <i>transform faults</i> , and the <i>slip</i> vectors of earthquakes on transforms and at <i>subduction zones</i> . They can also be derived for time spans of a few years using space-based geodesy. See Stein and Klosko (2002).                                                                                                                        |
| <b>global tectonics</b>           | See <i>tectonics</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>glowing avalanche</b>          | Incandescent and turbulent mixture of <i>pyroclasts</i> , <i>volcanic gases</i> , and air that flows under <i>gravity</i> down the flanks of a <i>volcano</i> ; essentially synonymous with <i>nuée ardente</i> , <i>pyroclastic flow</i> , and <i>pyroclastic surge</i> .                                                                                                                                                                                                                                                                                                                                      |
| <b>Gondwana</b>                   | The proto-continent of the southern hemisphere in the late <i>Paleozoic</i> (see <i>geologic time</i> ). It included Australia, South America and Antarctica, as well as India, before they were drifting away from one another. It was named by the Austrian geologist Eduard Suess after Gondwana, a historic region in Central India. See Uyeda (2002), Duff (1993), and Hancock and Skinner (2000).                                                                                                                                                                                                         |
| <b>graben</b>                     | A down-dropped block in the Earth's <i>crust</i> resulting from extension, or pulling, of the <i>crust</i> . See also <i>horst</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>gradient</b>                   | The rate of change, e.g., of the <i>seismic velocity</i> with depth in the Earth.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>gradiometry (seismic wave)</b> | The study of the compatibility relationship between a <i>wave field</i> and its spatial and temporal <i>gradients</i> . Spatial gradients of the wavefield may be constructed using a dense <i>seismic array</i> of instruments or obtained from <i>strain</i> or <i>rotation</i> measurements. The compatibility relationship requires a model for a propagating seismic wave in 1, 2, or 3 dimensions that can be used to determine wave propagation <i>azimuth</i> , <i>wave slowness</i> , change in <i>geometrical spreading</i> , and change in azimuthal <i>radiation pattern</i> . See Langston (2007). |
| <b>gravity</b>                    | The attraction between two masses, such as the Earth and an object on its surface. Commonly referred to as the <i>acceleration</i> of <i>gravity</i> <i>g</i> . Changes in the gravity field can be used to infer information about the structure of the Earth's <i>lithosphere</i> and                                                                                                                                                                                                                                                                                                                         |

*upper mantle*. Interpretations of changes in the gravity field are generally applied to gravity values corrected for extraneous effects. The corrected values are referred to by various terms, such as free-air gravity, Bouguer gravity, and isostatic gravity, depending on the number and kind of corrections made.

|                         |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|-------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>gravity anomaly</b>  | The difference between the observed <i>gravity</i> and the theoretically calculated value based on a model.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>gravity loads</b>    | In <i>earthquake engineering</i> , the vertical loads created in the elements of a structure by the force of <i>gravity</i> . Gravity loads are sometimes separated into dead load (the weight of the structure) and live load (the weight of furnishings and other contents). See Hall (2003).                                                                                                                                                                                                                                                                                                                                                                      |
| <b>gravity waves</b>    | <i>Normal modes</i> in a surface layer with very low <i>shear-wave velocity</i> , such as unconsolidated sediments, may be affected significantly by <i>gravity</i> at long periods like by a <i>tsunami</i> in the ocean. Waves similar to the gravity waves in a fluid layer are possible in addition to the shortening of <i>wavelength</i> of normal modes by <i>gravity</i> . So-called “visible waves” with large <i>amplitude</i> and relatively long <i>periods</i> observed in the <i>epicentral area</i> of a <i>great earthquake</i> have been suggested to be gravity waves. See Satake (2002, p. 442-443) and Ben-Menahem and Singh (1981, p. 776-796). |
| <b>great earthquake</b> | An earthquake with <i>magnitude</i> greater than about $7\frac{3}{4}$ (or 8) is often called “great”.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Green's function</b> | In seismology, the vector displacement field generated by an impulsive force applied at a point in the Earth. When combined with the source function describing the discontinuities in <i>displacement</i> and traction across an internal surface, a Green's function can represent the seismic displacements caused by earthquake <i>faults</i> and explosions. Green's function was first introduced by George Green (1793–1841) as a solution to an inhomogeneous hyperbolic equation for a point source in space and time. See Aki and Richards (2002, p. 27–28).                                                                                               |
| <b>ground failure</b>   | A general reference to <i>landslides</i> , <i>liquefaction</i> , <i>fault displacement</i> , lateral spreads, and any other lasting consequence (permanent deformation) of shaking that affects the stability of the ground. See Youd (2003).                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>ground motion</b>    | Vibration of the ground, primarily due to <i>earthquakes</i> or other <i>seismic sources</i> . Ground motion is produced by <i>seismic waves</i> that are generated, e.g., by sudden <i>slip</i> on a <i>fault</i> , the collapse of sub-surface cavities or sudden pressure released by <i>explosions</i> , and - after traveling through the Earth - arrive at its surface. It is measured by a <i>seismograph</i> that records <i>acceleration</i> , <i>velocity</i> or <i>displacement</i> . In <i>engineering seismology</i> , it is usually given                                                                                                              |

in terms of a *time series* (an *accelerogram*), a *response spectrum* or *Fourier spectrum*. See Chapters 2, 4 and 15 of this Manual and Anderson (2003).

- ground motion relation**      See *attenuation relation*.
- ground-motion (strong-motion) parameter**      In *engineering seismology*, a parameter characterizing ground motion, such as *peak acceleration*, peak velocity, and peak displacement (peak parameters) or ordinates of *response spectra* and *Fourier spectra* (spectral parameters). See Chapter 15 of this Manual and and Campbell (2003).
- ground oscillation**      A geotechnical term specifying a ground condition in which *liquefaction* of subsurface layers decouples overlying non-liquefied layers from the underlying firm ground, allowing large transient *ground motions* or ground waves to develop within the liquefied and overlying layers. See Youd (2003).
- ground roll**      A term used in exploration seismology to refer to *surface waves* generated from explosions. They are characterized by low *velocity*, relatively low *frequency*, and high *amplitude*, and are observed in regions where the near-surface-layering consists of poorly consolidated, low-velocity sediments overlying more competent beds with higher velocity.
- ground settlement**      Permanent vertical *displacement* of the ground surface due to compaction or consolidation of underlying soil layers. See Youd (2003).
- ground shaking scenario**      A representation for a site or region depicting the possible ground shaking level or levels due to earthquake, in terms of useful descriptive parameters. See Chapter 15 of this Manual, and Faccioli and Pessina (2003).
- ground truth**      A jargon coming from the nuclear-test detection seismology, referring to information about the location of a *seismic event* that is derived from information not usually available to an analyst studying the event with a typical monitoring network. The ground truth can come from satellite observations, or from *fault traces* that break out at the Earth's surface, or from operation of a local network, drilling site or similar. Ground truth (GT) locations are sometimes associated with a number, e.g., "GT5", meaning a location believed to be accurate to within 5 km, obtained from ground truth. Today, ground truth events are essential for testing *seismic source* location algorithms and to validate seismic velocity models. See IS 8.5, 8.6 and 11.1 in this Manual and Richards (2002).
- group velocity**      For dispersive waves with frequency-dependent *phase velocity*,  $c(\omega)$ , the wave packet or energy propagates with the group velocity given by  $d\omega/dk$ , where  $k$  is the *wave number* given by

$\omega/c(\omega)$ . See Aki and Richards (2002, p. 253-254) and Fig. 2.10 in Chapter 2 of this Manual.

### **guided waves**

Guided waves are trapped in a waveguide by total *reflections* or bending of rays at the top and bottom boundaries. An outstanding example is the acoustic waves in the *SOFAR channel*, a low-velocity channel in the ocean. Since the absorption coefficient for sound in seawater is quite small for frequencies in the order of a few 100 Hz, transoceanic transmission is easily achieved. If the Earth's surface is considered as the top of a waveguide, *surface waves*, such as *Rayleigh*, *Love*, and their higher modes, are guided waves. The fault-zone trapped mode is a guided wave in the *low velocity fault zone*. Where they can exist, guided waves may propagate to considerable distances, because they are effectively spreading in only two spatial dimensions.

### **Gutenberg discontinuity**

The seismic velocity discontinuity marking the *core-mantle boundary* (CMB) at which the velocity of *P waves* drops from about 13.7 to about 8.0 km/s and that of *S waves* from about 7.3 to 0 km/s. The CMB reflects the change from the solid *mantle* material to the fluid outer *core*. See Fig. 2.79 in Chapter 2 of this Manual and Lay (2002).

### **Gutenberg-Richter relation**

An empirical relation expressing the *frequency (of occurrence)* distribution of *magnitudes* of *earthquakes* occurring in a given area and time interval. It is given by  $\log N(M) = a - bM$ , where  $M$  is the earthquake magnitude,  $N(M)$  is the number of earthquakes with magnitude greater than or equal to  $M$ , and  $a$  and  $b$  are constants (See *b-value*). It was obtained by Gutenberg and Richter (1941; 1949), and is equivalent to the *power-law* distribution of earthquake energies or moments. See Utsu (2002b, p. 723), and Eqs. (7.1) - (7.2) in Ben-Zion (2003).

### **G-waves (Gn)**

Another name for very long-period *Love waves*. Because the group velocity of Love waves in the Earth is nearly constant (4.4 km/sec) over the period range from about 40 to 300 seconds (see Fig. 2.10 in Chapter 2 of this Manual), their *waveform* is rather impulsive, and they have received this additional name. They are called G-waves after Gutenberg. It takes about 2.5 hours for G-waves to make a round trip of the Earth. After a large earthquake, a sequence of G-waves may be observed. They are named G1, G2, ..., Gn, according to their arrival times. The odd numbers refer to G-waves traveling in the direction from epicenter to station, and the even numbers to those leaving the *epicenter* in the opposite direction and approaching the station from the antipode of the epicenter. See section 2.3.4, Fig. 2.19 in Chapter 2 of this Manual and Kulhanek (2002, p. 346).



**gyroscope (Gyro)** A device, which contains a fast rotating axial-symmetric body, measuring or compensating for *rotational motions*; commonly used for navigation. The principle of its action is based on the law of balance of *kinetic moment*. Because this term can mean both a stable platform and a *rotational sensor*, it should be avoided or defined by the authors.

## H

**Hadean** An eon in *geologic time*, from 4.6 to 4 billion years ago. It is named after Hades, the underworld of Greek mythology.

**Hales discontinuity** See *N discontinuity*.

**half-life** The time required for a mass of a radioactive isotope to decay to one half of its initial amount. See King and Igarashi (2002).

**halfspace** A mathematical model bounded by a planar surface but otherwise infinite. Properties within the model are commonly assumed to be *homogeneous* and *isotropic*, unlike the Earth itself, which is heterogeneous and anisotropic. Nevertheless, the half space model is frequently used to perform theoretical calculations (forward modeling) in seismology.

**hanging wall** The overlying side of a non-vertical *fault* surface. See *footwall* and Figure 3.107 in Chapter 3 of this Manual.

**harmonic tremor** Oscillatory *ground motion* sustained for minutes to days or longer, commonly associated with volcanic *eruptions* and often observed during the precursory phase of eruptive activity. The spectral content of harmonic tremor is commonly dominated by one or two spectral peaks in the frequency range 1 to 5 Hz. See *volcanic tremor*; also Chapter 13 of this Manual and McNutt (2002).

**Haskell model** A representation of the earthquake source by a ramp-function *slip* propagating unilaterally at a uniform velocity along the long dimension of a narrow *rectangular fault* surface (Haskell, 1964). It is specified by 5 parameters, i.e., *fault length*, *fault width*, final slip, *rupture velocity* and *rise time*. See Aki and Richards (2002, p. 498-503), and Purcaru and Berckhemer (1982) for a compilation of earthquakes with these parameters determined by various authors. See IS 3.1 in this Manual, Udias (2002, p. 99-100), and Eq. (3.8) in Ben-Zion (2003).

**Hawaiian eruption** Eruption of highly fluid (low-viscosity) *magma* that commonly produces *lava fountains* and forms thin and widespread *lava flows*, and generally very minor *pyroclastic deposits* around the *vent*.

|                             |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|-----------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>hazard</b>               | See <i>earthquake hazard</i> and <i>volcano hazard</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>hazard master model</b>  | A type of master model of earthquakes in which the <i>Probabilistic Seismic Hazard Analysis</i> (PSHA) is used for unifying multi-disciplinary data regarding earthquakes in a given region. The integration of the multi-disciplinary data is done in the end product, in contrast to a physical master model. See Aki (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>head waves</b>           | Head waves are interface waves observed in a <i>halfspace</i> that is in welded contact with another halfspace with higher velocity when the <i>seismic source</i> is located in the lower-velocity medium. The wave enters the higher velocity medium at the critical <i>incidence angle</i> and, after traveling along the interface at the higher velocity, re-enters the lower velocity medium at the critical angle. The <i>wavefront</i> in the lower-velocity halfspace is a part of the surface of an expanding cone. For this reason, head waves are sometimes called “conical waves.” See section 2.5.3.5 and Figs. 2.36 and 2.37 in Chapter 2 of this Manual and Aki and Richards (2002, p. 203-209).                                                                  |
| <b>heat flow</b>            | In geophysics, heat flow usually means the rate of heat flowing from the Earth’s interior per unit area of the surface. It is measured near the Earth’s surface as the product of thermal conductivity and vertical temperature gradient. The average heat flow of the Earth is estimated to be about $2 \mu\text{cal cm}^{-2} \text{sec}^{-1}$ , corresponding to about $84 \text{ mW m}^{-2}$ (milli-Watt per square meter) in SI units. Heat flow conditions have a great influence on the <i>rheological properties</i> and thus on the depth of the <i>brittle-ductile transition</i> and the occurrence of earthquakes in the <i>Earth’s crust</i> . See also <i>heat flow unit</i> and <i>thermal conductivity</i> , Chapter 81.4 in Lee et al. (2002), and Morgan (2002). |
| <b>heat flow unit (HFU)</b> | Terrestrial <i>heat flow</i> values are commonly given in the SI units $\text{mW m}^{-2}$ (milli-Watt per square meter). However, in older publications and maps the heat flow is usually given in cgs units of $\mu\text{cal/cm}^2 \text{sec}$ , or $10^{-6} \text{ cal cm}^{-2} \text{sec}^{-1}$ . The conversion relationship is $1 \mu\text{cal cm}^{-2} \text{sec}^{-1} = 41.84 \text{ mW m}^{-2}$ . Occasionally some authors used the term “heat flow unit” or “HFU” for “ $\mu\text{cal cm}^{-2} \text{sec}^{-1}$ ”.                                                                                                                                                                                                                                                      |
| <b>helioseismology</b>      | Study of the Sun's <i>free oscillations</i> (both acoustic and gravity modes), typically through spatially resolved Doppler spectroscopy (see <i>Doppler effect</i> ).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>hertz (Hz)</b>           | The <i>SI</i> derived unit for frequency; expressed in cycles per second ( $\text{s}^{-1}$ ). It is named after Heinrich Rudolph Hertz (1857-94) who discovered electromagnetic waves.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |

|                                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|---------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>heterogeneity</b>                  | A medium is heterogeneous when its physical properties change along the space coordinates. A critical parameter affecting seismic phenomena is the scale of heterogeneities as compared with the seismic <i>wavelengths</i> . For a relatively large wavelength, for example, an intrinsically <i>isotropic</i> medium with oriented heterogeneities may behave as a <i>homogeneous</i> anisotropic medium. See Cara (2002).                                                                                                                                                                                                                                                                                                                                                                           |
| <b>hexagonal symmetry</b>             | Symmetry by rotation of 60° around one axis, the 6-fold symmetry axis. Hexagonal symmetry and cylindrical symmetry are equivalent for the elastic properties. There are 5 independent elastic parameters in hexagonal symmetry systems. See Cara (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>higher order statistics</b>        | <p>Assuming the expectation of a continuous distribution as</p> $E[X] = \int_{-\infty}^{\infty} xp(x)dx$ <p>with the distribution function <math>p(x)</math> of the random variable <math>x</math>, the central statistic moment <math>m</math> of order <math>k</math> is given by <math>m_k = E[(X - E[X])^k]</math>, <math>k &gt; 1</math>. The second central moment is the variance, the third central moment normalized by the variance is the <i>skewness</i> <math>S</math>, and the fourth central moment normalized by the variance is the <i>kurtosis</i> <math>K</math>:</p> $S = \frac{m_3}{m_2^{3/2}}$ $K = \frac{m_4}{m_2^2}$ <p>Skewness and kurtosis are parameters that describe deviations of a time series or a random value distribution from a <i>Gaussian distribution</i>.</p> |
| <b>high-frequency (HF) earthquake</b> | In volcano seismology, generally synonymous with “volcano-tectonic”, “brittle-failure”, or “A-type” <i>earthquake</i> . See Chapter 13 in this Manual and McNutt (2002). In general, earthquakes that radiate more than (magnitude-dependent) average of high-frequencies are characterized by higher than average <i>stress drop</i> and/or <i>rupture velocity</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>high-pass filter</b>               | <i>Filter</i> which removes the low frequency portion of the input signal. See IS 5.2 of this Manual and Scherbaum (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>hodograph</b>                      | In the early era of seismology, the term sometimes used for the time of <i>arrival</i> of various <i>seismic phases</i> as a function of <i>epicentral distance</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>Holocene</b>                       | The most recent epoch in the <i>geologic time</i> , from 10,000 years ago to the present.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |

|                                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
|------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Holocene fault</b>                          | Applied to <i>faults</i> , this term indicates the time of most recent fault <i>slip</i> within the <i>Holocene</i> . Faults of this age are commonly considered <i>active</i> , based on the observation of historical or <i>palaeoseismic</i> activity on faults of this age in other locales.                                                                                                                                                                                                                                                                                                                             |
| <b>homogeneous</b>                             | Being uniform, of the same nature; the opposite of <i>inhomogeneous</i> . Although the real Earth as a whole is largely inhomogeneous and weakly <i>anisotropic</i> , it can be considered for <i>seismic modeling</i> in a first approximation, at least in parts, as being homogeneous and isotropic.                                                                                                                                                                                                                                                                                                                      |
| <b>horizontal-to-horizontal spectral ratio</b> | <i>Spectral ratio</i> of horizontal components at a target site to those at a reference site. If a reference site is virtually <i>site-effect</i> free, this corresponds to the true <i>site amplification</i> of the target site. Care must be taken to use a deep borehole station as a reference site, because seismic motions at depth are modified by the presence of the free surface differently from those at the surface. See Chapter 14 of this Manual and Kawase (2003).                                                                                                                                          |
| <b>horst</b>                                   | An upthrown block lying between two steep-angled <i>fault</i> blocks.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>hot spot</b>                                | A long-lived volcanic center that is thought to be the surface expression of a persistent rising <i>plume</i> of hot <i>mantle</i> material. The position of hot spots is used in many <i>plate</i> movement studies as reference coordinate system. See Uyeda (2002, p. 59-61) and Olson (2002). See also <i>melting anomaly</i> .                                                                                                                                                                                                                                                                                          |
| <b>hummocks</b>                                | A term in volcanology, referring to characteristic topographic features for <i>debris-avalanche</i> deposits (e.g., hummocky topography), as demonstrated during the 18 May 1980 eruption of Mount St. Helens, Washington State; their occurrence provides tell-tale evidence for one or more sector collapses of a volcanic edifice in its eruptive history.                                                                                                                                                                                                                                                                |
| <b>H/V spectral ratio</b>                      | Ratio of the <i>Fourier amplitude spectra</i> of horizontal and vertical components of <i>ambient noise</i> , or of earthquake <i>ground motions</i> , recorded at a site. Typically used to identify the presence of site-specific dominant frequencies in such motions. If this technique is applied to <i>microtremor</i> data, it is sometimes called “ <i>Nakamura method</i> ”. If applied to the P and S portion of seismic data from distant sources to invert for a one-dimensional structure, it is called the “ <i>receiver function method</i> ”. See Chapter 14 of this Manual and Faccioli and Pessina (2003). |
| <b>hyaloclastic rocks</b>                      | A general term for a wide variety of glassy volcanic rocks formed by the fragmentation of <i>magma</i> or <i>lava</i> in the presence of water. The adjective hyaloclastic has Greek roots (hyalos, “glass”; klastos, “broken”).                                                                                                                                                                                                                                                                                                                                                                                             |

|                             |                                                                                                                                                                                                                                                                                                                                                                                                              |
|-----------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>hybrid earthquake</b>    | In volcano seismology, an earthquake with characteristics of both “ <i>high-frequency</i> ” and “ <i>long-period</i> ” earthquakes. Often applied to earthquakes that have an impulsive, high-frequency onset with distinct <i>P</i> and <i>S</i> waves followed by an extended, low-frequency coda. Usage is variable.                                                                                      |
| <b>hydroacoustic</b>        | Pertaining to compressional (sound) waves in water, in particular in the ocean. Hydroacoustic waves may be generated by explosions, volcanic eruptions, <i>airgun</i> shots or earthquakes. See <i>T-phase</i> and <i>SOFAR channel</i> .                                                                                                                                                                    |
| <b>hydrology</b>            | The science that deals with water on and under the ground surface, including its properties, circulation, and distribution.                                                                                                                                                                                                                                                                                  |
| <b>hydrophone</b>           | A device for measuring <i>pressure</i> fluctuations in water using a piezoelectric sensor.                                                                                                                                                                                                                                                                                                                   |
| <b>hydrostatic pressure</b> | There are two variations on the usage of this term. In one usage, it simply refers to a <i>stress</i> state that is <i>isotropic</i> , i.e., shear stresses are zero. Another common usage is that “hydrostatic pressure” is a special case of “ <i>lithostatic pressure</i> ” where the density of the overlying column is chosen to be water density. See Ruff (2002).                                     |
| <b>hypocenter</b>           | A point in the Earth (also called <i>earthquake focus</i> ) where the rupture of the rocks initiates during an earthquake. Its position, in practice, is determined from <i>arrival times</i> of the <i>seismic phase onsets</i> recorded by <i>seismographs</i> . The point on the Earth's surface vertically above the hypocenter is called the <i>epicenter</i> . See Lee and Stewart (1981, p. 130-139). |
| <b>hypocentral distance</b> | Distance from an observation point to the <i>hypocenter</i> of an earthquake.                                                                                                                                                                                                                                                                                                                                |

## I

|               |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
|---------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>I, i</b>   | The symbol <i>I</i> is used to indicate that part of a <i>ray</i> path that has traversed the Earth's <i>inner core</i> as a <i>P</i> -wave. On the other hand, <i>i</i> is used to indicate <i>reflection</i> at the boundary between outer and inner core in the same manner that <i>c</i> is used for reflections at the <i>core-mantle boundary</i> . For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). |
| <b>IASPEI</b> | An acronym for the International Association of Seismology and Physics of the Earth's Interior ( <a href="http://www.iaspei.org/">http://www.iaspei.org/</a> ). See Adams (2002) and Engdahl (2003).                                                                                                                                                                                                                                                                                        |

|                                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|---------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>IASPEI magnitude standards</b>     | Standards for <i>magnitude</i> measurements elaborated by an international working group of the Commission on Seismic Observation and Interpretation (IASPEI, 2005 and 2013). These standards aim at a) the reduction of measurement errors due to the application of different procedures for calculating magnitudes of the same type, b) the wider use of <i>broadband magnitudes</i> from <i>body</i> and <i>surface</i> waves that are in better agreement with original definitions and c) maximizing the benefits of modern digital broadband records, simulation <i>filters</i> and signal processing tools for collecting many more globally compatible <i>amplitude</i> , <i>period</i> and magnitude data for research and other applications. |
| <b>identity tensor</b>                | A tensor of second rank, possessing the following property: the scalar product (from right or left side) of this tensor and of any vector (tensor) gives the same vector (tensor). The identity tensor is equal to $\mathbf{ii} + \mathbf{jj} + \mathbf{kk}$ , where $\mathbf{i}$ , $\mathbf{j}$ , $\mathbf{k}$ are orthogonal unit vectors, and the sign of the tensor product is omitted.                                                                                                                                                                                                                                                                                                                                                              |
| <b>igneous rock</b>                   | A rock that has formed from molten or partly molten material. See Christensen and Stanley (2003), and Press and Siever (1994).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>impedance</b>                      | In seismology, the product of <i>seismic wave velocity</i> and <i>density</i> ( $v\rho$ ).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>impedance contrast</b>             | The ratio of the <i>impedances</i> $v\rho$ at the <i>discontinuity</i> (or <i>interface</i> ) between two adjacent media: $v_1\rho_1/v_2\rho_2$ . It largely controls, besides the <i>incidence angle</i> , the <i>refraction angle</i> , the <i>reflection</i> and transmission coefficients at the discontinuity. For the ratio of the <i>amplitudes</i> $A_1$ and $A_2$ of an acoustic wave group travelling through two adjacent media (neglecting <i>geometrical spreading</i> and <i>attenuation</i> ) holds $A_1/A_2 = [(v_2\rho_2)/(v_1\rho_1)]^{1/2}$ .                                                                                                                                                                                         |
| <b>impulse response</b>               | Output <i>signal</i> of a stationary linear system to an impulsive (delta function) input.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>incidence angle <math>i</math></b> | The angle between a seismic <i>ray</i> , incident on an <i>interface</i> , and the vertical (normal) to this <i>discontinuity</i> . See also <i>reflection angle</i> and <i>refraction angle</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>incident wave</b>                  | The wave that propagates toward an <i>interface</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>induced earthquake</b>             | An earthquake that results from changes in crustal <i>stress</i> and/or strength due to man-made sources (e.g., underground mining and filling of a high dam), or natural sources (e.g., the <i>fault slip</i> of a major earthquake). As defined less rigorously, “induced” is used interchangeably with “ <i>triggered</i> ” and applies to any earthquake associated with a stress change, large or small. See <i>triggered earthquake</i> ; also McGarr et al. (2002).                                                                                                                                                                                                                                                                               |

|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|---------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>inertial sensor</b>          | A <i>translational</i> or <i>rotational</i> ground motion sensor sensitive to <i>accelerations</i> , <i>velocities</i> or <i>displacements</i> affixed and referenced to some stated frame (generally to the non-inertial frame of the rotating, gravitating Earth or a built structure), and generally measuring one of the six degrees of freedom of whole-body motion. Inertial sensors respond to accelerations but, because of the inductive readout method, often output a signal proportional to ground-motion velocity or ground motion. A well built inertial sensor has a bob moving “freely” in only one degree of freedom and is sensitive to accelerations for signal frequencies (much) smaller than the <i>eigenfrequency</i> of the sensor or displacements for frequencies (much) larger than the sensor’s eigenfrequency in that degree of freedom (see Fig. 5.3 upper left and lower left in Chapter 5 of this Manual). Because of hinging effects, very often inertial sensors intrinsically mix the linear and angular acceleration. To make sure that this does or does not happen, one has to examine the hinging mechanism of the sensor. |
| <b>inertia tensor</b>           | See <i>tensor of inertia</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>infragravity waves</b>       | Infragravity waves are the component of ocean waves at <i>periods</i> longer than 25 seconds. Infragravity waves control long period (>30 seconds) vertical component seismic <i>noise</i> levels at the seafloor and at coastal sites on land. Infragravity waves are trapped along shorelines as edgewaves (see <i>edge effect</i> ) or surfbeat and are generated by nonlinear mechanisms by the interaction of shorter period (wind driven) ocean waves along coastlines. See Webb (2002, p. 307).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>infrasonic</b>               | Pertaining to low-frequency (sub-audible) compressional (sound) waves in the atmosphere (see <i>infrasound</i> ).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>infrasound</b>               | Acoustic waves at <i>frequencies</i> approximately from the seismic to the audible range. Typical for sound waves is that they propagate over great distances through the atmosphere and that the propagation conditions strongly depend on the changing parameters in the dynamic medium. See Chapter 17 of this Manual and Richards (2002, p. 370).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>inhomogeneous</b>            | The opposite of <i>homogeneous</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>inhomogenous plane waves</b> | <i>Plane waves</i> with <i>amplitudes</i> varying in a direction different from the direction of propagation. The propagation velocity is lower than that for regular plane waves. For example, <i>Rayleigh waves</i> are composed of inhomogeneous <i>P</i> and <i>SV waves</i> propagating horizontally with amplitudes decaying with depth. They are also called “evanescent waves.” See Aki and Richards (2002, p. 149-157).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |

|                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|--------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>inner core</b>              | The central part of the Earth's <i>core</i> . It extends from a depth of about 5200 km to the Earth's center. It has non-zero <i>rigidity</i> , in contrast to the outer core and was discovered in 1936 by Inge Lehmann.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>inner core anisotropy</b>   | Refers to elastic anisotropy of the <i>inner core</i> . The P velocity along the north-south direction in the inner core is about 3% faster than along east-west directions although the amount appears to vary spatially. See Song (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>inner core rotation</b>     | The <i>geodynamo</i> is expected to force the <i>inner core</i> to rotate relative to the <i>mantle</i> . The observational evidence for such a differential inner core rotation has been reported but it is still under debate. See Chapter 56 in Lee et al. (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>instrumental noise</b>      | Unwanted data part originating from the instrument used for the measurement. See <i>self noise</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>instrument response</b>     | In order to recover the <i>ground motion</i> at a recording site, one must deconvolve (see <i>deconvolution</i> ) the contribution of the recording instrumentation. A modern instrument response can be broken down into two stages: a) the transformation of ground motion to an electrical energy followed b) by the transformation of that electrical energy into an output <i>signal</i> which can produce a permanent record (historically paper or photographic record, today mostly a digital stream recorded on a computer storage unit). See Chapter 5 of this Manual, as well as Wielandt (2002), Scherbaum (2002) and Eq. (3.10) in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>intensity, macroseismic</b> | A cumulative measure of the effects of an <i>earthquake</i> at a particular place at the Earth's surface on humans and (or) structures and thus a measure of the strength of earthquake shaking. Different from <i>magnitude</i> , the intensity at a point depends not only on the earthquake strength (released <i>seismic energy</i> ) or earthquake size (released <i>seismic moment</i> ), but also on the <i>epicentral distance</i> , the <i>focal depth</i> , the position of the observation point with respect to the type of <i>focal mechanism</i> and its <i>directivity</i> , and the local site conditions. Intensity values are usually given as integer Roman numerals ranging from 0 to 12 (see <i>intensity scales</i> , <i>macroseismic</i> ). Although coarse, they are a valuable complementary analogue for physical <i>ground motion parameters</i> , correlating best with ground velocity. Moreover, intensity of shaking, based on the assessment of real post-event damage, is a suitable site-related empirical risk estimator for scenario calibrations. See Chapter 12 of this Manual. Intensity is now also quantitatively estimated using ground motion measurements; such quantitative intensity measurement is called "Instrumental Intensity" to distinguish it from the qualitative seismic intensity. See Musson (2002). |



**intensity scales, macroseismic** *Macroseismic intensity* assessments assign integer numerals to observed physical damages and perceived motions, thus classifying the shaking strength observed at a site. When these intensity classes are arranged in a systematic and mutually linked reasonable way they form a macroseismic scale. Macroseismic intensity scales can be considered as an analogue to the 12-degree Beaufort wind-strength scale which is also based on perceptions and physical effects. The first, also internationally widely used macroseismic intensity scale was the ten-degree Rossi-Forel Scale of 1883. The scale of Sieberg (1912, 1923) became the foundation of all modern twelve-degree scales. A later version became known as the *Mercalli-Cancani-Sieberg Scale*, or *MCS Scale* (Sieberg, 1932). It was translated into English by Wood and Neuman (1931), becoming the *Modified Mercalli Scale (MM Scale)*, and after an extensive revision by Richter (1958) the “Modified Mercalli Scale of 1956” (MM56). Since the 1960s, the *Medvedev-Sponheuer-Karnik Scale (MSK Scale)*, Sponheuer and Karnik, 1964) was widely used in Europe. It was based on the MCS, MM56 and previous work of Medvedev (1962) in Russia, but greatly developed the quantitative aspects of the scale. Since the late 1990s the much more elaborated *European Macroseismic Scale 98 (EMS-98)*; Grünthal, 1998) has essentially replaced the usage of the old MSK scale although content-wise the EMS is more or less compatible with it. A more recent modification of the *MM Scale*, termed *MMI*, was proposed by Stover and Coffman (1993). All these scales have twelve degrees of intensity. Only the scale used in Japan, the *JMA intensity scale*, uses less degrees, originally 6, and since recent modifications (JMA 1996, 2009) 10 degrees. See Chapter 12 of this Manual, the specific entries in this glossary for the specific scales, and Musson et al. (2010) for a comparison between the scales and a discussion on their evolution.

**interface** In a continuum, a surface of *discontinuity* that separates two different media.

**interference** In physics, interference is the phenomenon in which two waves superpose each other to form a resultant wave of greater or lower *amplitude*. Interference usually refers to the interaction of waves that are *coherent* with each other, either because they come from the same source or they have the same or nearly the same *frequency*. When two waves with the same frequency combine, the resulting pattern is determined by the *phase* difference between the two waves.

**interferogram** The photographic or digital record of the *interference* pattern of electromagnetic or other types of superposing waves that has diagnostic potential to decipher the original state and/or the travelpath history of the interfering waves.

|                                            |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|--------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>interferometry</b>                      | Refers to a family of techniques in which electromagnetic waves are superimposed in order to extract information about the waves, respectively about the position and/or shape of the object(s) that receive(s) or reflect(s) the electromagnetic waves radiated from an electromagnetic source. Interferometry is an important investigation technique in many fields, amongst them astronomy, engineering and optical metrology, remote sensing, seismology, oceanography, nuclear and particle physics.                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>interferometry, 2-pass differential</b> | Approach to calculating <i>interferograms</i> , which uses two radar images and a digital elevation model (DEM). Also called DEM-elimination. See Feigl (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>interferometry, 3-pass</b>              | Approach to calculating <i>interferograms</i> , which uses three radar images, but no elevation model. See Feigl (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>International System of Units (SI)</b>  | An internationally agreed system of units, which is abbreviated as SI for “Système International d’Unités”, and is recommended for use in all scientific and technical fields (NBS, 1981). The international system of units is comprised of 7 base units: (1) length in meter (m), (2) mass in kilogram (kg), (3) time in second (s), (4) electric current in ampere (A), (5) thermodynamic temperature in kelvin (K), (6) amount of substance in mole (mol), and (7) luminous intensity in candela (cd). It also has a number of derived and supplementary units. Some derived units with special names are: <i>frequency</i> in hertz (Hz), force in newton (N), <i>pressure</i> or <i>stress</i> in pascal (Pa), energy, work, or quantity of heat in joule (J), power in watt (W), electric charge in coulomb (C), and electric potential in volt (V). See Lide (2002) and NBS (1981) for a detailed review of SI and conversion factors. |
| <b>interplate</b>                          | Interplate pertains to processes between <i>lithospheric plates</i> (see <i>lithosphere</i> ).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>interplate coupling</b>                 | The qualitative ability of a <i>subduction thrust fault</i> to lock and accumulate <i>stress</i> . Strong interplate coupling implies that the fault is locked and capable of accumulating stress whereas weak coupling implies that the fault is unlocked or only capable of accumulating low stress. A fault with weak interplate coupling could be <i>aseismic</i> or could slip by <i>creep</i> . See <i>locked fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>interplate earthquake</b>               | An earthquake that has its source in the <i>interface</i> between two lithospheric <i>plates</i> , such as in a <i>subduction zone</i> or on a <i>transform fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>intraplate</b>                          | Intraplate pertains to processes within the Earth’s lithospheric <i>plates</i> (see <i>lithosphere</i> ).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |

|                                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|-------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>intraplate earthquake</b>                    | An earthquake that has its source in the interior of a lithospheric <i>plate</i> . Often the reason for the specific location of such an earthquake or its relation to an identifiable <i>fault</i> is not evident.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>intrinsic attenuation</b>                    | <i>Attenuation of seismic waves</i> due to intrinsic absorption caused by internal <i>friction</i> (anelasticity). See Section 5.5 in Aki and Richards (2002, p. 161-177).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>inverse problem</b>                          | A problem in which the intent is to infer the values of the parameters characterizing a system, given some observed data. Solving an inverse problem may be summarized as $\rightarrow$ observed data $\rightarrow$ model $\rightarrow$ estimates of model parameters. An inverse problem is the opposite of a <i>forward problem</i> ; it can be formulated as a problem of iterative data fitting or, more generally, as a problem of probabilistic inference. Usually, there is no direct and unique solution to the inverse problem. A rare exception is the solution of the <i>Wiechert-Herglotz</i> formula under strict conditions. See <i>inversion</i> , Mosegaard and Tarantola (2002), Curtis and Snieder (2002), Menke (1984), and Tarantola (2005). |
| <b>inverse refraction diagram (for tsunامي)</b> | A <i>refraction</i> diagram drawn from coastal observation points. Iso-time contours correspond to the <i>travel time</i> encompassed by the ray path to each station from the source region on the map. See Suyehiro and Mochizuki (2002, p. 443).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>inversion</b>                                | Given a model hypothesized for explaining a set of observed data, inversion is a procedure for determining the model parameters from the observed data. Various issues involved in this procedure, such as the evaluation of the resolution and error in the estimated parameters, constitute the <i>inverse problem</i> . In contrast the prediction of observables for a model with assumed parameters is called the <i>forward problem</i> . See Mosegaard and Tarantola (2002), Curtis and Snieder (2002), Menke (1984) or Tarantola (2005).                                                                                                                                                                                                                 |
| <b>Ishimoto-Iida's relation</b>                 | An empirical relation expressing the <i>frequency of occurrence</i> of the maximum <i>amplitudes</i> of earthquakes recorded at a <i>seismograph</i> station obtained by Ishimoto and Iida (1939). Although the concept of <i>magnitude</i> was not known to them, the power law found for the amplitude distribution is reducible to the power law for earthquake energy implied in <i>Gutenberg-Richter's</i> frequency-magnitude <i>relation</i> . They recognized the significance of its departure from the Boltzmann exponential law for statistical energy distribution. See Utsu (2002b), and Eqs. (7.1) - (7.2) in Ben-Zion (2003).                                                                                                                     |
| <b>isochron</b>                                 | In seismology, a line on a map or a chart connecting points of equal time or time interval, e.g., <i>arrival times</i> of the <i>onset</i> of <i>seismic waves</i> from a source. In <i>geochronology</i> , a line                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |

connecting points of equal age. In marine geophysics, the *magnetic lineation* with equal age, often abbreviated as chron. See Cox (1973), Uyeda (1978), and Duff (1993).

**isopach** As used in *volcanology*, a line joining points of equal thickness in a *pyroclastic* deposit.

**isopleth** As used in *volcanology*, a line joining points where the sizes of the largest *pyroclasts* are the same.

**isoseismal** A closed curve bounding the area within which the *intensity* from a particular earthquake was predominantly equal to or higher than a given value. See Chapter 12 of this Manual and Musson and Cčić (2002).

**isostasy** The application of the principle of buoyancy on the Earth's *lithosphere*. If the mass of any column of rocks above a specified level is everywhere the same, this part of the lithosphere is in isostatic equilibrium. Dynamic processes can destroy this balance, but buoyancy will force to restore the equilibrium.

**isotope** Various species of atoms of a chemical element with the same atomic number, but different atomic weights.

**isotropic** Means the physical properties of a medium, e.g., its *density* or *velocity* of *seismic wave* propagation, are independent on any direction in space.

## J

**J** The symbol J is used to indicate that part of a *ray* path which has traversed the Earth's *inner core* as an *S wave*. Unambiguous observations of such waves have not yet been achieved, but they may be possible in future. For this and other such symbols used in *seismic phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**jerk** Rate of change of *acceleration*.

**JMA magnitude ( $M_J$ )** *Magnitude* for earthquakes in Japan and its vicinity, assigned by JMA using data from the JMA network.  $M_J$  is fairly close to  $M_s$  for shallow earthquakes and  $mB$  for intermediate and deep earthquakes. See Chapter 3 in this Manual and Chapter 44 in Lee et al. (2002).

**JMA intensity scale** The *macroseismic intensity scale* used by the Japanese Meteorological Agency (JMA). Modified four times since its first definition in 1884. From 1996 onward, the scale has 10 grades designated as 0, 1, 2, 3, 4, 5-, 5+, 6-, 6+, and 7. See

*intensity scale*, Chapter 12 in this Manual, and Musson et al. (2010).

**joint transverse-angular coherence function** Measure of *coherency* between two transmitted *wave fields* as a function of *incident angles* and the transverse-separation of receivers. See Sato et al. (2002, p. 202).

**joule (J)** The SI derived unit for energy, or work, or quantity of heat, expressed in  $\text{m}^2 \text{kg s}^{-2}$ , or Newton meter (Nm). It is named after James Joule (1818-89), who established the mechanical equivalence of heat, and is today used instead of the outdated unit calorie ( $1 \text{ cal} = 4.184 \text{ J}$ ).

**Julian day** As used in seismology, the number of a day during the year (also referred to as Day-Of-Year, DOY), counted continuously from January 1 as Day 1. For example, February 2 is Julian day 33 in this method of counting. Note that for non-leap and leap years, the Julian Day for the 28<sup>th</sup> of February differs by one day.

**juvenile** An adjective commonly used to describe eruptive materials produced from the *magma* involved in the *eruption*, in contrast to *lithic* materials derived from preexisting rocks, volcanic or non-volcanic. See *lithics*.

## K

**K** *Seismic phase* symbol that stands for the *P waves* lag of *seismic waves* that traveled through the outer *core* (with K for the German word for core = Kern). For this and other such symbols used in *seismic phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**kelvin (K)** A base unit of thermodynamic temperature in the *international system of units (SI)*. The official definition is: “The kelvin, unit of thermodynamic temperature, is the fraction  $1/273.16$  of the thermodynamic temperature of the triple point of water.” See NBS (1981).

**kilogram (kg)** A base unit of mass in the *international system of units (SI)*. The official definition is: “The kilogram is the unit of mass; it is equal to the mass of the international prototype of mass of the kilogram.” See NBS (1981).

**kiloton** An energy unit, now defined as 4.184 trillion ( $10^{12}$ ) Nm (or Joules, J). The term originates from an estimate of the energy released by the explosion of a thousand tons of the chemical explosive Trinitrotoluol (TNT). See Richards (2002).

|                                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
|------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>kinematic</b>                         | Referring to the general movement patterns and directions, usually expressed in physical units of <i>displacement</i> (m), <i>velocity</i> (m/s) or <i>acceleration</i> (m/s <sup>2</sup> ).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>kinematic units</b>                   | See <i>kinematic</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>kinematic ray equations</b>           | The ordinary differential equations governing the geometry of a <i>ray</i> path. The independent variable is typically the <i>travel time</i> or ray path length, and the dependent variables, the ray position and direction. See Chapman (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>kinematics (of seismic source)</b>    | A description of the <i>deformation</i> or motions within a <i>seismic source</i> region in geometrical terms such as <i>fault slip</i> or transformational <i>strain</i> , without reference to the <i>stresses</i> or forces that are responsible. See <i>dynamics (of seismic source)</i> ; also Aki and Richards (2002 or 2009, Box 5.3, p. 129).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>kinetic moment</b>                    | Rotational analogue of impulse. It is equal to the sum of moments of momentum and of proper kinetic moment (sometimes called “dynamic spin”). Kinetic moment of a body is calculated relatively to a fixed point in a system of reference and depends upon the chosen “pole” (another point, fixed in the body). For a rigid body, if its center of mass is taken as a “pole,” and the origin of an inertial system of reference is taken as a fixed point, the kinetic moment is equal to $\mathbf{r} \times \mathbf{p} + \mathbf{J} \boldsymbol{\omega}$ , where $\mathbf{r}$ is the radius-vector of the center of mass, $\mathbf{p}$ is the momentum (impulse) of the body, $\mathbf{J}$ is the <i>tensor of inertia</i> calculated relative to the center of mass, and $\boldsymbol{\omega}$ is the angular velocity of the body. The time derivative of the kinetic moment represents the inertial term in the law of balance of <i>torque</i> (the <i>torques</i> must be calculated relative to the same fixed point). In <i>micropolar</i> theories, the existence of proper kinetic moment of particles and resistance of the medium to their rotation causes the existence of <i>rotational waves</i> . |
| <b>Kirchhoff surface integral method</b> | This method originates from Kirchhoff’s solution of the Cauchy problem, i.e., finding a solution of the wave equation for given initial conditions on the wave function and its time derivative defined for the whole space. The solution was expressed as a surface integral of functions given as initial conditions, properly time-shifted and weighted over a spherical surface centered at the observation point. It can be generalized to an integral over a non-spherical surface enclosing the region where the homogeneous wave equation is valid. It can be used, for example, to represent scattered waves from a rough surface as a surface integral. The integral reduces to ray-theoretical solution for the planar surface. See Chapman (2002), and biographies of Augustin Louis Cauchy and Gustav Kirchhoff in Howell (2003).                                                                                                                                                                                                                                                                                                                                                                     |

|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                            |
|---------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Knopoff model</b>            | A shear crack model for earthquakes investigated by Knopoff (1958). See Madariaga and Olsen (2002), and Scholz (2002). See also <i>fault model</i> .                                                                                                                                                                                                                                                                       |
| <b>Kostrov model</b>            | A self-similar circular rupture model for earthquakes introduced by Kostrov (1966). See Madariaga and Olsen (2002), Das and Kostrov (1988). See also <i>fault model</i> .                                                                                                                                                                                                                                                  |
| <b>Kramers-Kronig relations</b> | Also called <i>dispersion relations</i> . Relationships between the <i>amplitude</i> and <i>phase</i> of the frequency response resulting from the requirement that the corresponding impulse response is causal. See <i>frequency response</i> , <i>impulse response</i> and <i>causal signal</i> . See also Aki and Richards (2002 or 2009, p. 167-169).                                                                 |
| <b>kurtosis</b>                 | A quantitative measure of the deviation of a random value distribution from the “normal” bell-shaped <i>Gaussian distribution</i> . Kurtosis is a measure of the bulge of the distribution. It is 3 for normally distributed random variables. Positive (negative) deviations from 3 indicate widening (narrowing) of the distribution. See <i>higher-order statistics</i> .                                               |
| <b>L</b>                        |                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>L (LQ, LR)</b>               | The symbol L is used to designate long-period <i>surface waves</i> . When the type of surface wave is known, LQ and LR are used for <i>Love</i> and <i>Rayleigh waves</i> , respectively. For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).                                                                                                |
| <b>lahar</b>                    | A widely used Indonesian term for volcanic debris flow. See <i>mudflow</i> .                                                                                                                                                                                                                                                                                                                                               |
| <b>Lamé constants</b>           | Two physical constants, $\lambda$ and $\mu$ , that linearly relate <i>stress</i> to <i>strain</i> in a <i>homogeneous</i> and <i>isotropic</i> elastic solid in the classical elasticity theory. $\mu$ is the <i>rigidity</i> ( <i>shear modulus</i> ) and $\lambda$ a combination between <i>bulk modulus</i> $\kappa$ and rigidity $\mu$ ( $\lambda = \kappa - 2\mu/3$ ). See section 2.2.2 in Chapter 2 of this Manual. |
| <b>landslide</b>                | The down-slope movement of soil and/or rock.                                                                                                                                                                                                                                                                                                                                                                               |
| <b>lapilli</b>                  | Of Italian origin (meaning “little stones”), a term given to <i>tephra</i> ranging in size between 2 and 64 mm.                                                                                                                                                                                                                                                                                                            |
| <b>Laplace transformation</b>   | An integral transform mapping signals $r(t)$ defined in the positive <i>time domain</i> ( $r(t) = 0$ for $t < 0$ ) into the complex domain by multiplying by $\exp(-st)$ and integrating over $t$ from 0 to infinity. See section 5.2.2 of Chapter 5 in this Manual and Scherbaum (2002).                                                                                                                                  |
| <b>lapse time</b>               | Time measured beginning from the earthquake <i>origin time</i> .                                                                                                                                                                                                                                                                                                                                                           |

|                                         |                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|-----------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Late Quaternary faults</b>           | <i>Faults</i> of this age, between the present and 500,000 years before the present, are sometimes considered <i>active</i> based on the observation of historical or <i>palaeoseismic</i> activity on faults of this age in some locales.                                                                                                                                                                                                               |
| <b>lateral blasts</b>                   | Sharing the characteristics of <i>pyroclastic flows</i> and <i>surges</i> , such phenomena have in addition an initial low-angle component of explosive energy release; they are often associated with sudden decompression of a highly pressurized magmatic system upon <i>sector collapse</i> , as was well documented for the 18 May 1980 eruption of Mount St. Helens, Washington State.                                                             |
| <b>lateral-force coefficient</b>        | A term in <i>earthquake engineering</i> referring to a numerical coefficient, specified in terms of a fraction or percentage of the weight of a structure, used in earthquake resistant design to specify the horizontal seismic forces to be resisted by the structure. See Borchardt et al. (2003).                                                                                                                                                    |
| <b>lateral spread and flow</b>          | Terms referring to <i>landslides</i> that commonly form on gentle slopes and that have rapid fluid-like flow movement, like water.                                                                                                                                                                                                                                                                                                                       |
| <b>lateral spread in ground failure</b> | Lateral displacement of the ground down gentle slopes or toward a free face as a consequence of <i>liquefaction</i> accompanied by flow within subsurface granular layers. See Youd (2003).                                                                                                                                                                                                                                                              |
| <b>Laurasia</b>                         | The proto-continent of the northern hemisphere in the late <i>Paleozoic</i> time. It included most of North America, Greenland, and most of Eurasia (excluding India), before they were drifting away from one another. See Duff (1993), and Hancock and Skinner (2000).                                                                                                                                                                                 |
| <b>lava</b>                             | An all-inclusive term for any <i>magma</i> once it is erupted from a <i>volcano</i> ; rocks and deposits formed from lava can vary widely in texture and appearance depending on chemical composition, gas and (or) crystal content, eruption mode (effusive or explosive), and depositional environment.                                                                                                                                                |
| <b>lava cone</b>                        | A relatively small <i>volcanic</i> edifice built largely of <i>lava flows</i> around a central <i>vent</i> , but it is distinguished from a <i>lava shield</i> by having a slightly concave upward profile becoming steeper toward the top — similar to those of <i>cinder cones</i> or <i>stratovolcanoes</i> . Depending on the viscosity of the lava flows involved, there can be a continuum in characteristics between lava cones and lava shields. |
| <b>lava delta</b>                       | A wide, fan-shaped area of new land created by prolonged <i>lava</i> entry — commonly via a <i>lava-tube</i> system — into the ocean, as demonstrated by the 1983-present eruption at Kilauea Volcano                                                                                                                                                                                                                                                    |



|                                             |                                                                                                                                                                                                                                                                                                                               |
|---------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                             | in Hawaii. Such new land is usually built on sloping layers of loose lava fragments and flows.                                                                                                                                                                                                                                |
| <b>lava dome</b>                            | Roughly symmetrical and hemispherical mound of <i>lava</i> formed during extrusion of <i>magma</i> that is so viscous that it accumulates over the <i>vent</i> .                                                                                                                                                              |
| <b>lava flow</b>                            | A stream of molten rock that pours from a <i>volcanic vent</i> during an <i>effusive eruption</i> ; lava flows can exhibit wide differences in size, fluidity, textures, and physical aspects depending on <i>lava</i> viscosity and eruption rates.                                                                          |
| <b>lava fountain</b>                        | A pillar-like, uprushing jet of incandescent <i>pyroclasts</i> that falls to the ground near the <i>vent</i> ; lava fountains can also occur as sheet-like jets from one or more fissure vents (“curtain of fire”). Lava fountains can attain heights of hundred of meters. See <i>Hawaiian eruption</i> .                    |
| <b>lava lake (or pond)</b>                  | An accumulation of molten <i>lava</i> , usually <i>basaltic</i> , contained in a <i>vent</i> , <i>crater</i> , or broad depression. It describes both lava lakes that are molten and those that are partly or completely solidified. If one or more vents sustain a lava lake, it can remain active for many months to years. |
| <b>lava shield</b>                          | A broadly convex, gently sloping volcanic landform built by multiple thin flows of fluid <i>lava</i> from a <i>volcanic vent</i> or a cluster of closely spaced vents. While the term applies to such landforms of any size, the larger lava shields tend to be called <i>shield volcanoes</i> .                              |
| <b>lava tube</b>                            | Lava tubes are natural conduits through which <i>lava</i> travels beneath the surface. They form by the crusting over of lava channels and flows. A broad <i>lava-flow</i> field often consists of a main lava tube and a series of smaller tubes that supply lava to the front of one or more separate flows.                |
| <b>Layer 2 (and 1 of the oceanic crust)</b> | Upper part of the oceanic <i>crust</i> , characterized by <i>velocities</i> lower than about 6.5 km/sec and by high velocity <i>gradients</i> (typically 0.5-1.0 km/sec per kilometer). Layer 1 in this nomenclature is the water layer itself. See Minshull (2002).                                                          |
| <b>Layer 2A (of the oceanic crust)</b>      | A layer of particularly low <i>velocities</i> (typically 2-4 km/sec) found in the vicinity of ocean ridge axes, bounded by a sharp velocity increase at its base, and commonly interpreted as an extrusive <i>basalt</i> layer. See Minshull (2002).                                                                          |
| <b>Layer 3 (of the oceanic crust)</b>       | Lower part of the oceanic <i>crust</i> , characterized by <i>velocities</i> normally in the range 6.5-7.2 km/sec and by low velocity <i>gradients</i> (typically 0.1-0.2 km/sec per kilometer). See Minshull (2002).                                                                                                          |

|                                    |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>leaking modes</b>               | <i>Normal modes</i> in a layered <i>halfspace</i> , in general, have cutoff <i>frequencies</i> below which the <i>phase velocity</i> exceeds the P and/or S velocities of the halfspace, and the energy leaks through the half-space as <i>body waves</i> . Because of the leakage, the <i>amplitude</i> of leaking modes attenuates exponentially with distance. See Aki and Richards (2002 or 2009, p. 312-324).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>Least Significant Bit (LSB)</b> | The smallest change in value that can be represented in the output of an <i>analog-to-digital converter</i> (digitizer).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>least-squares fit</b>           | An approximation of a set of data with a curve such that the sum of the squares of the differences between the observed points and the assumed curve is a minimum.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>left-lateral fault</b>          | A <i>strike-slip fault</i> across which a viewer would see the block on the other side moves to the left. Also known as a <i>sinistral fault</i> . See <i>right-lateral fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>left-lateral movement</b>       | See <i>left-lateral fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>Levi-Civita tensor</b>          | A tensor that is equal to $-\mathbf{I} \times \mathbf{I}$ , where $\mathbf{I}$ is the <i>identity tensor</i> . Its components are known as Ricci symbols. It appears in expressions including vector products.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>Lenz's law</b>                  | An induced electric current in a circuit always flows in a direction so as to oppose the change that causes it. This law was deduced in 1834 by the Russian physicist Heinrich Friedrich Emil Lenz (1804–65).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>Lg waves</b>                    | Originally, short-period (1-6 seconds) large <i>amplitude arrivals</i> with predominantly transverse motion (Ewing et al., 1957). The maximum energy within the Lg-wave group propagates along the surface with <i>velocities</i> around 3.5 km/s, which is close to the average shear velocity in the upper part of the continental <i>crust</i> . Lg waves are observed with frequencies up to 5 or 10 Hz and caused by superposition of multiple <i>S-wave</i> reverberations and SV to P and/or P to SV conversions inside the whole crust. The waves are observed only when the wave path is entirely continental and not disturbed by abrupt changes in crustal thickness or lithological composition along major <i>tectonic</i> boundaries. As little as 2° of intervening ocean can be sufficient to eliminate the waves. See Kulhanek (2002, p. 336-337). For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). |
| <b>lifelines</b>                   | Man-made structures that are important or critical for a community to function, such as roadways, pipelines, power lines, sewers, communications, and port facilities.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>linear analysis</b>             | A calculation or theoretical study that assumes material properties stay within their linear ranges and that the deflections                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |

are small in comparison to characteristic lengths. These assumptions permit the response to be described by linear differential or matrix equations. See Chapter 2 of this Manual and Jennings (2003).

**linearity**

In mathematics, a function  $F$  is linear if  $F(ax) = aF(x)$ , and  $F(x+y) = F(x) + F(y)$ , where  $x$  and  $y$  are variables and  $a$  is a constant. A system is linear if its output can be expressed as the *convolution* of the input and the *impulse response*. When the input is a function of time in a linear system, the output can be obtained from using its frequency response alone. See *convolution* and *frequency response*. See Section 5.7 of Chapter 5 in this manual and Wielandt (2002, p. 297) for testing the linearity of seismometers.

**linearize**

To linearize an equation means to use a linear approximation to that equation, often found by keeping only the first-order terms in the *Taylor expansion*, or, in equivalent linearization, by choosing the parameters of a linear system to minimize the difference between some measure of the responses of a nonlinear system and its linear approximation.

**linear system**

A system which is completely determined by its frequency response. When the frequency response is time-invariant, it is called *stationary linear system*. See *linearity* and *frequency response*.

**liquefaction**

Process by which water-saturated sediment or soil temporarily loses strength and acts as a fluid. The transformation of a granular material from a solid state into a liquefied state is caused by increased pore water pressures and reduced *effective stress*. This effect can be caused by *ground motion* (earthquake shaking) and be associated with *sand boil*. See Youd (2003).

**lithics**

Non-juvenile fragments in *pyroclastic* deposits that represent preexisting solid volcanic or non-volcanic rock incorporated and ejected during explosive *eruptions*.

**lithology**

The description of rock composition and texture.

**lithosphere**

The rigid upper portion of the near-surface thermal boundary layer of the *Earth* at temperatures below about 650°C that can store elastic energy under long-term loads. This definition may or may not agree with the seismic lithosphere, which is composed of the *crust* and the uppermost *mantle* “lid” above the *low velocity layer*. The lithosphere is about 100 km thick, although its thickness depends on age and *heat-flow* conditions (older and cooler lithosphere is thicker). At some locations the lithosphere below the crust is *brittle* enough to produce earthquakes by faulting, such as within subducted oceanic *plates*. See Lay (2002) and Mooney et al. (2002).

|                                           |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|-------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>lithosphere-asthenosphere boundary</b> | The (gradual) transition between the rigid-brittle <i>lithosphere</i> and the ductile (partially molten) <i>asthenosphere</i> in the Earth's <i>upper mantle</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>lithospheric plates</b>                | See <i>tectonic plates</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>lithostatic pressure</b>               | See <i>hydrostatic pressure</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>lithostatic stress</b>                 | There are two variations on the usage of this term. One usage, also called the "lithostatic load" or " <i>lithostatic pressure</i> ", gives the vertical <i>normal stress</i> as a function of depth due to the gravitational force of the overlying rock column. Another usage, the "lithostatic stress state", combines the "lithostatic load" with assumptions about the <i>strain</i> history and rheology of the rock column to produce the complete <i>stress tensor</i> . See Ruff (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>Li-waves</b>                           | These are similar to <i>Lg-waves</i> , but their existence is not as widely accepted as that of <i>Lg</i> . The <i>velocity</i> of Li-waves is 3.8 km/sec (as compared to 3.5 km/sec for <i>Lg</i> ) and may be associated with the lower continental <i>crust</i> (Båth, 1954).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>local magnitude (<math>M_L</math>)</b> | A <i>magnitude</i> scale introduced by Richter (1935) for earthquakes in southern California. $M_L$ was originally defined as the logarithm of the maximum <i>amplitude</i> of <i>seismic waves</i> on a <i>seismogram</i> written by the Wood-Anderson <i>seismograph</i> (Anderson and Wood, 1925) at a distance of 100 km from the <i>epicenter</i> . In practice, measurements are reduced to the standard distance of 100 km by a <i>calibration</i> function established empirically. Because Wood-Anderson seismographs have been out of use since the 1970s, $M_L$ is now computed with simulated Wood-Anderson records and a slightly modified standard formula. The original $M_L$ scale calibration function is only valid for the seismic wave <i>attenuation</i> conditions in California. Therefore, when adapting the $M_L$ scale to other regions, many authors use the original Richter definition together with a modified calibration function. See the biography of Charles Francis Richter in Howell (2003), Chapter 3, DS 3.1 and IS 3.3 in this Manual, the IASPEI (2013) magnitude standards, and Bormann (2011). |
| <b>local site conditions</b>              | Qualitative or quantitative description of the topography, geology and soil profile at a site which affect ground motions at that site during an earthquake. See Chapter 14 of this Manual and Campbell (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>locked fault</b>                       | A <i>fault</i> that is not slipping because frictional resistance on the fault is greater than the <i>shear stress</i> across the fault. Such faults may store <i>strain</i> for extended periods, which is eventually released in an earthquake when frictional resistance is                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |

|                                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|---------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                       | overcome. In contrast, a <i>fault segment</i> which slips sometimes without the occurrence of an earthquake is called creeping segment. See <i>creep events</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>log-periodic behavior</b>          | Power-law (fractal) behavior when the power is not real, but complex. See Turcotte and Malamud (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>longitudinal waves</b>             | Another name for <i>P waves</i> . See section 2.2.3 in Chapter 2 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>long-period</b>                    | In traditional <i>seismometry</i> with limited <i>dynamic range</i> , <i>seismographs</i> were naturally divided into two types; long-period seismographs for <i>periods</i> above the spectral peak (periods around 5 seconds) of the microseismic noise (see <i>microseisms</i> ) generated by ocean waves and short-period seismographs for periods below it. See Chapters 4 and 5 in this Manual.                                                                                                                                                                                             |
| <b>long-period earthquake</b>         | Small, local <i>seismic events</i> occurring under volcanoes with emergent <i>onset</i> of <i>P waves</i> , no distinct <i>S waves</i> , and a low-frequency content (1 to 5 Hz) as compared with the usual <i>tectonic earthquakes</i> of the same <i>magnitude</i> ; also called <i>B-type</i> or <i>low-frequency earthquake</i> . The observed features are attributed to the involvement of fluid such as <i>magma</i> and/or water in the source process. See Chouet (1996a) for the review of related observations and their interpretation, Chapter 13 in this Manual, and McNutt (2002). |
| <b>long (oceanic) wave</b>            | A kind of oceanic gravity wave. When the <i>wavelength</i> is much larger than the water depth, it is called long wave or shallow water wave. Most <i>tsunamis</i> can be treated as long waves. See Satake (2002).                                                                                                                                                                                                                                                                                                                                                                               |
| <b>look vector</b>                    | A vector parallel to the line of sight between a radar and its target on the ground. See Feigl (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>loss</b>                           | In earthquake damage, an adverse economic or social effect (or cumulative effects) of an earthquake (or earthquakes), usually specified as a monetary value or as a fraction or percentage of the total value of a property or portfolio of properties.                                                                                                                                                                                                                                                                                                                                           |
| <b>loss function</b>                  | A mathematical expression or graphical relationship between a specified <i>loss</i> and a specified <i>ground-motion</i> parameter (often the <i>seismic intensity</i> ) for a given structure or class of structures.                                                                                                                                                                                                                                                                                                                                                                            |
| <b>Love-Rayleigh wave discrepancy</b> | Observed <i>phase velocities</i> of <i>Love</i> and <i>Rayleigh waves</i> require a <i>transverse isotropy</i> of seismic velocities in the <i>crust</i> and the uppermost <i>mantle</i> . Love waves are observed to be faster than predicted for <i>isotropic</i> models, which are based on the inversion of Rayleigh waves. This still not totally understood discrepancy was first discovered by Aki and Kaminuma (1963)                                                                                                                                                                     |

for Japan, and has been found universally, both in tectonically active and stable regions. See Cara (2002).

**Love wave**

*SH*-waves trapped near the surface of the Earth and propagating along it. Their existence was first predicted by A. E. H. Love (1911) for a *homogeneous* layer overlying *halfspace* with an *S* velocity greater than that of the layer. They can exist, in general, in vertically *heterogeneous* media, but not in a homogeneous halfspace with a planar surface. See Chapter 2 in this Manual, Malischewsky (1987), Malischewsky and Scherbaum (2004), Kulhanek (2002), and the biography of Augustus Edward Hough Love in Howell (2003).

**lower mantle**

Lowermost part of the Earth's *mantle*. It contains the *D''* zone, which is located directly above the *core-mantle boundary*.

**low-frequency earthquake (LFE)** A term introduced by the Japan Meteorological Agency (JMA) to differentiate a new class of events in their *seismicity* catalog. LFEs occur as part of an extended duration *tremor signal* in *subduction* environments and the visible *arrivals* of LFEs are primarily *S*-waves. *Cross-correlation* methods can be used to accurately locate source regions of LFEs and evidences exist that LFEs originate from shear failure on the *plate interface* in-between locked and slipping portions of the interface (Shelly et al., 2006). See also *non-volcanic tremor* and *long-period earthquake*.

**low-pass filter**

*Filter* which removes the high *frequency* portion of the input *signal*. See IS 5.2 in this Manual and Scherbaum (2002).

**low velocity layer/zone**

Any layer in the Earth in which *seismic wave velocities* are lower than in the layers above and below. The wider low-velocity zone in the *upper mantle* with seismic velocities lower than in the overlying lid of the uppermost layer of the mantle is termed *asthenosphere* and is commonly associated with the presence of partial melt and *ductile* flow. See section 2.5.3.3 and Fig. 2.33 in Chapter 2 of this Manual and Lay (2002).

**Lyapunov exponent**

Solutions to deterministic equations are chaotic if adjacent solutions diverge exponentially in phase space; the exponent is known as the Lyapunov exponent. Solutions are chaotic if the exponent is positive. See *chaos*.

**M****Ma**

An abbreviation for one million years ago (Megaannum).

**macroscopic precursor**

An *earthquake precursor* detected by human sense organs, for example, anomalous animal behavior, gush of well water,

|                                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|--------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                      | earthquake lights, and rumbling sounds. See Rikitake and Hamada (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>macroseismic</b>                  | Pertaining to the observed (felt) effects of earthquakes. See Chapter 12 of this Manual, and Musson and Cekić (2002).                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>macroseismic intensity</b>        | See <i>intensity</i> and <i>intensity scales</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>macroseismology</b>               | The study of any effects of earthquakes that are observable without instruments, such as felt by people, <i>landslides</i> , fissures, knocked-down chimneys, etc. See Chapter 12 in this Manual.                                                                                                                                                                                                                                                                                                                                     |
| <b>magma</b>                         | Molten rock that forms below the Earth's surface; contains variable amounts of dissolved gases (principally water vapor, carbon dioxide, sulfur dioxide) and crystals (principally silicates and oxides).                                                                                                                                                                                                                                                                                                                             |
| <b>magmatism</b>                     | Production of molten igneous rock called " <i>magma</i> ". The magma either erupted at the Earth's surface as <i>lava</i> or <i>ash</i> (volcanism), or trapped in subterranean chambers within the <i>crust</i> or <i>upper mantle</i> as rock bodies termed intrusions (plutonism).                                                                                                                                                                                                                                                 |
| <b>magnetic anomaly</b>              | The difference between the observed value of the magnetic field and the calculated value based on a standard model of the Earth's magnetic field. See Uyeda (2002, p. 55-56).                                                                                                                                                                                                                                                                                                                                                         |
| <b>magnetic lineation</b>            | Stripe patterned <i>magnetic anomalies</i> observed in oceanic areas. They are generally 10-20 km wide and many hundred kilometers long with amplitudes of many hundred nT (nano Tesla). Their origin is explained by the Vine-Matthews-Morley theory (Vine and Matthews, 1963, Morley and Laroche, 1964). See Uyeda (2002), Raff and Mason (1961), Cox (1973), Uyeda (1978), Duff (1993), and Merrill et al. (1998).                                                                                                                 |
| <b>magnetic poles (of the Earth)</b> | They correspond to a point on the Earth's surface where the magnetic inclination is observed to be +90° and -90°, respectively. The North Pole and the South Pole are not exactly opposite each other and their positions wander with time due to changes in the geomagnetic dynamo in the Earth's <i>core</i> . As of 2005 the position of the North Pole was calculated to lie at 82.7°N and 114.4°W, and that of the South Pole at 64.5°S and 137.7°E. See Merrill et al. (1998). See <i>geomagnetic pole</i> , <i>geodynamo</i> . |
| <b>magnetic quiet zone</b>           | An area of the sea-floor without <i>magnetic lineations</i> , because the Earth's magnetic field did not reverse its polarity for a long interval, e.g., during the long Cretaceous normal magnetic epoch. See Chapter 6, and Larson and Pitman (1972).                                                                                                                                                                                                                                                                               |

**magnetic reversal**

A change of the Earth's magnetic field to the opposite polarity that has occurred at irregular intervals during *geologic time*. Polarity reversals can be preserved in sequences of magnetized rocks and compared with standard polarity-change time scales to estimate geologic ages of the rocks. Rocks created along the *oceanic spreading ridges* (*sea-floor spreading*) commonly preserve this pattern of polarity reversals as they cool, and this pattern can be used to determine the rate of ocean-ridge spreading. The reversal patterns recorded in the rocks are termed sea-floor magnetic lineaments. See *magnetic lineation*.

**magnification curve**

A diagram showing the *amplification*, e.g., of the seismic ground motion by a *seismograph*, as a function of *frequency*. See *amplification* and *amplitude response*.

**magnitude, of earthquakes**

In seismology, a quantity intended to measure the size of *earthquakes* and is independent of the place of observation. The Richter magnitude or *local magnitude* ( $M_L$ ) was originally defined by Charles F. Richter (1935) as the logarithm of the maximum *amplitude* in micrometers of *seismic waves* in a *seismogram* written by a standard Wood-Anderson *seismograph* at a distance of 100 km from the *epicenter*. Empirical tables were constructed to reduce measurements to the standard distance of 100 km (see also *magnitude calibration function*) and the zero of the scale was fixed arbitrarily to fit the smallest earthquake then recorded. The concept was extended later to construct magnitude scales based on other data, resulting in many types of magnitudes, such as *body-wave magnitude* ( $m_B$  and  $mb$ ), *surface-wave magnitude* ( $M_s$ ), *moment magnitude* ( $M_w$ ) and *energy magnitude* ( $M_e$ ). In some cases, magnitudes are estimated from seismic *intensity* data, *tsunami* data, or the duration of *coda waves*. The word “magnitude” or the symbol  $M$ , without a subscript, is sometimes used when the specific type of magnitude is clear from the context, or is not really important. According to Hagiwara (1964), earthquakes may be classified by magnitude ( $M$ ) as: *major* if  $M \geq 7$ , as *moderate* if  $M$  ranges from 5 to 7, as *small* if  $M$  ranges from 3 to 5, as *micro* if  $M$  ranges from 1 to 3, and as ultra-micro  $< 1$ . Later usages include: as *nano* if  $M < 0$ , as *great* if  $M \geq 8$  (or sometimes  $7 \frac{3}{4}$ ), and as *mega* if  $M \geq 9$ . In principal, all magnitude scales could be cross calibrated to yield the same value for any given earthquake, but this expectation has proven to be only approximately true, thus the need to specify the magnitude type as well as its value. For details also on other specific types of magnitudes, on IASPEI recommendations for standard magnitude measurement procedures, as well as cross-correlations of magnitude scales see Chapter 3 and IS 3.3 of this Manual, Kanamori (1983), Utsu (2002b), Bormann and Saul (2009), Bormann et al. (2009), Bormann (2011), and IASPEI (2013).



|                                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|---------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>magnitude calibration function</b> | That a <i>magnitude</i> becomes a defined and comparable quantity, which relates to the size or strength of an <i>earthquake</i> independent of the place of observation, one has to correct the related measured values such as <i>amplitude</i> , <i>period</i> , <i>duration</i> , <i>waveform</i> , <i>moment</i> or <i>energy</i> content for <i>path effects</i> ( <i>geometrical spreading</i> , <i>intrinsic attenuation</i> , <i>scattering</i> , ...), source mechanism, and/or recording <i>site effects</i> . These correction terms are then accounted for in a magnitude calibration function. Because of the complexity of all these effects and the difficulty to model them theoretically, especially for short-period data, these “functions” are usually derived as average functions or tabulated values from large empirical data sets. For a commented summary of calibration functions and tables see DS 3.1. |
| <b>mainshock</b>                      | The largest earthquake in an <i>earthquake sequence</i> , sometimes preceded by one or more <i>foreshocks</i> , and almost always followed by many <i>aftershocks</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>major earthquake</b>               | An earthquake with <i>magnitude</i> equal to or greater than 7 is often called “major”. See <i>magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>mantle (Earth’s)</b>               | Zone of the <i>Earth’s</i> interior below the <i>crust</i> and above the <i>core</i> . The mantle represents about 84% of the <i>Earth’s</i> volume and is divided into the <i>upper mantle</i> and the <i>lower mantle</i> , with a transition zone between. See Fig. 2.79 in Chapter 2 and Figs. 11.65 and 11.66 in Chapter 11 of this Manual, Lay (2002), and Jeanloz (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>mantle convection</b>              | The slow overturning of the solid, crystalline <i>mantle</i> by subsolidus <i>creep</i> , driven by internal buoyancy forces. The term subsolidus creep refers to the fact that mantle convection occurs in a crystalline solid. The hot interior of the mantle deforms like an elastic solid on short time scales but like a viscous fluid on <i>geologic time</i> scales. The primary surface expression of mantle convection is <i>plate tectonics</i> and <i>continental drift</i> . See Uyeda (2002) and Olson (2002).                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>mantle magnitude Mm</b>            | The mantle <i>magnitude</i> Mm uses <i>surface waves</i> with <i>periods</i> between about 60 s and 410 s that penetrate into the <i>Earth’s mantle</i> . The concept has been introduced by Brune and Engen (1969) and further developed by Okal and Talandier (1989 and 1990). Mm is firmly related to the <i>seismic moment</i> $M_0$ and does not, or only marginally, saturate even for <i>great</i> , <i>slow</i> or complex <i>earthquakes</i> . Therefore, it is routinely used for rapid magnitude determination of great earthquakes within about 10 min after <i>origin time</i> in the context of <i>tsunami</i> early warning. See Chapter 3 of this Manual, Okal and Talandier (1989 and 1990), Weinstein and Okal (2005).                                                                                                                                                                                             |

## Glossary

- mantle plume** A concentrated *mantle* upwelling that forms a *hot spot* in the Earth's surface. See Uyeda (2002), and Olson (2002). See also *melting anomaly*.
- mantle Rayleigh waves (R waves)** Just as long-period *Love waves* are given another name, *G-waves*, long-period *Rayleigh waves* are sometimes called "mantle Rayleigh waves" or R waves. See Figs. 2.20 and 2.21 in Chapter 2 of this Manual.
- marine microseisms** An ever-present, nearly periodic *seismic signal* originating from ocean waves and surf. Its *amplitude* is typically between 0.1 and 10  $\mu\text{m}$  and its *period* between 4 s and 8 s. The marine microseisms divide the spectrum of seismic signals into a short-period and a long-period band. See *double frequency microseisms* and *microseisms*.
- Maslov asymptotic ray theory** The traditional *ray theory* breaks down at *caustics* as the rays are focused and the equation for the *amplitude* predictions becomes singular. V. P. Maslov's asymptotic ray theory (1965) generalizes the ray theory by using an *ansatz*, which integrates over neighboring rays with appropriate phase delays. The Maslov asymptotic ray theory remains valid at caustics but reduces to the original ray theory when the latter is valid. See Chapman (2002).
- master model of earthquakes** A concept developed at the Southern California Earthquake Center to use as a framework for integrating multi-disciplinary earth science information regarding earthquakes in Southern California for the purpose of transmitting it to people living there. A first generation master model for southern California is described in WGCEP (1995). See Aki (2002).
- maximum credible earthquake (MCE)** The maximum earthquake, compatible with the known tectonic framework, that appears capable of occurring in a given area. See Faccioli and Pessina (2003).
- maximum probable earthquake (MPE)** The maximum earthquake that could strike a given area with a significant probability of occurrence. See Faccioli and Pessina (2003).
- Maxwell material** A theoretical viscoelastic material in which the short-time response to an applied force is that of an elastic solid, while the long-time response is that of a viscous fluid. See also *viscoelasticity*.
- Maxwell's equations** A set of four partial differential equations that describe the large-scale electromagnetic phenomena by the behavior of the electric and magnetic vectors and how the electric and magnetic fields are related to each other. These equations were formulated by James Clerk Maxwell (1831-1879) based on the

experimental results of Ampere, Coulomb, Faraday, and Lenz. See Maxwell (3<sup>rd</sup> edition, 1991), Stratton (1941).

**mb** See *body-wave magnitude*.

**mB** See *body-wave magnitude*.

**MCS scale** See *Mercalli-Cancani-Sieberg scale*.

**Medvedev-Sponheuer-Karnik scale** A *macroseismic* twelve-degree intensity scale (abbreviated MSK). It was based on MCS, MM56 (see *intensity scales*) and previous work by Medvedev (1962) in Russia. Published first in 1964 (Sponheuer and Karnik, 1964), it received several later minor modifications. The MSK developed greatly the quantitative aspects of the scale, thus making it more powerful than its predecessors. It became widely used in Europe. See Chapter 12 of this Manual and Musson et al. (2010).

**melting anomaly** Another name for *hot spot*. A region on Earth (often in an *intraplate tectonic* setting) of anomalously high rates of volcanism over many millions of years, presumably sustained by a persistent source of hot *mantle* material; the Hawaiian melting anomaly (or hot spot) is arguably the best example.

**MEMS devices** Sensor devices based on microelectromechanical systems. Originally developed for use in airbags, they are nowadays built-in in smartphones, harddisks, Wii stations, etc. Of special interest for *seismic exploration*, *seismology* and *engineering seismology* are *accelerometers* based on MEMS technology. Mechanical components and required electronic circuits are often integrated on a silicon chip that means small dimensions and often low costs. Low g chips have a resolution of about 1 mg at costs of a few \$. They can be used for *strong-motion* sensing (e.g., in earthquake *early warning systems* such as SOSEWIN, Fleming et al., 2009);) There is also some progress in the development of high resolution accelerometers for different applications. The resolution is up to ~100ng/sqrt(Hz). Other applications are in strong-motion seismic sensing with high *dynamic range* of 120dB or more and in *seismic prospection* where MEMS accelerometers combined with 24 bit digitizer modules result in a modern replacement for classical *geophones*. Costs of these high resolution sensors are still high (several 100\$ per unit), due to a more expensive technology. It is not likely that MEMS can replace *broadband seismometers* in near future since their sensitivity at very low *frequencies* is more than 2 orders less than for broadband velocity-meters.

**Mercalli-Cancani-Sieberg scale** A *macroseismic* twelve-degree *intensity scale* (abbreviated MCS). It is essentially based on the earlier developed and upgraded twelve-degree scale by August Sieberg

(Sieberg, 1912, 1923). Sieberg's version of 1923 has been translated into English by Wood and Neumann (1931), becoming the *Modified Mercalli Scale (MM Scale)*.

|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
|---------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Mesozoic</b>                 | An era in <i>geologic time</i> , from 245 to 66.4 million years ago.                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>metamorphic facies</b>       | A set of metamorphic rocks characterized by particular mineral associations, indicating origin under a specific pressure-temperature range.                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>metamorphic rock</b>         | A rock derived from pre-existing rocks by mineralogical, chemical, and/or structural changes, essentially in the solid-state, in response to changes in pressure, temperature, shearing stress, and chemical environment.                                                                                                                                                                                                                                                                                                               |
| <b>meter (m)</b>                | A base unit of length in the <i>international system of units (SI)</i> . The official definition is “The meter is the length of the path traveled by light in vacuum during a time interval of 1/299792458 of a second”. See Lide (2002).                                                                                                                                                                                                                                                                                               |
| <b>Metropolis method</b>        | A general method for sampling a probability, including high-dimensional spaces. See Mosegaard and Tarantola (2002).                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>microcontinent</b>           | An isolated, relatively small mass of sialic <i>crust</i> surrounded by oceanic crust. It is commonly plastered onto continents at a <i>convergent margin</i> where it becomes a part of the orogenic belt identifiable as an exotic terrane or “suspect terrane.”                                                                                                                                                                                                                                                                      |
| <b>microcrack</b>               | An ensemble of tiny cracks that can join to form a crack. See Teisseyre and Majewski (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>microearthquake</b>          | An <i>earthquake</i> that is not perceptible by man and can be recorded by <i>seismographs</i> only. Typically, a microearthquake has a magnitude of 2 or less on the Richter scale.                                                                                                                                                                                                                                                                                                                                                    |
| <b>micromorphic (continuum)</b> | A continuum in which each material point is equipped with a kind of micro-structure, described in a local system of coordinates, whereby each of its “points” comprises a microvolume with micro- <i>displacements</i> , micro- <i>rotations</i> , micro- <i>deformations</i> , and micro-inertia tensor (see <i>tensor of inertia</i> ). A number of material constants permit the description and solution of many complicated problems in which materials display non-classical responses to applied external loads (Eringen, 1999). |
| <b>micropolar (continuum)</b>   | A synonym for the <i>Cosserat continuum</i> (Pujol, 2009). A continuum defined similarly to the <i>micromorphic continuum</i> , but for which only micro- <i>displacements</i> and micro- <i>rotations</i> of the microelements are permitted, but not micro- <i>deformations</i> , i.e., the microelements are taken to be rigid. Each material point is equipped with three degrees of freedom for rigid rotation in                                                                                                                  |

addition to the classical translational degrees of freedom (Eringen, 1999).

**microseisms**

Continuous *ground motion* constituting the background *noise* for any seismic measurement. Microseisms with *frequencies* higher than about 1 Hz are usually caused by artificial sources, such as traffic and machinery, and are sometimes called “microtremors”, to be distinguished from longer-period microseisms due to natural disturbances. At a typical station in the interior of a continent, the microseisms have predominant *periods* of about 6 seconds. They are caused by surf pounding on steep coastlines and the pressure from standing ocean waves, which may be formed by waves traveling in opposite directions in the source region of a storm or near the coast (Longuet-Higgins, 1950). Also called *double frequency microseism* and *marine microseism*. See Chapter 4 in this Manual, Webb (2002), Kulhanek (2002); also Aki and Richards (2002 or 2009), p. 616-617).

**microzonation**

The identification and mapping at local or site scales of areas having different potentials for hazardous earthquake effects, such as *ground shaking intensity*, *liquefaction* or *landslide* potential.

**mid-ocean ridge**

An undersea mountain range extending through the Arctic Ocean, the North and South Atlantic Oceans, the Indian Ocean, and the South Pacific Ocean, and consists of many small and slightly offset segments. Each segment is characterized by a central *rift* where two *tectonic plates* are being pulled apart and new oceanic *lithosphere* is being created. See Cox (1973), Uyeda (1978), Duff (1993), Hancock and Skinner (2000), and Uyeda (2002).

**mitigation**

The proactive use of information, land use, technology, and societies' resources to prevent or minimize the effects of disasters on society. See O'Brien and Mileti (2003).

**mixed dislocation**

See *dislocation*.

**Mm**

See *mantle magnitude*.

**modal**

A term referring to natural modes of vibration or their properties, for example, modal *damping*. See Jennings (2003).

**mode 1 crack**

An opening crack in which *displacement* is directed normal to the crack surface. See section 4 and Figure 9 in Ben-Zion (2003).

**mode 2 crack**

A shear crack in the plane strain mode. See *plane strain*, and section 4 and Figure 9 in Ben-Zion (2003).

## Glossary

|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|----------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>mode 3 crack</b>              | A shear crack in the anti-plane strain mode. See <i>anti-plane strain</i> , and section 4 and Figures 9 - 10 in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>moderate earthquake</b>       | An earthquake with magnitude that ranges from 5 to 7 is often called “moderate”. See <i>magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>mode shape</b>                | The shape a structure assumes when it oscillates solely at one of its <i>natural frequencies</i> . Mathematically, a mode shape is an eigenvector of the eigenvalue problem posed by linear, free vibration of a structure. See Jennings (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>Modified Mercalli scale</b>   | A <i>macroseismic</i> twelve-degree <i>intensity scale</i> (I to XII) published by Wood and Neumann (1931) (abbreviated MM). It is based on the translation into English by Wood and Neumann (1931) of the macroseismic scale published by Sieberg (1923), which in its later version (Sieberg, 1932) became widely known as the <i>Mercalli-Cancani-Sieberg scale</i> . Therefore, the name Modified Mercalli scale is inappropriately chosen. Even more, after Richter (1958) overhauled the Wood and Neumann version completely, thus becoming the “Modified Mercalli Scale of 1956” (MM56), which is only very remotely linked to Mercalli. Reference to the MM scale is further complicated by recent modifications introduced by Stover and Coffman (1993) to the 1931 version of the MM in conjunction with the revision of the <i>seismicity</i> of the United States between 1568 and 1989. This latest upgrade is now referred to as <i>MMI</i> . The MM scale is widely used, not only in the USA (although MMI has become there now most popular because of the “Did you feel it?” system maintained by the USGS), but also in many countries in and outside of North America. For details see Chapter 12 of this Manual, Musson et al. (2010) and Atkinson and Wald (2007). |
| <b>MM scale</b>                  | See <i>Modified Mercalli scale</i> . A later revision by Richter (1958) became known as the version MM56.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>MMI scale</b>                 | Recently modified version of the <i>MM scale</i> proposed by Stover and Coffman (1993). See <i>Modified Mercalli seismic intensity scale</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>Moho</b>                      | The abridged “nickname” for <i>Mohorovičić discontinuity</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>Mohorovičić discontinuity</b> | A <i>discontinuity</i> in seismic <i>velocities</i> that defines the boundary between <i>crust</i> and <i>mantle</i> of the Earth. Named after the Croatia seismologist Andrija Mohorovičić (1857-1936) who discovered it. The boundary is at a depth of 20 to 60 km beneath the continents and 5 to 10 km beneath the ocean floor. See Fig. 2.11 in Chapter 2 of this Manual, Mooney et al. (2002), Minshull (2002), and the biography of Andrija Mohorovičić in Howell (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |

|                                         |                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
|-----------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Mohr diagram</b>                     | A two dimensional diagram of a particular plane (Mohr plane) of the three dimensional <i>stress</i> state. The Mohr plane contains two of the <i>principal stresses</i> $\sigma_i > \sigma_j$ , and depicts the shear and <i>normal stresses</i> on planes of all possible orientation whose normals lie in the Mohr plane. Details on use and construction of Mohr diagrams can be found in Jaeger and Cook (1971). See Lockner and Beeler (2002). |
| <b>moment connection</b>                | A connection between elements of a structure that is capable of transmitting moment as well as shear force. In a <i>moment frame</i> , the joints between the beams and columns are typically designed to take the full moment developed in the beam. See Hall (2003), Dowrick (1977), and Salmon and Johnson (1996).                                                                                                                               |
| <b>moment frame</b>                     | An unbraced planar frame. Moment frames resist lateral forces by flexure of the comprising beams and columns. See Hall (2003), and Dowrick (1977), Park and Paulay (1975), and Salmon and Johnson (1996).                                                                                                                                                                                                                                           |
| <b>moment magnitude (M<sub>w</sub>)</b> | A non-saturating <i>magnitude</i> computed using the <i>scalar seismic moment</i> ( $M_0$ ). It was introduced by Kanamori (1977) via the Gutenberg-Richter magnitude-energy relation. See Chapter 3 in this Manual and Bormann and Di Giacomo (2011) for details, also with respect to the relationship of $M_w$ to the <i>energy magnitude</i> $M_e$ and the problem of <i>saturation (of magnitude scales)</i> .                                 |
| <b>moment of force</b>                  | The tendency of a force applied to an object to cause the object to rotate about a given point, and this tendency is expressed by the equation, $\mathbf{N} = \mathbf{r} \times \mathbf{F}$ , where $\mathbf{N}$ is the <i>torque</i> , $\mathbf{r}$ is the position vector from the origin to the point of application of force, and $\mathbf{F}$ is the total force acting on the point.                                                          |
| <b>moment of inertia</b>                | A component of <i>tensor of inertia</i> $J$ , calculated as $\mathbf{k}^* \mathbf{J}^* \mathbf{k}$ , is called moment of inertia relative to the axis with unit vector $\mathbf{k}$ .                                                                                                                                                                                                                                                               |
| <b>moment tensor</b>                    | A symmetric second-order tensor that characterizes an internal <i>seismic point source</i> completely. For a finite source, it represents a point source approximation and can be determined from the analysis of <i>seismic waves</i> whose <i>wavelengths</i> are much greater than the source dimensions. See IS 3.9 of this Manual, Aki and Richards (2002 or 2009 p. 49-58), Sipkin (2002), and Section 2 in Ben-Zion (2003).                  |
| <b>monitoring system</b>                | A system for monitoring <i>earthquakes</i> , <i>volcanic eruptions</i> , <i>tsunami</i> , and/or other phenomena, usually consisting of a network of <i>seismic stations</i> and/or <i>arrays</i> , sometimes complemented by other types of sensors.                                                                                                                                                                                               |
| <b>monogenetic volcano</b>              | A small volcano that is of a type that generally has a single <i>eruption</i> , or short-lived eruptive sequence, during its lifetime.                                                                                                                                                                                                                                                                                                              |

|                                            |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|--------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Monte Carlo method</b>                  | Any mathematical method using at its core a random (or, more frequently, pseudo-random) generation of numbers. Monte Carlo methods are typically used to randomly explore high-dimensional spaces, where a systematic exploration is unfeasible. See Mosegaard and Tarantola (2002) and Tarantola (2005).                                                                                                                                                                             |
| <b>moveout</b>                             | The time difference between <i>arrivals</i> (such as P) at different stations, or between different arrivals at the same stations (like P and pP), which is also known as <i>step-out</i> .                                                                                                                                                                                                                                                                                           |
| <b>M<sub>s</sub></b>                       | See <i>surface-wave magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>MSK scale</b>                           | See <i>Medvedev-Sponheuer-Karnik scale</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>M<sub>t</sub></b>                       | See <i>tsunami magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>mudflow</b>                             | Essentially synonymous with volcanic debris flow and <i>lahar</i> .                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>multi-degree of freedom</b>             | A term applied to a structure or other dynamical system that possess more than a single degree of freedom. The number of degrees of freedom of a complex system is equal to the number of independent variables needed to fully describe the spatial distribution of its response at any instant. See Jennings (2003).                                                                                                                                                                |
| <b>multiple lapse time window analysis</b> | A method to measure <i>scattering</i> loss and <i>intrinsic absorption</i> from the analysis of whole <i>S-wave seismogram</i> envelopes. This method is based on the radiative transfer theory for S-wave propagation through <i>scattering</i> media. See Sato et al. (2002), and Sato and Fehler (1998, p. 189).                                                                                                                                                                   |
| <b>M<sub>w</sub></b>                       | See <i>moment magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>mylonite</b>                            | A cohesive, penetratively foliated (and usually lineated) <i>fault-rock</i> (L-S tectonite) with at least one of the major mineral constituents (usually quartz) having undergone grain-size reduction through dynamic recrystallization accompanying crystal plastic flow. The mylonite series (protomylonite - mylonite - ultramylonite) reflects progressive reduction in grain size, generally under greenschist facies <i>metamorphic</i> conditions. See Sibson (2002, p. 458). |
| <b>mylonitic gneiss</b>                    | A comparatively coarse-grained, banded mylonitic <i>fault-rock</i> (L-S tectonite) developed in a high-strain ductile shear zone under amphibolite facies <i>metamorphic</i> conditions, in which most of the mineral constituents have undergone dynamic recrystallization. See Sibson (2002).                                                                                                                                                                                       |



## N

|                                        |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|----------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Nakamura method</b>                 | See <i>H/V spectral ratio</i> and <i>spectral ratio</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>natural frequency</b>               | The discrete <i>frequency</i> at which a particular elastic object or system vibrates when it is set in motion by a single impulse and not influenced by other external forces or <i>damping</i> . The reciprocal of <i>fundamental period</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>natural period</b>                  | The reciprocal of <i>natural frequency</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>N discontinuity</b>                 | A seismic boundary within the continental <i>lithosphere</i> at a depth of 60-120 km reported particularly in Eurasia. Also called “ <i>Hales discontinuity</i> ”. See Mooney et al. (2002, p. 898).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>Nakamura method</b>                 | See <i>H/V spectral ratio</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>near field</b> and <b>far field</b> | A seismic source radiates different types of seismic waves with different <i>amplitude</i> decays. Some of these seismic waves can only be observed in the vicinity of the source. In theoretical seismology usually all seismic wave types with an amplitude decay of $r^{-1}$ , are called “far-field” terms, while those with a amplitude decay of $r^{-2}$ are called “near-field” terms. If the <i>hypocentral distance</i> is large enough for the “near field” terms to become negligible one is in the “far field”. The exact hypocentral distance, which separates “far field” and “near field”, depends on the <i>frequency</i> of the seismic waves. Usually a distance of a few <i>wavelengths</i> from the source is considered as “far field”, and a distance within a fraction of a wavelength is “near field”, with intermediate distances requiring the examination of the effects of the individual terms (Aki and Richards, 1980, p. 88 and eqns. 4.23 and 4.35). However, in geology and earthquake engineering a simpler approach is taken: Locations more than about ten source dimensions from the source are in the <i>far field</i> , and those within one to several dimensions are in the <i>near field</i> . |
| <b>neotectonics</b>                    | The study of the post-Miocene structures and structural history of the Earth’s <i>crust</i> . The Miocene ended about 5 million years ago.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>Newmark analysis</b>                | A numerical technique that models a potential <i>landslide</i> as a rigid block resting on a frictional slope, describing dynamic forces on the block from assumed <i>ground shaking</i> records in order to calculate the expected <i>displacement</i> of the block.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>newton (N)</b>                      | The SI derived unit for force. It is defined as the force which accelerates a mass of 1 <i>kilogram</i> to 1 <i>meter per second</i> per                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |

second, i.e.,  $1 \text{ N} = 1 \text{ kg m s}^{-2}$ . It is named after Isaac Newton (1642-1727) who formulated the three laws of motion.

**noble gas**

A gas in group 0 of the periodic table of the elements (Neon, Argon, Krypton, Xenon, and Radon); it is monatomic and chemically inert. See King and Igarashi (2002).

**node**

A reference point, typically a corner, in *finite element* analysis. Because the variation of field quantities of interest, such as *displacement*, *strain*, *velocity*, etc., are assumed to vary within finite elements in specified ways, their values become functions of values of key variables at the nodes. See *finite element* and Hall (2003).

**noise**

In general: Unwanted part of data superimposed on a wanted *signal*. Noise is often modeled as a stationary random process, but may also contain signal-dependent components. One man's noise can be the other man's signal. More specific in seismology: Incoherent, natural or artificial perturbations caused by a diversity of agents and distributed sources. One usually differentiates between *ambient* background *noise* and *instrumental noise*. The former is due to natural (ocean waves, wind, rushing waters, animal migration, ice movement, etc) and/or man-made sources (traffic, machinery, etc.), whereas instrumental (internal) noise may be due to the “flicker” noise of electronic components and/or even Brownian molecular motions in mechanical components. Digital *data acquisition* systems may add *digitization noise* due to their finite discrete resolution (*least significant bit*). Very sensitive seismic recordings may contain all these different noise components, however, usually their resolution is chosen so that only *seismic signals* and to a certain degree also the ambient noise are resolved. Disturbing noise can be reduced by selecting recording sites remote from noise sources, by underground installation of seismic sensors (e.g., in boreholes, tunnels or abandoned mines) or by suitable *filter* procedures (improvement of the *signal-to-noise ratio*). See Chapter 4 of this Manual, Webb (2002), and Wielandt (2002).

**noise level** (of an instrument)

See *self noise*.

**non double-couple earthquake**

An earthquake produced by a source that includes *displacement* components other than simple shear displacements along a *fault plane* (a *double-couple* earthquake). Common examples include the abrupt extensional opening of a crack, the collapse of a cavity or a *compensated linear vector dipole* (CLVD). See Julian et al. (1998).

**non-linear analysis**

In *earthquake engineering*, a calculation or theoretical development that includes material changes such as yielding or cracking, or the non-linear effects of large *displacements* such

|                                        |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
|----------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                        | as the increased <i>moment</i> at the base of a structure caused by lateral displacements of floor masses. These effects cannot be considered within the framework of linear equations. See Jennings (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>nonstructural</b>                   | The parts of a building system that do not support loads (e.g., partitions, exterior claddings, lighting fixtures, etc.).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>non-volcanic tremor</b>             | Non-volcanic tremor is characterized by long lasting wave-trains of low-frequency components between 1 and 10 Hz and lack of distinct <i>P</i> or <i>S</i> phases. Unlike the frequently observed deep, low-frequency micro-earthquakes beneath active volcanoes, non-volcanic tremors are extremely large phenomena characterized by long time duration from hours to weeks and a wide source area, over 600 km in length (Obara et al., 2004). Non-volcanic tremor was first detected within <i>subduction zones</i> in southwest Japan and Cascadia and appears to be related to <i>slow-slip events</i> in these regions. Similar tremor is also observed in other <i>fault</i> settings such as the San Andreas fault system. Non-volcanic tremor is often associated with <i>aseismic slip</i> . See also <i>low frequency earthquakes</i> , <i>slow slip-events</i> , <i>episodic tremor and slip</i> .                                                                                                                                                                                                     |
| <b>normal distribution (or errors)</b> | Synonym for <i>Gaussian distribution</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>normal fault</b>                    | A <i>fault</i> that involves lateral extension, where one block (the <i>hanging wall</i> ) moves over another (the <i>footwall</i> ) down the <i>dip</i> of the <i>fault plane</i> , i.e., the movement of the hanging wall is downward relative to the footwall. The maximum compressive stress is vertical; the least compressive stress is horizontal. See <i>reverse fault</i> ; also Figs. 3.105 and 3.106 in Chapter 3 of this Manual, Jackson (2002), and Figure 13 of Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>normal modes</b>                    | Normal modes were originally defined as the linear, free vibrations of a system with a finite number of degrees of freedom, such as a finite number of mass particles connected by massless springs. Each mode is a simple harmonic vibration at a certain frequency called an “ <i>eigenfrequency</i> ” or <i>natural frequency</i> . There are as many independent modes as the number of degrees of freedom. An arbitrary motion of the system can be expressed as a superposition of normal modes. Free vibrations of a finite continuum body, such as the Earth, are also called normal modes. In this case, there are an infinite number of normal modes, and an arbitrary motion of the body can be expressed by their superposition. The concept of normal modes has been extended to wave-guides in which free waves with a certain <i>phase velocity</i> can exist without external force. Examples are <i>Rayleigh waves</i> in a <i>halfspace</i> and <i>Love waves</i> in a layered halfspace. In these cases, however, one cannot express an arbitrary motion by superposing normal modes. See Udias |

(2002, p. 90-92), Lognonne and Clevede (2002), and Aki and Richards (2002 or 2009, Chapters 7 and 8).

**normal stress** Force per unit area exerted upon an infinitesimal area in the direction of the normal to that area. In most disciplines, tension across the area is reckoned positive; in geology, the opposite convention may be used. See also *shear stress*, Ruff (2002), and Figure 1 in Ben-Zion (2003).

**nuée ardente** A French term (“glowing cloud”) that was first used to describe the deadly 1902 eruption of Montagne Pelée (Island of Martinique) that devastated the port city of St. Pierre (Lacroix, 1904). A *nuée ardente* is a density current consisting of hot *pyroclastic* fragments, volcanic gases, and air that flows at high speed from the *crater* region of a volcano — generally associated with the collapse of an eruption column or unstable *lava dome*. See *glowing avalanche*, *pyroclastic flow*, and *pyroclastic surge*.

**Nyquist frequency** Half of the digital sampling rate. It is the minimum number of counts per second needed to define unambiguously a particular *frequency*. If the *seismic signal* contains energy in a frequency range above the Nyquist frequency the signal distortions are called *aliasing*.

## O

**oblique slip** A combination of *strike slip* and *normal slip* or *reverse slip*.

**occurrence rate** See *frequency of occurrence*.

**ocean bottom seismometer/hydrophone (OBS/OBH)** A receiver for seismic signals which is located at or near the seabed. An OBH is commonly moored a few meters above the seabed and has a single *hydrophone* sensor. An OBS commonly rests on the seabed and has in addition to a hydrophone a three-component *seismograph*, which may be inside the recording package or deployed as a separate package for better coupling. See sub-Chapter 7.5 of Chapter 7 in this Manual, and Minshull (2002).

**oceanic spreading ridge** A fracture zone along the ocean bottom that accommodates upwelling of *mantle* material to the surface, thus creating new *crust*. A line of ridges, formed as molten rock reaches the ocean bottom and solidifies, topographically marks this fracture. See *mid-ocean ridge* and *sea-floor spreading*.

**oceanic trench** A narrow and elongated deep depression of the sea-floor associated with a *subduction zone*. See Uyeda (1978), and Duff (1993).

- octave (filtering)** A term coming from music, defining distance between the eighth full tone above (or below) a given reference tone. The *frequency* doubles (or halves) between two tones separated by one octave. *Seismographs* typically record (filter) *ground motion* oscillations within a range of one (narrow-band) to 12 octaves (very broadband).
- Okada model** Closed form of analytic solutions for the static *displacement* field due to shear and tensile *faults* in a *homogeneous halfspace*. It has become a standard model used for calculating the displacement field for earthquake *fault slips* and *pressure* sources beneath volcanoes. See Okada (1985), and Feigl (2002, p. 612-613) with a computer program included on the attached Handbook CD#1, under the directory \37Feigl.
- omega-squared model** A widely used earthquake source model associated with the *far-field body-waves*. It is based on the *displacement amplitude spectrum* and is characterized by having a flat low-frequency part and a *power law* decay (with the power of -2) in the high-frequency part, separated by the *corner frequency*. The level of the flat low-frequency part is proportional to the *seismic moment* and the corner frequency is inversely proportional to the linear dimension of the source. See Hanks (1979) and Aki and Richards (2002, Chapters 10 and 11) for its observational and physical bases. The observed seismic spectra tend to be lower than predicted by this model at frequencies higher than a limit called *fmax*. A practical procedure for computing the time history of strong ground motion, including the *fmax* effect, was given by Boore (1983) based on this model. See *Brune model*, *seismic moment*, *corner frequency*, *fmax*, Fig. 3.5 in Chapter 3 of this Manual or on p. 1871 in Bormann et al. (2009), and Eq. (6.1) in Ben-Zion (2003).
- Omori's relation (Omori's law)** A relation expressing the temporal decay of the *frequency of occurrence* of *aftershocks*. Occasionally called “hyperbolic law”. The modified Omori relation generalizes it to a *power law* dependence on time with the power allowed to deviate from -1. See Utsu (2002b), and Eq. (7.4) in Ben-Zion (2003), and the biography of Fusakichi Omori in Howell (2003).
- one-dimensional amplification of seismic waves** In *site effect* studies a one-dimensional structure means a structure varying only in the vertical direction. It is also called a horizontally-layered structure. The *amplification* for incident *body waves* can be theoretically calculated by considering multiple transmission and *reflection* at each interface of horizontal layers including the free surface. There is no amplification for *surface waves* incident on such a structure since it is assumed to be *homogeneous* in the horizontal directions. See Kawase (2003).

## Glossary

|                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|--------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>onset (of seismic wave)</b> | The first appearance of an seismic <i>wavelet</i> on a record.                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>ophiolite</b>               | A suite of mainly ultramafic and mafic rocks formed at <i>mid-ocean spreading centers</i> and along zones of convergence of oceanic crustal <i>plates</i> . On-land exposures of tectonically accreted sections of ophiolite are a primary means of studying the rocks that make up the oceanic <i>crust</i> and uppermost <i>mantle</i> . See Minshull (2002) and Dickinson (2002).                                                                            |
| <b>optimal survey design</b>   | Optimal design of experiment or survey based on the <i>inverse</i> theory. See Curtis and Snieder (2002, p. 868-870).                                                                                                                                                                                                                                                                                                                                           |
| <b>organized noise</b>         | Seismic <i>noise</i> whose power spectrum is not uniformly distributed, but is concentrated at particular <i>wavenumbers</i> and <i>frequencies</i> . See Douglas (2002).                                                                                                                                                                                                                                                                                       |
| <b>origin time</b>             | The instant of time when an earthquake begins at the <i>hypocenter</i> . It is usually determined from <i>arrival times</i> of seismic waves by the <i>Geiger's method</i> .                                                                                                                                                                                                                                                                                    |
| <b>orogenesis</b>              | See <i>orogeny</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>orogeny</b>                 | The process of crustal uplift, folding, and faulting by which systems of mountains are formed. See Uyeda (2002, p. 62-64).                                                                                                                                                                                                                                                                                                                                      |
| <b>orthogonal regression</b>   | See <i>regression analysis</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>orthorhombic symmetry</b>   | Symmetry with respect to three mutually orthogonal planes. There are 9 independent elastic parameters for an orthorhombic material. Olivine crystals are orthorhombic. See Cara (2002).                                                                                                                                                                                                                                                                         |
| <b>oscillator</b>              | A mass that moves with oscillating motion under the influence of external forces and one or more forces that restore the mass to its stable at-rest position. In <i>earthquake engineering</i> , an oscillator is an idealized damped mass-spring system used as a model of the <i>response</i> of a structure to earthquake <i>ground motion</i> . A <i>seismograph</i> is also an oscillator of this type. See <i>single-degree-of-freedom</i> (SDOF) system. |
| <b>outer rise</b>              | Upward flexure of the oceanic <i>plate</i> before subducting beneath the <i>trench</i> .                                                                                                                                                                                                                                                                                                                                                                        |
| <b>outer arc ridge</b>         | A zone landward from the trace of the <i>subduction thrust fault</i> of elevated sea floor probably related to the compression of the rocks in the <i>accretionary wedge</i> . Also referred to as the outer arc high.                                                                                                                                                                                                                                          |
| <b>oversampling</b>            | A sampling technique in which the analog input <i>signal</i> is first sampled at a sampling rate much higher than the final sampling rate and subsequently decimated digitally. See Chapter 6 of this Manual and Scherbaum (2002).                                                                                                                                                                                                                              |

## P

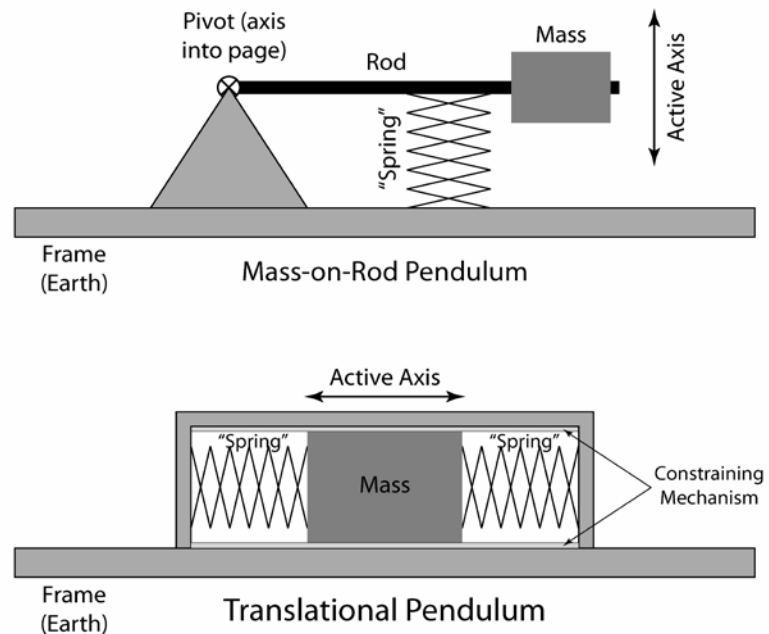
|                           |                                                                                                                                                                                                                                                                                                                                                                                       |
|---------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>P</b>                  | General symbol for a <i>P-wave</i> . For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).                                                                                                                                                                                                                |
| <b>P'</b>                 | Alternative phase name for PKP. For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).                                                                                                                                                                                                                     |
| <b>pahoehoe</b>           | A common type of fluid <i>lava flow</i> produced by <i>Hawaiian eruptions</i> , pahoehoe in solidified form is characterized by a continuous, smooth, billowy or ropy surface. Like <i>aa</i> , the term is also of Hawaiian origin; both <i>aa</i> and pahoehoe have become universally accepted to describe lava flows of similar appearance and characteristics in the world over. |
| <b>paleoearthquake</b>    | An ancient, usually prehistoric earthquake.                                                                                                                                                                                                                                                                                                                                           |
| <b>paleomagnetics</b>     | The study of the geomagnetic field existing at the time of formation of a rock, based on the remanent magnetization that was induced in the rock at the time of its formation and is still preserved today. See Uyeda (2002, p. 52-54), and Merrill and McFadden (2002).                                                                                                              |
| <b>paleomagnetic pole</b> | An estimate of the geomagnetic pole of the Earth at a certain time in the past by averaging the paleomagnetic field over a sufficient long period of time, e.g., $10^4$ to $10^5$ years, from rocks which acquired the paleomagnetic field at that time. See Uyeda (2002), and Merrill et al. (1998).                                                                                 |
| <b>paleomagnetism</b>     | All rocks exhibit magnetic properties imparted to them by the Earth's magnetic field at the time they were formed, and preserved to the present time as remanent magnetism. Paleomagnetism is the study of these properties, or the properties themselves. See Merrill and McFadden (2002).                                                                                           |
| <b>palaeoseismic</b>      | Referring to the prehistoric seismic record as inferred from ruptural <i>displacements</i> in young geologic sediments in combination with $^{14}\text{C}$ age dating.                                                                                                                                                                                                                |
| <b>paleoseismic event</b> | A <i>paleoearthquake</i> that caused surface faulting, <i>displacements</i> in young sediments, or other near-surface phenomena such as <i>liquefaction</i> . See Grant (2002).                                                                                                                                                                                                       |
| <b>paleoseismology</b>    | The study of (usually prehistoric) <i>paleoearthquakes</i> , i.e., centuries or millennia after their occurrence. See Grant (2002), and McCalpin and Nelson (1996).                                                                                                                                                                                                                   |

|                                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
|------------------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Paleozoic</b>                               | An era in <i>geologic time</i> , from 545 to 245 million years ago.                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>Pangea</b>                                  | A supercontinent (including most of the continental <i>crust</i> of the Earth) that was postulated by Alfred Wegener in 1912 to have existed from about 300 to 200 million years ago. Pangea is believed to have split into two protocontinents – <i>Gondwana</i> in the south and <i>Laurasia</i> in the north, from which the modern continents are.                                                                                                                                                                                  |
| <b>parabolic approximation</b>                 | An approximation to the wave equation in which the second derivative with respect to the global <i>ray</i> direction is neglected. It can be justified when the <i>wavelength</i> is much smaller than the characteristic length of the <i>heterogeneity</i> . The wave equation then becomes parabolic in type. This approximation includes <i>diffraction</i> and <i>forward scattering</i> , but neglects <i>back-scattering</i> . See Sato et al. (2002).                                                                           |
| <b>parameter data</b>                          | A quantitative <i>attribute</i> of a seismic <i>arrival</i> , such as <i>onset</i> or <i>arrival time</i> , ( <i>back</i> ) <i>azimuth</i> , <i>slowness</i> , <i>polarisation</i> , <i>period</i> , and <i>amplitude</i> .                                                                                                                                                                                                                                                                                                             |
| <b>paraxial ray</b>                            | Originally defined for an optical system with a symmetrical axis of <i>rotation</i> . A light ray near this axis and which has only a small inclination to the axis is called a paraxial ray. In <i>seismic ray</i> theory it is defined as a ray generated by perturbing the source position or take-off direction of a central ray. The <i>dynamic ray equation</i> governs the properties of the paraxial ray. See Chapter 9 in Lee et al. (2002).                                                                                   |
| <b>partial autocorrelation function (PACF)</b> | Plays an important role in data analyses aimed at identifying the extent of the lag in an autoregressive model (see <i>autoregressive process</i> ). The use of this function was introduced as part of the Box-Jenkins approach (Box et al., 2008) to time-series modeling, where by plotting the partial autocorrelative functions one could determine the appropriate lags <b>p</b> in an autoregressive AR( <b>p</b> ) model or in an extended AutoRegressive Integrated Moving Average ARIMA( <b>p,d,q</b> ) model. See Wikipedia. |
| <b>participation factor</b>                    | A factor that apportions scaled versions of the input to a <i>multi-degree of freedom</i> system to the various modes of the system. The factor is not unique and depends on how the mode shapes are normalized. In one common normalization scheme the participation factors add to unity; in this case the participation factor for a mode is the fraction of excitation that mode receives. See Jennings (2003).                                                                                                                     |
| <b>pascal (Pa)</b>                             | The <i>SI</i> derived unit for <i>pressure</i> or <i>stress</i> , named after Blaise Pascal (1623-62) for his pioneering work on hydrostatics. It is defined as a force of one <i>newton</i> per square <i>meter</i> ( $\text{N m}^{-2}$ ), or ( $\text{m}^{-1} \text{ kg s}^{-2}$ ). In older seismological literature the unit for                                                                                                                                                                                                    |



|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|----------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                  | pressure and stress is usually expressed in the cgs unit bar, with 1 MPa (i.e., $10^6$ Pa) = 10 bar.                                                                                                                                                                                                                                                                                                                                                          |
| <b>passband</b>                  | The portion of the <i>frequency spectrum</i> that a <i>filter</i> passes with less than 3- <i>decibel attenuation</i> . See Ballou (1987).                                                                                                                                                                                                                                                                                                                    |
| <b>passive margin</b>            | A continental margin that is not a <i>plate boundary</i> , formed, for example, as <i>sea-floor spreading</i> carries the continental block away from the spreading center. If earthquakes occur on such a margin, they are <i>intraplate</i> events (q.v.). See also <i>active margin</i> .                                                                                                                                                                  |
| <b>passive-source seismology</b> | Seismic investigations that utilize data from naturally occurring <i>seismic events</i> . See Mooney et al. (2002).                                                                                                                                                                                                                                                                                                                                           |
| <b>path effect</b>               | The effect of the propagation path on seismic <i>ground motions</i> . It is implicitly assumed that the <i>source</i> , path and <i>site effects</i> on ground motions are separable.                                                                                                                                                                                                                                                                         |
| <b>Pb, P*</b>                    | Designates <i>P waves</i> critically refracted (see <i>refraction</i> ) through an intermediate layer in the Earth's <i>crust</i> with a <i>velocity</i> between ~6.5 and ~7.5 km/sec. The upper boundary of this layer has been called the <i>Conrad discontinuity</i> . An alternative, still accepted phase name is P*. For these and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). |
| <b>P coda</b>                    | The portion of <i>P waves</i> after the <i>arrival</i> of the primary waves. They may be due to P-to-S-wave <i>conversions</i> at <i>interfaces</i> or to multiple <i>reflections</i> in layers or to <i>scattering</i> by three-dimensional <i>inhomogeneities</i> .                                                                                                                                                                                         |
| <b>peak acceleration</b>         | See <i>peak ground acceleration</i> .                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>peak ground acceleration</b>  | The maximum <i>acceleration amplitude</i> measured (or expected) in a strong-motion <i>accelerogram</i> of an earthquake, abbreviated PGA. See Anderson (2003).                                                                                                                                                                                                                                                                                               |
| <b>pedogenic</b>                 | Pertaining to processes that add, transfer, transform, or remove <i>soil</i> .                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Peléan eruption</b>           | A type of eruption named after the 1902 eruption of Montagne Pelée, Martinique. It is associated with relatively viscous <i>magmas</i> (e.g., andesitic to rhyolitic compositions). The most violent and destructive activity generally occurs early in the eruption and commonly involves <i>nuées ardentes</i> , followed by emplacement of <i>lava domes</i> and thick <i>lava flows</i> .                                                                 |
| <b>pendulum</b>                  | In relation to the design of an <i>inertial sensor</i> , a pendulum is either (1) a mass moving strictly in a translational manner (Figure 1, bottom), as in most simple <i>geophones</i> ; or (2) a mass attached by a rigid arm to a pivot axis orthogonal to that arm                                                                                                                                                                                      |

and to the active axis of the sensor, as in many *accelerometers* and broadband velocity sensors (Figure 1, top). A mass moving on a simple cantilever, as in many *MEMS devices*, is equivalent to (2), whereas the restoration of classical pendulums derives solely from the gravitational force, with springs.



**Figure 1.** Schematic diagrams of a mass-on-rod pendulum (*top*) and a translational pendulum (*bottom*). The mass-on-rod pendulum moves in a circular arc about its pivot, here pointing into the page, and is parallel to the putative active axis only while at zero deflection. The translational pendulum is constrained by its suspension to move only in the active-axis direction. “Spring” is shorthand for all mechanical and force feedback restoring forces in combination.

They provide, either acting alone or in conjunction with *gravity*, the restoring force in common seismic pendulums. Nearly all pendulums are purposely damped by some means. The intended direction of (infinitesimal) mass motion relative to the instrument frame (thus the ground) is the *active axis* of the pendulum. Most seismic sensors are based on one or the other of these forms of constrained pendulums, although a few are based on magneto-hydrodynamic and other phenomena. In the simple or common pendulum, the mass is below the pivot, and constrained to move in a vertical plane; in the inverted pendulum, the mass is directly over the pivot; and in the horizontal pendulum, the mass is to one side of the pivot, and is constrained to move in a nearly horizontal circle. See Chapter 5 (pp. 15-18 and Fig. 5.4-5.7) in this Manual.

**pendulum (gravitational)**

A classic instrument first studied by Galileo and Newton, a structure without springs, in which the line between the *rotation* axis (in the original, a fixed point) and the center of mass

|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|---------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                 | oscillates with dampened simple harmonic motion (in the linear approximation) about the direction of the local gravitational field of the Earth. Its equilibrium orientation is the plumb line. See Fig. 5.5 in Chapter 5 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>pendulum (mass-on-rod)</b>   | A pendulum (Figure 1, top) that rotates about some axis with the mass on the other end of a rigid rod, or with the mass at the end of a stiff cantilever spring (the pendulum of many modern force-balance <i>accelerometers</i> and <i>broadband seismographs</i> ). Because the rod can also rotate away from the normal to the active axis, this design may be affected by translational accelerations along the axis of the rod while it is offset, although such offsets are minimized in force-feedback and stiff (e.g., <i>MEMS</i> ) designs. Mass-on-rod pendulums also are directly affected by <i>rotational accelerations</i> about the pivot axis. See also Figs. 5.4 to 5.7 of Chapter 5 in this Manual. |
| <b>pendulum, translational</b>  | A mass-on-spring pendulum (Figure 1, bottom) where the mass moves only along a linear translational path relative to the sensor's case (the mass and multiple constraining cantilever springs of a common geophone or a few <i>MEMS devices</i> in which multiple cantilevers and careful design obviate any effects of rotation). This design has minimum sensitivity to off-axis translational and <i>rotational accelerations</i> but is sensitive to <i>Coriolis forces</i> .                                                                                                                                                                                                                                      |
| <b>peridotite</b>               | Rock comprising the uppermost <i>mantle</i> , composed mainly of olivine [(Mg, Fe) $2\text{SiO}_4$ ] and pyroxene [(Mg, Fe, Ca) $\text{SiO}_3$ ]; other minerals, such as garnet, can also be present.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>period</b>                   | The length of time required to complete one cycle or a single oscillation of a periodic process. See <i>natural period</i> or <i>natural frequency</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>period (of a signal)</b>     | Most <i>seismic signals</i> are aperiodic but can be mathematically represented as a superposition of periodic sinusoidal signals. The “period of a seismic signal” usually means the period of the dominant sinusoidal signal.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>permeability (hydraulic)</b> | Rate of fluid flow through rock per unit <i>pressure gradient</i> . See Johnston and Linde (2002), and Johnston (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>permeability (magnetic)</b>  | A coefficient relating magnetic flux density to magnetic field intensity. See Johnston (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>permittivity</b>             | A constant relating the force between two charges separated by a distance. See Johnston (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>perovskite</b>               | A dense, high-pressure phase of pyroxene [(Mg, Fe, Ca) $\text{SiO}_3$ ]; it is only stable at high <i>pressures</i> (above ~20 GPa), but is                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |

thought to be the predominant mineral of the Earth's *mantle*. See Jeanloz (2002).

**Pg,  $\bar{P}$**

*Travel-time* curves at short distances (up to a few hundred km) for *seismic sources* in the Earth's *crust* usually consist of two intersecting straight lines; one with a *velocity* of about 6 km/sec at shorter distances and the other with a velocity of about 8 km/sec at greater distances. The former is attributed to direct *P* waves propagating through the crust and is designated as Pg, which stands for granitic layer. An alternative, older name has been  $\bar{P}$ , which is no longer standard. For these and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**PGA**

See *peak ground acceleration*.

**Phanerozoic**

An eon in *geologic time*, from 545 million years ago to the present. It is from Greek and means “apparent life”.

**phase**

(1) A stage in periodic motion, such as wave motion or the motion of an *oscillator*, measured with respect to a given initial point and expressed in angular measure. (2) A pulse, *wavelet* or group of seismic *body-* or *surface-wave* energy arriving at a definite time, which passed the Earth on a specific path (see also *phase nomenclature*). (3) Stages in the physical properties of rocks or minerals under differing conditions of *pressure*, temperature, and water content.

**phase names**

*Seismograms* consist of distinguished *onsets* of different *phases*. During the last 120 years, seismologists deciphered the principle character and the travel paths through the Earth of most of these onsets and developed a nomenclature to name them. The newest version of these naming rules was accepted by IASPEI in 2003 during its General Assembly in Sapporo. A detailed description of these rules, many examples of phase names and explaining plots to the different *seismic wave* paths can be found in IS 2.1 in this Manual or in Storchak et al. (2003 and 2011).

**phase response**

The phase of the complex *frequency response*, e.g., of a *seismograph*.

**phase velocity**

*Frequency-dependent velocity* of propagation of a specific *phase* position (angle) in a dispersive wave train, such as a *surface-wave* group. See also *group velocity*.

**phreatic**

Adjective of Greek origin (from phrear, “well”) relating to ground water. Commonly applied to explosive *eruptions* triggered by interaction of water with *magma* or solidified but still hot volcanic rock that eject steam and fragments of pre-

existing solid rock (lithics), but not magma. “Phreatic” is synonymous with “*steam-blast*”.

- phreatomagmatic eruption** Volcanic *eruption* that results partly from the interaction of external water and *magma*, resulting in explosive conversion of water to steam and disruption of the erupted *lava*.
- physical master model** A type of *master model of earthquakes* in which universally applicable laws of physics and chemistry are applied to geologic processes to integrate earth science information regarding earthquakes in a given region. Its construction has been an ultimate goal beyond that of the hazard master model at the Southern California Earthquake Center. See Aki (2002).
- piercing line** A term in *paleoseismology* indicating a feature that crosses a *fault*. The point where a piercing line intersects a fault is called “piercing point”. See Grant (2002).
- piezomagnetism** Changes in rock magnetism resulting from *stress* loading. See Johnston (2002).
- pillow lava** Type of submarine *lava flow* resulting from underwater volcanic *eruptions*, or from the entry of lava erupted on land into the ocean, with formation of bulbous and tubular masses with glassy surfaces upon chilling in water.
- pinned connection** A connection that can transmit shear but not *moment*, like a hinge. Historically, actual pins were used in such connections, particularly in bridges, but in modern construction most such “pinned joints” are welded or bolted joints not designed to transmit significant moments. See Dowrick (1977), and Salmon and Johnson (1996).
- PKIKP** Symbol for a *P wave* transmitted through the outer and *inner core* of the Earth; an alternative phase name for the PKPdf branch. See Chapter 2, Fig. 2.54, Chapter 11, Figs. 11.73 and 11.77, Figures 1, 2 and 5 of EX 11.3 in this Manual; also Kulhanek (2002, p. 342), and Cara (2002, p. 880-881). For this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).
- PKJKP** A *P wave* that travels as a converted *shear wave* through the solid *inner core* (known as a J wave). The PKJKP phase, though ardently sought, has yet to be reliably observed. See Song (2002). For this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).
- PKPpre** PKP phases become the first geometric *arrivals* beyond the *core shadow*. However, sometimes on short period *seismograms* at a distance range of about 128° to 145°, clearly

visible waves arrive a few to 10 seconds prior to the PKP phases. The addition pre stands for precursory. These precursors are explained by *scattering at heterogeneities* in the lowermost *mantle (D'')* or near or at the *core-mantle-boundary*. See Kulhanek (2002, p. 341-342), and Song (2002). For this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**PKP triplication**

This *triplication*, which has three distinct *branches* of a *seismic phase* at the same distance, is the result of wave propagation through a zone of steep *velocity* changes with depth within the Earth's *core*. At distances between  $\sim 145^\circ$  and  $155^\circ$  all three PKP phases PKPab, PKPbc, and PKPdb can be observed. (see Figs. 11.72, 11.73, 11.76 and 11.77 in Chapter 11 of this Manual). For these and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**planar frame**

An assemblage of vertical columns and horizontal beams, with or without inclined braces, where all members lie in a single plane. See Park and Paulay (1975).

**plane strain**

Also called in-plane *strain* in contrast to *anti-plane strain*; used to simplify mathematical analysis of elastic problem, in which *displacement* depends on only two spatial coordinates and its direction lies in the plane of those coordinates. *Dip-slip* displacement that does not vary along *strike* is an example of plane-strain *deformation*. Another example is a *Rayleigh wave* from a line source. See *anti-plane strain*; also Love (1944).

**plane wave**

A wave for which the constant-phase surfaces are planes perpendicular to the direction of propagation. See *spherical wave*, *inhomogenous plane waves*; also Chapter 9 of this Manual, Aki and Richards (2002 or 2009, p. 119-187), and section 3 of Ben-Zion (2003).

**plastic hinge**

A term given to a point on a structural element, typically a beam or column, where the element has yielded to the extent that increased curvature occurs without an increase in the resisting moment. See Bolt and Abrahamson (2003).

**plastic strain rate**

Irreversible *strain* rate expressed by a tensor. See Teisseyre and Majewski (2002).

**plate**

Here the same as *tectonic plate*. In *tectonics*, a plate is an internally rigid segment of the Earth's *lithosphere* which is in motion relative to other plates and to deeper mantle beneath the lithosphere. See *plate motion*, *plate tectonics*, and Uyeda (2002), and Stein and Klosko (2002).

|                              |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>plate boundary (zone)</b> | The zone of diffuse <i>deformation</i> , often marked by a distribution of <i>seismicity</i> , <i>active faulting</i> , and rough topography, within which relative <i>plate motion</i> is accommodated. Plate boundary zones are typically narrow under oceans and broad under continents. See Stein and Klosko (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>plate driving force</b>   | A forces that drives the motion of Earth's <i>lithospheric plates</i> . Terms such as “ <i>ridge push</i> ”, “ <i>slab pull</i> ” and “basal shear” have been used to describe them. See Zoback and Zoback (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>plate motion</b>          | The geometrical combination of three fundamental kinds of relative plate motions produces the observed <i>plate</i> patterns on Earth. Pairs of plates in contact at mutual boundaries either move apart (plate divergence), toward one another (plate convergence), or slip sideways past one another along structural ruptures ( <i>transform faults</i> ). The topology of <i>plate boundaries</i> on a sphere also dictates that three plates locally meet at common junctures ( <i>triple junctions</i> ), which may involve any of the three kinds of plate boundaries in any combination. The present Earth may be divided into six major and six smaller plates, with various additional “microplates” that jostle between the larger plates along some plate boundaries. See <i>Euler vector</i> . See also Uyeda (2002), and Stein and Klosko (2002), and Dickinson (2002). |
| <b>plate tectonics</b>       | The <i>lithosphere</i> of the Earth is broken into six major, and six smaller <i>plates</i> and various microplates that are in relative motion with respect to each other, over the underlying hotter and more fluid <i>asthenosphere</i> . Plate tectonics is the single coherent theory which reconciles <i>continental drift</i> , <i>sea-floor spreading</i> and deep focus earthquakes and explains (most) of the <i>tectonic</i> phenomena such as folding, faulting, rifting as well as related seismic and volcanic activity as the result of the interaction between or fragmentation (break-off) of lithosphere plates. Thus, plate tectonics provides a modern paradigm for the whole Earth science community. See Uyeda (2002), Stein and Klosko (2002), and Dickinson (2002). See also <i>plate motion</i> , <i>inter-plate</i> and <i>intra-plate</i> .                |
| <b>Pleistocene</b>           | An epoch in <i>geologic time</i> between about 10,000 and 1,6 million years before present. As a descriptive term applied to rocks or <i>faults</i> , it marks the period of rock formation or the time of most recent <i>fault slip</i> , respectively. Faults of Pleistocene age may be considered active though their activity rates are commonly lower than for younger faults.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>Plinian eruption</b>      | Explosive <i>eruption</i> that generates a high and sustained <i>eruption column</i> , producing <i>fallout</i> of <i>pumice</i> and <i>ash</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>plume</b>                 | See <i>mantle plume</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |

|                       |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
|-----------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>PL waves</b>       | A train of long- <i>period</i> (30 to 50 seconds) waves observed in the interval between <i>P</i> waves and <i>S</i> waves for distances less than about 30°. They show normal <i>dispersion</i> (longer periods arriving earlier). They are explained as a <i>leaking mode</i> of the <i>crust-mantle</i> waveguide with <i>P</i> to <i>Rayleigh</i> wave conversion. See Aki and Richards (2002, p. 323). For these and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>Pn</b>             | According to the <i>IASPEI</i> standard <i>phase name</i> list Pn is a <i>P</i> wave bottoming in the <i>upper mantle</i> or an upgoing <i>P</i> wave from a source in the uppermost mantle (see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). This differs to some extent from earlier related definitions and considerations. Beyond a certain critical distance, usually beyond 100 to 200 km depending on crustal thickness and source depth, the first <i>arrival</i> from <i>seismic sources</i> in the <i>crust</i> corresponds to waves refracted under a (near) critical angle from the top of the mantle. The <i>amplitudes</i> of Pn waves are relatively small, with long-period motion followed by larger and sharper waves of shorter period called <i>Pg</i> , which propagated directly through the crust. Yet, at <i>epicentral distances</i> between some 300 to 600 km the <i>Moho</i> reflected PnPn may be the first strong arrival after Pn with sometimes rather large amplitudes. The Pn wave has long been interpreted as a <i>head (conical) wave</i> along the <i>interface</i> between two <i>homogeneous</i> media - namely, crust and mantle. The observed amplitude, however, is usually greater than that predicted for head waves, implying that the <i>velocity</i> change at the crustal base ( <i>Mohorovičić discontinuity</i> ; <i>Moho</i> ) is not exactly step-like but has a finite <i>gradient</i> at or below the transition zone. The designation Pn is commonly applied to short-period <i>P</i> waves that propagate over considerable distances (even up to 20° in continental platform regions) with horizontal <i>phase velocities</i> in the range of ~7.5-8.5 km/s. This can also not been explained in terms of head wave propagation at the <i>Moho</i> (although the horizontal velocity and <i>travel times</i> would agree with it), because head waves must decay rapidly with distance. More likely is an explanation in terms of <i>guided waves</i> , within a high-Q layer several tens of kilometers in thickness at the top of the mantle. See <i>Sn</i> ; also record examples in Chapter 2 (Figs. 2.17, 2.46 and 2.47), Chapter 11 (Figs. 11.58-11.60) and DS 11.1 of this Manual, Kulhanek (2002, p. 335-338), and Mooney et al. (2002). For these and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). |
| <b>point process</b>  | A series of events distributed according to a probabilistic rule. See Utsu (2002b).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>point rotation</b> | Rotational motions nominally of an infinitesimal point, generally measured by a dedicated <i>rotational sensor</i> either                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |



|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|---------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                 | deployed on a small rigid pad or attached to some point in a structure.                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>point seismic array</b>      | Co-located, inertial three-component <i>seismograph</i> , <i>rotational sensor</i> , and/or <i>strain meter</i> that can be used to perform seismic wave <i>gradiometry</i> . Although wave measurements are taken at a single observation point on the Earth, wave gradiometry produces estimates of wave parameters, such as horizontal <i>slowness</i> and <i>azimuth</i> , similar to large-aperture <i>arrays</i> of seismographs constructed for <i>beamforming</i> or <i>frequency-wavenumber analysis</i> . See <i>seismic array</i> . |
| <b>point source</b>             | Idealized, mathematical model of a <i>seismic source</i> , considered as the assumed startpoint (in time and space) of a rupture process.                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>point string deformation</b> | The point <i>deformation</i> with the simultaneous extension and contraction perpendicular to each other and of the equal values.                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>Poisson distribution</b>     | A probability distribution that characterizes discrete events occurring independently of one another in time. See Vere-Jones and Ogata (2003).                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>Poisson process</b>          | A point process in which events are statistical independently and uniformly distributed with respect to the independent variable, e.g., time. See Utsu (2002b).                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Poisson's ratio</b>          | An elastic body like a metal bar under uniaxial tension shows elongation along the tension axis and shrinkage in the transverse direction. The absolute ratio of the transverse <i>strain</i> to the longitudinal strain is called Poisson's ratio. Poisson's ratios for rocks are typically around 0.25. See Fig. 2.2 in Chapter 2 of this Manual as well as Christensen and Stanley (2003), Eq. (1.6) in Ben-Zion (2003), and the biography of Simeon Denis Poisson in Howell (2003).                                                        |
| <b>polarity</b>                 | In <i>seismology</i> , the direction of <i>first motion</i> on a vertical component <i>seismogram</i> ; either up (compression) or down (dilatation or relaxation).                                                                                                                                                                                                                                                                                                                                                                            |
| <b>polarity reversal</b>        | In seismology, the occurrence of <i>waveforms</i> that are mirror images of their own, e.g., the waveforms of <i>depth phases</i> with respect to their primary <i>P</i> or <i>S</i> waves.                                                                                                                                                                                                                                                                                                                                                    |
| <b>polarization</b>             | Direction of particle motion relative to the direction of wave propagation. It differs for different types of <i>seismic waves</i> such as <i>P</i> , <i>S</i> and <i>surface waves</i> and may be $\pm$ linear or elliptical, prograde or retrograde. It is also influenced by <i>heterogeneities</i> and <i>anisotropy</i> of the medium in which the seismic waves propagate and depends on their <i>frequencies</i> or <i>wavelengths</i> , respectively. The polarization of <i>ground motion</i> may be                                  |

reconstructed by analyzing three-component seismic recordings. See *Christoffel matrix*, *shear wave splitting*, *seismic anisotropy* and *earthquake focal mechanism*.

|                                        |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|----------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>polarization anisotropy</b>         | A term used (in contrast to <i>azimuthal anisotropy</i> ) to refer to an <i>anisotropic medium</i> in which the <i>seismic-wave</i> properties depend apparently on the <i>polarization</i> of the waves, and not on the <i>azimuth</i> of their propagation direction. <i>SH/SV velocity</i> differences and <i>Love-Rayleigh wave</i> discrepancies are often referred to as “polarization anisotropy” while azimuthal velocity variations would be due to “azimuthal anisotropy”. See Cara (2002).                                    |
| <b>polar tensor (vector or scalar)</b> | An object which does not depend on the orientation of a co-ordinate system. Examples of polar objects: <i>stress tensor</i> , force, translational velocity, and mass. Polar objects cannot be equal to axial objects and are often related to the translational motions.                                                                                                                                                                                                                                                                |
| <b>polar wandering</b>                 | The movement of geographic or <i>paleomagnetic poles</i> over <i>geologic time</i> . See Uyeda (2002), Uyeda (1978), and Duff (1993).                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>poloidal</b>                        | See <i>spheroidal</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>polygenetic volcano</b>             | A long-lived volcano that has erupted repeatedly throughout its lifetime, often in an episodic manner.                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>pore-fluid factor</b>               | The ratio of fluid pressure to overburden <i>pressure</i> ( $\lambda_v = P_f / \sigma_v$ ) employed as a measure of the degree of overpressuring. For typical rock <i>densities</i> , $\lambda_v \sim 0.4$ and $\lambda_v = 1.0$ denote hydrostatic and lithostatic fluid-pressure conditions, respectively. See Sibson (2002).                                                                                                                                                                                                          |
| <b>pore pressure</b>                   | In soil mechanics, the <i>pressure</i> within the fluid that fills voids between soil particles. In rock mechanics, the pressure due to liquids in formations that offsets the strengthening effect of <i>normal stresses</i> acting on a <i>fault</i> . At a given depth, the ambient pore pressure is often observed to be nearly equivalent to that of a column of water extending upward to the ground surface, a pore pressure regime termed the hydrostatic. See Johnston and Linde (2002), McGarr et al. (2002), and Youd (2003). |
| <b>porosity</b>                        | The volume of voids per unit volume of a porous material usually expressed as a percentage.                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>poststack seismic data</b>          | A term in <i>reflection seismology</i> ; seismic data set reduced by the <i>stacking process</i> such that each surface location is represented by a single stacked seismic trace. See <i>prestack seismic data</i> .                                                                                                                                                                                                                                                                                                                    |
| <b>potency</b>                         | See <i>seismic potency</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |

|                               |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|-------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>power</b>                  | In general terms, the source of physical or mechanical force or energy; in mathematics, the product of the multiplication of a quantity by itself (e.g., 4 is the second power of 2, i.e., $2^2$ ); in a more general sense, the exponent to the base, which may also be a non-integer, and negative number.                                                                                                                                                                                                                              |
| <b>power law</b>              | The mathematical relationship between two quantities X and Y, if $Y = X^n$ with a power (exponent) $n \neq 1$ . E.g., when the frequency of occurrence of earthquakes varies as a power of earthquake size, then the <i>frequency</i> is said to follow a power law. Many physical, biological, geological, man-made and other phenomena follow a power law.                                                                                                                                                                              |
| <b>power spectral density</b> | The power spectral density of a stationary time series is defined as the <i>Fourier transform</i> of its <i>auto-correlation function</i> . The power spectral density integrated over the whole <i>frequency</i> range is equal to the mean square of the time series, hence the name “spectral density”. There is a factor 2 difference depending on whether the frequency range is taken from 0 to $\infty$ or $-\infty$ to $+\infty$ . See Chapter 4 in this Manual and Aki and Richards (2002 or 2009, p. 610-611).                  |
| <b>PP, PPP, PS, PPS etc.</b>  | Single, respectively double <i>reflection</i> (or <i>conversion to S</i> ) at the free surface of a <i>P</i> wave leaving a source downward. See also <i>PzP</i> . For these and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).                                                                                                                                                                                                                                     |
| <b>precision</b>              | The closeness with which separate observations (or parameter values calculated from observed data on the basis of a model) agree with each other. According to Bevington and Robinson (2003) a measure of how carefully a result is determined without reference to any “true” value (in contrast to <i>accuracy</i> ).                                                                                                                                                                                                                   |
| <b>precursor time</b>         | The time span between the onset of an anomalous phenomenon and the occurrence of the <i>mainshock</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>pre-event memory</b>       | In a triggered recording system of a digital <i>seismograph</i> , it refers to a signal buffering memory in the <i>data logger</i> from which the <i>ground motion</i> can be retrieved for a time segment prior to the time the instrument was triggered.                                                                                                                                                                                                                                                                                |
| <b>preliminary tremors</b>    | In early seismological studies, the term was used for the small motions at the start of an earthquake record. For <i>local earthquakes</i> , this corresponds to the waves arriving between the (in modern terms) first <i>P</i> and main <i>S</i> arrival. Omori found that its duration (or time difference between <i>S</i> and <i>P</i> onset) is proportional to the <i>epicentral distance</i> by a factor of about 8 km/sec; distance = $8 \times (S - P \text{ onset time})$ . See biography of Fusakichi Omori in Howell (2003). |

|                                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>pressure</b>                          | If a material element has zero “shear stress”, then the <i>principal stresses</i> all share the same value; this value is the negative pressure. Any rotation of the coordinate system produces the same value for the <i>normal stresses</i> , hence a stress state with zero shear stress is <i>isotropic</i> and the <i>stress tensor</i> can be written as the product of the scalar quantity “pressure” and the identity matrix. Even if shear stresses are present, the isotropic component of the stress state can be extracted from the general stress tensor. The pressure is simply the average value of the three normal stresses regardless of the coordinate system orientation. See Ruff (2002).                                                                                                                                                                                                                                                                                                                        |
| <b>pressure solution</b>                 | Water-assisted <i>ductile deformation</i> mechanism whereby minerals are sequentially dissolved from highly stressed regions, transported through fluid by diffusion or advection, and precipitated in low <i>stress</i> regions. See Lockner and Beeler (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>prestack seismic data</b>             | A term in <i>reflection seismology</i> referring to the entire seismic data set before reduction by the <i>stacking process</i> . See <i>poststack seismic data</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>primary seismic station</b>           | <i>Seismic station</i> or <i>array</i> of the International Monitoring System (IMS) that is part of the detection network to monitor compliance with the Comprehensive Test-Ban Treaty (CTBT). See Chapter 17 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>principal axes</b>                    | In general, the principal axes are the eigenvectors of a symmetric tensor, i.e., the three orthogonal axes at which the rotated tensor has only diagonal elements. They can, for example, characterize the orientation of the <i>stress</i> and the <i>strain</i> at a point, and need not be tied to the geographic coordinate system. For example, in the case of the <i>seismic moment</i> tensor corresponding to a <i>double couple</i> source, the principal axes are called the compression axis, tension axis and null axis. With respect to seismic signals, as proposed by Penzien and Watabe (1975), the principal axes are a system of Cartesian coordinates on the ground surface for which three seismic signals, viewed as stochastic processes along respective axes $u_x(t)$ , $u_y(t)$ , and $u_z(t)$ , are least correlated. The system of principal axes is expected to be situated in such a way that one of the three axes is perpendicular to ground surface and one is directed toward the <i>epicenter</i> . |
| <b>principal displacement zone (PDZ)</b> | The zone of concentrated <i>slip</i> (sometimes referred to as the <i>fault core</i> ) accommodating most of the <i>displacement</i> within the more distributed <i>deformation</i> (damage zone) defining a major <i>fault zone</i> . See Sibson (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |

- principal part (of seismogram)** In early seismological studies, the term used for the largest motions seen on the *seismogram* from an earthquake. See Agnew (2002).
- principal plane (of shear fracture)** The most favorably oriented plane for shear fracture. Given planes of equal shear strength and of all possible orientations, the principal plane is that which minimizes the work done in shear. The orientation is given by  $\tan 2\beta = 1/\mu_*$ , where  $\beta$  is the angle between the greatest *principal stress* and the normal to the plane of interest, and  $\mu_*$  is the coefficient of *friction* for a *fault* or the coefficient of internal friction for an intact rock. See Lockner and Beeler (2002, p. 508).
- principal stresses** The three *normal stresses* in the particular coordinate frame (principal axes) oriented so that the *stress tensor* reduces to diagonal form (i.e., no shear stresses). See Ruff (2002), Harris (2003), and Figure 1 in Ben-Zion (2003). See also *normal stress*, *shear stress*, and *principal axes*.
- principle of material objectivity** Also termed material frame indifference, because the constitutive laws governing the internal interactions between the parts of a physical system should not depend upon the external frame of reference used to describe them. For continuous media, this implies that *stress* and *couple-stress tensors* have to rotate together with the material when subjected to a rigid motion and not depend upon its translational part.
- prior information** The information one may have on the parameters characterizing a system, prior to any actual measurement. See Mosegaard and Tarantola (2002).
- probabilistic earthquake scenario** A representation, in terms of useful descriptive parameters, of earthquake effects with a specified probability of exceedance during a prescribed period in an area. See Faccioli and Pessina (2003).
- probabilistic seismic hazard analysis (PSHA)** Available information on earthquake sources in a given region is combined with theoretical and empirical relations among earthquake *magnitude*, distance from the source and local site conditions to evaluate the exceedance probability of a certain *ground motion* parameter, such as the *peak acceleration*, at a given site during a prescribed period. See Aki (2002), and Somerville and Moriawaki (2003).
- probability density** A function of a continuous statistical variable whose integral over a given interval gives the probability that the variable will fall within that interval. See Vere-Jones and Ogata (2003).

|                                    |                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
|------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>probability of exceedance</b>   | The probability that, in a given area or site, an earthquake <i>ground motion</i> will be greater than a given value during some time period. See Giardini et al. (2003).                                                                                                                                                                                                                                                                                           |
| <b>probable maximum loss (PML)</b> | A probable upper limit of the <i>losses</i> that are expected to occur as a result of a damaging earthquake, normally defined as the largest monetary loss associated with one or more earthquakes proposed to occur on specific <i>faults</i> or within specific source zones.                                                                                                                                                                                     |
| <b>prograde branch</b>             | See <i>branch (of travel-time curve)</i> .                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>Proterozoic</b>                 | An eon in geologic time, from 2.5 billion to 545 million years ago. It is from Greek and means “first life”. See <i>geologic time</i> .                                                                                                                                                                                                                                                                                                                             |
| <b>protolith</b>                   | The original rock from which a particular <i>metamorphic rock</i> is eventually formed.                                                                                                                                                                                                                                                                                                                                                                             |
| <b>PSD</b>                         | See <i>power spectral density</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>pseudotachylyte</b>             | A black aphanitic <i>fault-rock</i> retaining evidence of a melt origin (relic glass, devitrification or quench textures) attributed to heat generated by localized <i>seismic slip</i> . Also developed in association with impact structures or at the base of some <i>landslides</i> . See Sibson (2002).                                                                                                                                                        |
| <b>pseudo vector</b>               | A quantity that behaves as a vector as long as the handedness of the coordinate system (i.e., it is right- or left-handed) remains unchanged. Changing the handedness changes the direction of the quantity. For this reason, vectors cannot be equated to pseudo vectors (see, e.g., Pujol , 2009) An example of a pseudo-vector is the vector product of two vectors. Pseudo vectors are also known as axial vectors. See <i>axial tensor, vector or scalar</i> . |
| <b>PSHA</b>                        | See <i>probabilistic seismic hazard analysis</i> .                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>pull-apart basin</b>            | A topographic depression produced by extensional bends or stepovers along a <i>strike-slip fault</i> .                                                                                                                                                                                                                                                                                                                                                              |
| <b>pumice</b>                      | Low- <i>density</i> fragments of silicate glass (generally of dacitic to rhyolitic composition) containing few crystals but a large fraction of gas bubbles (voids) due to exsolution of gas from <i>magma</i> during eruption. See <i>scoria</i> .                                                                                                                                                                                                                 |
| <b>push over analysis</b>          | An analysis used to estimate the capacity of a structure to resist collapse from strong <i>ground motion</i> . Typically used in earthquake resistant design studies, the analysis employs a selected force profile that is increased in intensity until the analysis indicates a structural element will fail or yield. That member is replaced by its yielding resistance, or removed if failed, and the force profile increased until another member             |

yields or fails. The process is repeated until the overall structure has a constant or declining resistance under increased load. The deflection profile at this final stage approximates the structure's ultimate, non-linear, earthquake resistance. See Bolt and Abrahamson (2003).

**p value**

The exponent of the time factor in the modified *Omori relation* showing the rate of decay of *aftershock* occurrence. A higher p value indicates more rapid decay. Estimated p values for most aftershock sequences fall between 0.9 and 1.5. In the original Omori relation, the p value is fixed as 1. See Utsu (2002b).

**P wave**

A symbol for a *seismic body wave* that involves particle motion (alternating compression and dilatation) in the direction of propagation. P waves travel faster than S waves and, therefore, arrive earlier in the record of a *seismic event* (P stands for “*unda prima*” = primary wave). The particle *displacement* associated with P waves (for a classical linear elastic continuum in an *isotropic far field*) is parallel to the direction of wave propagation. For this reason, P waves are sometimes called longitudinal waves. See Chapter 2 in this Manual and Aki and Richards (2002, p. 122). In isotropic media, the *curl* of the P-wave field (i.e., rotational motion) is zero. However, this is not so in general, for instance, for anisotropic or *Cosserat continua*. For this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**pyroclast**

A general term of any fragment of *lava* ejected during an explosive volcanic *eruption*; by definition, pyroclastic deposits are composed of pyroclasts.

**pyroclastic**

Adjective of Greek derivation (from pyr, “fire”; klastos, “broken”) relating to fragmental materials formed by the shredding of molten or semi-molten *lava* and (or) the shattering of pre-existing solid rock (volcanic and other) during explosive *eruptions*, and to the volcanic deposits they form.

**pyroclastic flows**

Ground-hugging mixtures of hot, generally dry, pyroclastic debris and volcanic gases that sweep along the ground surface at extremely high velocities (< 250 km/h). A continuum exists between pyroclastic flow and *pyroclastic surge*, depending on the ratio of solid materials to gases. Because of its relatively high aggregate density, the movement of pyroclastic flows tends to be more controlled by topography, mostly restricted to valley floors.

**pyroclastic surges**

Though part of a continuum with *pyroclastic flows*, pyroclastic surges have relatively lower ratios of solid materials to gases (lower aggregate *density*). In contrast to pyroclastic flows, the less dense, more mobile surges are less controlled by

topography and can affect areas high on valley walls and even overtop ridges to enter adjacent valleys.

**PzP**

This wave is like the surface *reflection* *PP*, except that the reflection occurs at an *interface* at depth *z* (expressed in kilometers, e.g., P600P) instead of at the surface. Similarly, for *conversions from P to S* one writes PzS (or vice versa). The new *IASPEI* phase name standards additionally differentiate with a + or – sign, whether this reflection/conversion occurred on the outer (upper) side (+) or on the inner (lower) side (-) of a *discontinuity*; e.g., P660-P is a P reflection from below the 600 km discontinuity, which means this phase is a precursor to PP. For examples see Figs. 2.68-2.70 in Chapter 2 and for other such symbols used in *seismic phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).

**Q**

**Q factor**

So-called quality factor, which is a measure of the dissipative characteristics of a system or a material, proportional to the inverse of the fractional loss of energy per cycle of oscillation. See also *Q inverse* and Eq. (3.9) in Ben-Zion (2003).

**Q inverse ( $Q^{-1}$ )**

The reciprocal of the *Q* value is the fractional loss (*attenuation*) of wave energy per cycle divided by  $2\pi$ . It is called spatial  $Q^{-1}$  if measured from the *amplitude* attenuation with distance and temporal  $Q^{-1}$  *attenuation* if measured from that with time. They can be different for dispersive waves. See Aki and Richards (2002, Box 5.7, p. 162-163).

**quasi-isotropic ray theory**

Quasi-isotropic ray theory describes *rays* in weakly *anisotropic* media and includes the coupling between quasi-shear rays. The quasi-isotropic ray *ansatz* depends on the *signal period* and the anisotropic part of the model parameters. See Chapman (2002).

**quasi-plastic (QP) regime**

That portion of the *crust* or *lithosphere* where *fault* motion is governed by localized *ductile* shearing in which temperature-sensitive crystal plastic and/or diffusional flow mechanisms dominate. See Sibson (2002).

**Quaternary**

A period in *geologic time*, from 1.6 million years ago to the present.

**quiescence**

A significant decrease in the rate of background *seismicity* from the long-time average rate.



## R

|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
|----------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>radar</b>                     | An acronym for RAdio Detection And Ranging. Synthetic aperture radar (SAR) <i>interferometry</i> is increasingly used by seismologists for analyzing the real surface <i>deformation</i> due to strong <i>earthquakes</i> and for deriving therefrom information about the rupture mechanism. See <i>satellite interferometry</i> and Feigl (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>radial anisotropy</b>         | A term sometimes used to denote hexagonal symmetry with a vertical symmetry axis. Traditionally called <i>transverse isotropy</i> . See Cara (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>radiation damping</b>         | The dissipation of energy through the transmission of vibratory energy that is not reflected or confined. Radiation <i>damping</i> occurs, for example, in the idealized problem of a vibrating structure imbedded in a linear, elastic <i>halfspace</i> , even when the material of the halfspace is not dissipative. See Jennings (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>radiation efficiency</b>      | A term introduced by Husseini (1977), which corresponds to the ratio between radiated <i>seismic energy</i> and the potential energy available for <i>strain</i> release. In contrast to <i>seismic efficiency</i> , the radiation efficiency accounts only for strain energy loss due to fracturing but not due to the additional conversion into frictional heat and sound energy. Therefore, values of radiation efficiency are significantly larger than those of <i>seismic efficiency</i> . Whereas the former ranges for most earthquakes between 0.25 and 1 (although they may drop even below 0.1 for very slow rupturing <i>tsunami earthquakes</i> ) the seismic efficiency is estimated to be only a few percent. See Kanamori and Brodsky (2004), Bormann and Di Giacomo (2011) and <i>seismic efficiency</i> . |
| <b>radiation pattern</b>         | Dependence of the <i>amplitudes</i> of seismic <i>P</i> and <i>S</i> waves on the direction and <i>take-off angle</i> under which their <i>seismic rays</i> have left the <i>seismic source</i> . It is controlled by the type of <i>source mechanism</i> , e.g., the orientation of the <i>earthquake fault plane</i> and <i>slip direction</i> in space. See <i>earthquake focal mechanism</i> , Chapter 3 and Ex 3.2 in this Manual, and Eq. (3.7b) in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                   |
| <b>radiative transfer theory</b> | A phenomenological theory that describes a multiple <i>scattering</i> process based on causality, <i>geometrical spreading</i> , and the energy conservation law, which neglects the interference between wave packets. It is often applied to model the propagation of high-frequency <i>seismic-wave</i> energy in <i>heterogeneous</i> Earth media. See Sato et al. (2002), and Sato and Fehler (1998, p. 173).                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>radiocarbon dating</b>        | <i>Geochronology</i> using radioactive $^{14}\text{C}$ <i>age dating</i> . See Grant (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |

- radiogenic heat production** The rate of heat generated in a unit volume of rock by spontaneous radioactive decay of natural unstable isotopes in the rock. The primary heat producing isotopes in the Earth are  $^{232}\text{Th}$ ,  $^{238}\text{U}$ ,  $^{40}\text{K}$ , and  $^{235}\text{U}$ . See Morgan (2002).
- radiometric** Pertaining to the measurement of *geologic time* by the analysis of certain radioisotopes in rocks and their known rates of decay (see  *$^{14}\text{C}$  age date*).
- random medium** A mathematical model for a medium whose spatial variation in parameters is described by random functions. Their *autocorrelation* functions or *power spectral density* functions give their statistical properties. See Sato et al. (2002).
- random vibration theory (RVT)** A theory of the response of systems to probabilistic defined inputs. In *seismology*, a convenient way of estimating *time domain* properties, such as peak *amplitudes* and *frequency*, from the *Fourier spectrum* of *ground motion* and an estimate of the duration of the motion. For applications, see the review by Boore (2003).
- random walk** The typical path followed by a molecule in *Brownian motion*. By extension, the path followed in a parameter space when using a *Monte Carlo method*. See the biography of Robert Brown in Howell (2003), Mosegaard and Tarantola (2002), and Tarantola (2005).
- range** In satellite geodesy; distance along the line of sight between the satellite and the ground. See Feigl (2002).
- rate and state dependent friction** A class of constitutive relationships that describe sliding *friction* over a wide range of conditions and time scales, developed by Ruina (1983) and Rice and Ruina (1983) as a generalization of equations of Dieterich (1978, 1979). These relations consist of a base frictional resistance and depend on the sliding speed (rate) and contacts (state) on the *fault plane*. See Lockner and Beeler (2002, p. 521), Harris (2003), and Eqs. (5.4a)-(5.4f) and Figure 12 in Ben-Zion (2003).
- rate strengthening** A term used to describe a positive dependence of *fault* shear strength on sliding velocity; always produces *aseismic* fault *slip*. See Lockner and Beeler (2002, p. 521).
- rate weakening** A term used to describe a negative dependence of *shear* strength on sliding velocity; can lead to frictional instability (*seismic slip*) under some circumstances. See Lockner and Beeler (2002, p. 521), and material below Eq. (5.4f) in Ben-Zion (2003).

|                             |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
|-----------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>ray</b>                  | A trajectory along which the constituent components of a wave, e.g., <i>phase</i> and energy, propagates. See Chapman (2002), and Bleistein (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>ray expansion</b>        | The complete seismic response can be expanded into a summation of terms associated with <i>rays</i> . Normally the response can be approximated by a limited number of rays in the time window and <i>amplitude</i> range of interest. See Chapman (2002).                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>ray Green's function</b> | The approximate <i>Green's function</i> obtained using asymptotic <i>ray theory</i> . See Chapman (2002), and the biography of George Green in Howell (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>Rayleigh damping</b>     | A form of <i>damping</i> used in the matrix analysis of structures in which the damping matrix is a constant times the mass matrix plus another constant times the stiffness matrix. Rayleigh damping produces viscous damping factors in the modes that are dependent on the <i>natural frequencies</i> of the modes. The constants are chosen to produce desired damping factors over the important modes of response. See Jennings (2003), and the biography of Lord Rayleigh in Howell (2003).                                                                                                                                                                      |
| <b>Rayleigh wave</b>        | Coupled <i>P</i> and <i>SV</i> waves trapped near the surface of the Earth and propagating along it. Their existence was first predicted by Rayleigh (1887) for a <i>homogeneous halfspace</i> , for which the <i>velocity</i> of propagation is 0.88 to 0.95 times the shear velocity, depending on the <i>Poisson's ratio</i> , and their particle motion is retrograde elliptic near the surface. They can exist, in general, in a vertically heterogeneous medium bounded by a free surface. See Udias (2002, p. 87-88), Kulhanek (2002, p. 334-335), Aki and Richards (2002 or 2009, p. 155-157), and the biography of Lord Rayleigh (1842-1919) in Howell (2003). |
| <b>ray parameter p</b>      | See <i>wavenumber vector k</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>ray theory</b>           | Theoretical approach, which treats wave propagation like particles moving along <i>seismic rays</i> . It is a valid approximation when the medium properties such as wave <i>velocity</i> , <i>density</i> and <i>attenuation</i> vary smoothly within a <i>wavelength</i> (high-frequency approximation). See Chapman (2002) and Červený (2001).                                                                                                                                                                                                                                                                                                                       |
| <b>ray-tracing method</b>   | Computational method of calculating ground-shaking estimates that assumes that the <i>ground motion</i> is composed of multiple arrivals of <i>seismic rays</i> and related energy bundles (Gauss beams) that leave the source and are reflected or refracted at <i>velocity</i> boundaries according to <i>Snell's Law</i> . The <i>amplitudes</i> of reflected and refracted waves at each boundary are recalculated according to the energy conservation law.                                                                                                                                                                                                        |
| <b>receding branch</b>      | See <i>branch (of travel-time curve)</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |

|                                               |                                                                                                                                                                                                                                                                                                                                                                                                                    |
|-----------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>receiver function method</b>               | See <i>H/V spectral ratio</i> and <i>receiver functions</i> .                                                                                                                                                                                                                                                                                                                                                      |
| <b>receiver functions</b>                     | The spectral ratio of the horizontal component of <i>S waves</i> to the vertical component of <i>P waves</i> recorded at a single station from a <i>teleseismic</i> event. Assuming a horizontally layered structure beneath the station, it gives an estimate of the seismic <i>velocity</i> structure, particularly the nature of <i>discontinuities</i> , of the <i>crust</i> and the uppermost <i>mantle</i> . |
| <b>record section</b>                         | A display of <i>seismograms</i> recorded from a source (e.g., an earthquake or an explosion) according to their distance from that source in a time-distance graph.                                                                                                                                                                                                                                                |
| <b>rectangular fault</b>                      | <i>Fault plane</i> geometry with a rectangular shape. The long and short side is, respectively, called <i>fault length</i> and <i>fault width</i> . The <i>rupture front</i> is a line segment parallel to the width. It starts at a point on the fault and propagate along the fault length. See IS 3.1 in this Manual and <i>Haskell model</i> .                                                                 |
| <b>recurrence interval</b>                    | The average time interval between consecutive events which occur repeatedly; also termed average return period. See Grant (2002).                                                                                                                                                                                                                                                                                  |
| <b>recursive technique</b>                    | A computational technique based on a formula which relates certain coefficients, say for a <i>filter</i> , at the <i>n</i> th step with those at the ( <i>n</i> -1)th step. Desired coefficients are obtained at the <i>n</i> th step by an iterative use of the formula starting with the initial values at the 0th step. See Sipkin (2002).                                                                      |
| <b>reduced travel time (<math>T_r</math>)</b> | The observed <i>travel-time</i> <i>T</i> minus the distance from a source <i>X</i> divided by a suitable <i>reduction velocity</i> ( $V_r$ ).                                                                                                                                                                                                                                                                      |
| <b>reduction velocity (<math>V_r</math>)</b>  | A suitable reduction velocity mostly used for controlled-source <i>P wave</i> crustal studies is 6 km/sec, for upper- <i>mantle</i> studies 8 km/sec, for <i>S wave</i> studies the reduction velocity is usually that for <i>P wave</i> studies divided by $\sqrt{3}$ .                                                                                                                                           |
| <b>reflection</b>                             | The energy or wave from a <i>seismic source</i> that has been returned (reflected) from an <i>interface</i> between materials of different elastic properties within the Earth.                                                                                                                                                                                                                                    |
| <b>reflection angle</b>                       | The angle between a <i>seismic ray</i> , leaving an <i>interface</i> on the same side as reaching it, and the vertical (normal) to this <i>discontinuity</i> . See also <i>incidence angle</i> and <i>refraction angle</i> .                                                                                                                                                                                       |
| <b>reflection coefficient</b>                 | The <i>amplitude</i> ratio of the reflected wave relative to the incident wave at an <i>interface</i> . In seismology, the coefficients are found by imposing boundary conditions that the seismic <i>displacement</i> and the traction acting on the <i>discontinuity</i> surface are continuous across the surface. These conditions physically                                                                  |

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|-------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                     | mean a welded contact and absence of extra-seismic sources at the discontinuity. Coefficients have been defined for different normalizations and signs of the component waves, e.g., energy or displacement (Knott 1899; Zoeppritz, 1919). See <i>transmission coefficient</i> , Chapman (2002, p.110-112), and Aki and Richards (2002 or 2009, p. 128-149).                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>reflection seismology</b>        | The study of the Earth's structure by the use of <i>seismic waves</i> originating from a near-surface source and reflected back to the surface from sub-surface <i>discontinuities</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>reflector</b>                    | An <i>interface</i> between materials of different elastic properties that reflects <i>seismic waves</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>refraction</b>                   | (1) The deflection, or bending, of the <i>ray</i> path of a <i>seismic wave</i> caused by its passage from one material to another having different elastic properties. (2) Bending of a <i>tsunami</i> wave front owing to variations in the water depth along a coastline.                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>refraction angle</b>             | The angle between a <i>seismic ray</i> , leaving an <i>interface</i> after passing it, and the vertical (normal) to this <i>discontinuity</i> . See also <i>incidence angle</i> and <i>reflection angle</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>refraction diagram (tsunami)</b> | A chart indicating <i>wavefronts</i> of a <i>tsunami</i> from the source. It is used to predict tsunami <i>travel times</i> from a given source area. See Satake (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>refraction seismology</b>        | The study of the Earth's structure by the use of <i>seismic waves</i> from a near-surface source refracted back to the surface after propagation along deep <i>interfaces</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>regional strain</b>              | A description of the overall <i>deformation</i> of a region. See Jackson (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>regional waves</b>               | <i>Seismic waves</i> which propagate laterally in the <i>crust</i> and uppermost <i>mantle</i> . Examples are <i>Pg</i> , <i>Pn</i> , <i>Sn</i> , <i>Lg</i> and <i>Rg</i> . These waves are distinguished from <i>teleseismic</i> waves which propagate in the deeper interior of the Earth. Regional waves typically propagate over <i>epicentral distances</i> up to 1500 km (somewhat further for <i>Lg</i> ), and their travel speed can be very different for different regions. For example, a <i>Pn</i> wave, which travels along the top of mantle, can arrive 8 seconds earlier in some regions and 5 seconds later in other regions as compared to the global average <i>arrival time</i> for the same distance. See Richards (2002). |
| <b>regression analysis</b>          | A statistical technique applied to data to determine, for predictive purposes, the degree of <i>correlation</i> of a dependent variable with one or more independent variables, in other words, to see if there is a strong or weak cause-and-effect relationship between two or more parameters. More                                                                                                                                                                                                                                                                                                                                                                                                                                          |

specifically, standard regression analysis considers one of the two compared variables as independent (given and error free) and the other one as dependent and error afflicted, in contrast to correlation analysis which considers both variables as being in error. In this sense, the so-called *orthogonal regression*, preferable for deriving regression relationships between different types of *magnitude* data, is a correlation analysis between magnitude data with comparably large initial errors. See Bormann et al. (2007, 2009) and Castellaro and Bormann (2007).

**relative bandwidth**

The relative bandwidth (RBW) can be expressed by a number or in terms of *octaves* or *decades*. Increasing the *frequency* of a *signal* by one octave means doubling its frequency, and by one decade multiplying it by ten. Accordingly, a band-passed signal (or *filter*) with  $n$  octaves or  $m$  decades has a ratio between the upper and lower *corner frequencies*  $f_u/f_l = 2^n = 10^m$  and a *geometric center frequency*  $f_o = (f_u \times f_l)^{1/2} = f_l \times 2^{n/2} = f_l \times 10^{m/2}$ . From this follows for the relative bandwidth

$$RBW = (f_u - f_l) / f_o = (2^n - 1) / 2^{n/2} = (10^m - 1) / 10^{m/2}.$$

Note that signal and *noise amplitudes* can be made commensurate only when plotting them in a constant RBW over the whole considered frequency range. See Chapter 4 of this Manual.

**relaxation theory**

A concept in which radiated *seismic energy* is released from stored *strain energy* during the *slip* along a *fault* until the adjacent fault blocks reach a new state of equilibrium.

**renewal process**

A point process in which the statistical distribution of the intervals between consecutive events are statistically independent and stationary. See Utsu (2002b).

**residual**

The difference between measured and predicted values of some quantity.

**residual strength**

Shear resistance within cohesive soils at very large shear *deformations* or within contractive granular soils in the liquefied state. See Youd (2003).

**resonance**

Strong increase in the *amplitude* of vibration in an elastic body or system when the *frequency* of the shaking force is close to one or more times of the *natural frequencies* of a shaking body.

**response**

The motion in a system resulting from shaking under specified conditions.

**response function**

See *amplitude response* and *frequency response*.

- response-spectral ordinate** A term in *engineering seismology*, indicating the *amplitude* of a *response spectrum* at a specified value of undamped *natural period* or *frequency*. See Campbell (2003).
- response spectrum** A curve showing the computed maximum *response* of a set of simple *single-degree-of-freedom* harmonic *oscillators* with chosen levels of viscous *damping* and different *natural frequencies* to a particular record of *ground acceleration*. Response spectra, commonly plotted on tripartite logarithmic graph paper, show the oscillator's maximum acceleration, *velocity*, and *displacement* as a function of oscillator frequency for various levels of oscillator's damping. A computational approximation to the response spectrum is referred to as the pseudo-relative velocity response spectrum (PSRV). These curves are used by engineers to estimate the maximum response of simple structures to complex *ground motions*. See Anderson (2003).
- retrograde branch** See *branch (of travel-time curve)*.
- return period** The average time between exceedance of a specified level of *ground motion* at a specific location; it is equal to the inverse of the annual probability of exceedance. See also *recurrence interval*.
- reverse fault** A *fault* that involves lateral shortening, where one block (the *hanging wall*) moves over another (the *footwall*) up the *dip* of the fault, i.e., the movement of the hanging wall is upward relative to the footwall. The maximum *compressive stress* is horizontal, and the least compressive stress is vertical. Reverse faults that dip less than about 30 degrees are often called "*thrust faults*". See also *normal fault*, and Jackson (2002).
- reverse movement** See *fault*.
- Rg** Originally a symbol for the fundamental-mode *Rayleigh waves* observed for continental paths in the *period* range 8 to 12 seconds (Ewing et al., 1957). Now applied to dispersed wave trains propagating in the Earth's *crust* with periods between about 0.1 and 4 second (e.g., Fig. 2.18 in Chapter 2 of this Manual and Figure 4 in Kulhanek, 2002, p. 337). For this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).
- rheological properties** The properties of rocks that describe their ability to deform and flow as a function of temperature, *pressure*, and chemical conditions.
- Richter magnitude (of an earthquake)** The magnitude value of an earthquake based on Richter's *local magnitude ( $M_L$ ) scale* (Richter, 1935). In

|                                                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
|------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                                                  | popular use, the term is often mistaken for any magnitude. See <i>magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>Richter scale</b>                                             | Also called <i>Richter magnitude scale</i> or <i>local magnitude</i> ( $M_L$ ), introduced by Richter (1935). See <i>magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>ridge push</b>                                                | Gravitational force acting on the edges of diverging <i>plates</i> at <i>mid-ocean ridges</i> , due to the uplift of the ocean-ridge crest caused by the hot <i>asthenosphere</i> underneath it.                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>rift (rift valley)</b>                                        | An extended feature marked by a <i>fault</i> -caused trough that is created in a zone of divergent crustal <i>deformation</i> . Examples are the East African Rift Valley and the rift that often exists along the crest of a <i>mid-ocean ridge</i> .                                                                                                                                                                                                                                                                                                                                                      |
| <b>rift zone</b>                                                 | An elongate system of structural weakness — defined by eruptive fissures, <i>craters</i> , <i>volcanic vents</i> and geologic <i>faults</i> — that has undergone extension (ground has spread apart) by movement, storage, and rise of <i>magma</i> .                                                                                                                                                                                                                                                                                                                                                       |
| <b>right-lateral fault</b>                                       | A <i>strike-slip fault</i> across which a viewer would see the block on the other side moves to the right. Also known as <i>dextral fault</i> . See <i>left-lateral fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>right-lateral movement</b>                                    | See <i>right-lateral fault</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>rigidity</b>                                                  | Also termed <i>shear modulus</i> , is a measure of the resistance of the material to pure shearing, i.e., to changing the shape but not the volume of the material. See section 2.2 and Fig. 2.2 in Chapter 2 of this Manual.                                                                                                                                                                                                                                                                                                                                                                               |
| <b>Ring of Fire</b>                                              | Another popular name for the <i>Circum-Pacific belt</i> whereabout 90% of the world's earthquakes occur. The next most seismic region (5-6 %) is the <i>Alpide belt</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>ring laser</b>                                                | An instrument that detects the Sagnac beat frequency of two counter-propagating light beams (see <i>Sagnac effects</i> ). This beat frequency is directly proportional to the <i>rotation rate</i> around the surface normal to the ring laser system.                                                                                                                                                                                                                                                                                                                                                      |
| <b>ripple-firing, delay-firing, millisecond delay initiation</b> | These are all terms used to characterize quarry or mine-blasting activity, in which a particular shot is carried out as a series of separate charges, fired in a sequence instead of all together. Ripple-firing is the term commonly used, but experts usually refer to delay-firing and in some cases to millisecond delay initiation (where a specifically timed pattern of charges is executed that may entail delays that are designed for a particular blasting objective, such as fragmenting rock to a pre-determined size and moving the fragments a pre-specified distance). See Richards (2002). |



|                                             |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|---------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>rise-time</b>                            | In earthquake source studies, it refers to the time required for the completion of <i>slip</i> at a point on the <i>fault plane</i> . In the <i>Haskell model</i> , it is the parameter of ramp-function used for describing the slip time history. See <i>Haskell model</i> , and Eqs. (3.5b) and (3.8) in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>rocking</b>                              | <i>Rotation</i> about a horizontal axis or, as often used by engineers, of a whole structure about a horizontal axis. The term “rocking” should be used only with full, graphically supported definitions inclusive of frequency band, according to Evans et al. (2009).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>root mean square</b>                     | Square root of the mean value of a set of squared values.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>root n improvement</b>                   | More precisely, “square root of n” improvement in <i>signal-to-noise ratio</i> , as expected from <i>delay-and-sum processing</i> of n independent recordings with <i>noise</i> that has uniform power. See Chapter 9 in this Manual and Douglas (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>rotation(s)</b>                          | See <i>displacement (rotational)</i> and <i>rotation vector</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>rotation angles (rotational motions)</b> | The <i>rotational displacements</i> indicated $\theta_x$ , $\theta_y$ , and $\theta_z$ in Figure 1 of Evans et al. (2010). Terms like “X-rotation” may be used as shorthand though “rotation angle $\theta_x$ ” is more formal. Authors are cautioned that the term “rotational displacement” has other meanings and should be defined (e.g., P. Spudich, written comm., 2008; B. R. Julian, oral comm., 2008). Rotation angles are generally given in units of radians. For example, “ $\theta_x$ ” is the right-handed rotation angle about the X axis. The SI abbreviation for “radians” is “rad,” and for “seconds” it is “s” (not “sec”); thus, units in the style “mrad/s” are preferred to “milli-radians/sec”. In continuum mechanics, rotation angle and axis of rotation define the <i>rotation tensor</i> . |
| <b>rotation, average (volume or area)</b>   | <i>Rotation</i> of an extended region or some extended portion of a structure, often inferred from measured <i>translational motions</i> at the periphery.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>rotation, block</b>                      | See block rotation.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>rotation, point</b>                      | See <i>point rotation</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>rotation rate</b>                        | In the linear approximation, the first time derivative of <i>rotational displacement</i> . “Angular velocity” and <i>rotational velocity</i> are acceptable synonyms for “rotation rate”. In the nonlinear theory, it is defined via the <i>rotation tensor</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>rotation tensor</b>                      | A mathematical object that may refer to many kinds of “physical” <i>rotations</i> , in particular to the rotational part in the polar decomposition of the <i>gradient</i> of the radius vector in the classical continuum, to the (independent) rotation of a point body in the <i>Cosserat continuum</i> , or to any other rotation, for                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |

instance, to the rotation of a rigid body, of a co-ordinate system etc. In the following discussion, it is an orthogonal tensor characterizing rotation of vectors and rigid bodies. If  $\mathbf{d}$  is a vector subjected to rotation, the rotated vector equals  $\mathbf{d}' = \mathbf{P} \mathbf{d}$ , where  $\mathbf{P}$  is the rotation tensor,  $\det \mathbf{P} = 1$ . Each rotation tensor may be represented via the *rotation angle* and axis of rotation (theorem by Leonhard Euler):  $\mathbf{P} = (1 - \cos \theta) \mathbf{n} \mathbf{n} + \cos \theta \mathbf{I} + \sin \theta \mathbf{I} \times \mathbf{n}$ , where  $\mathbf{I}$  is the identity tensor ( $\mathbf{I} \mathbf{d} = \mathbf{d}$  for each vector  $\mathbf{d}$ ),  $\mathbf{n}$  is the axis of rotation. For linear motions,  $\mathbf{P}$  is approximately equal to  $\mathbf{I} + \theta \mathbf{n} \times \mathbf{I}$ . In the literature, the rotation tensor is also called the “turn tensor”. In *micropolar* media, one has to distinguish the rotation of the background continuum, related to the gradient of translational deformation, and the proper rotation of particles. In linear elasticity and seismology the rotation tensor refers to a different object, and the expression for  $\mathbf{P}$  is simplified by the infinitesimal nature of the *rotation angle*, not the linear motion. See Pujol (2009).

- rotation vector** In continuum mechanics, it is the vector  $\theta \mathbf{n}$ , where  $\theta$  is the *rotation angle* and  $\mathbf{n}$  is the unit vector along the axis of rotation. It corresponds to the *rotation tensor*,  $\mathbf{P} = (1 - \cos \theta) \mathbf{n} \mathbf{n} + \cos \theta \mathbf{I} + \sin \theta \mathbf{I} \times \mathbf{n}$ .
- rotational acceleration** The time derivative of the *rotational velocity*. In the linear approximation, the second time derivative of a *rotational displacement*. “Angular acceleration” is often used for “rotational acceleration.”
- rotational displacement** It is often used as a synonym for *rotation vector*.
- rotational ground motion** The rotational components of *ground motion*. In classical elasticity and assuming infinitesimal *deformations*,  $\boldsymbol{\omega} = \frac{1}{2} (\nabla \times \mathbf{u}(\mathbf{x}))$ , where  $\boldsymbol{\omega}$  is a *pseudo vector* and represents the angle of rigid rotation generated by the disturbance,  $\nabla \times$  is the *curl* operator, and  $\mathbf{u}$  is the *displacement* of a point at  $\mathbf{x}$ . See e.g., Aki and Richards (2002 and 2009, p. 13), and Cochard et al. (2006, p. 394).
- rotational-motion sensor** The term rotational-motion sensor is widely used instead of *rotational sensor* in aerospace, automotive, and mechanical engineering. In earthquake studies, the term *rotational seismometer* is preferred.
- rotational response spectrum** A plot of maxima of rotational responses of *oscillators* as a function of their *natural periods* (or *frequencies*). It is generalized from the conventional, translational *response spectrum* (see, e.g., Kalkan and Graizer 2007, Zembaty 2009a).
- rotational seismology** An emerging field of inquiry for studying all aspects of *rotational ground motions* induced by earthquakes, explosions, and ambient vibrations.

|                                    |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
|------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>rotational seismometer</b>      | A <i>rotational sensor</i> specifically designed as a <i>seismometer</i> for measuring <i>rotational ground motions</i> associated with <i>seismic waves</i> from <i>earthquakes</i> or explosions.                                                                                                                                                                                                                                                                                                                                                                               |
| <b>rotational seismic solitons</b> | A finite number of rotational pulses excited in the earthquake source; they can have a form of spin or twist <i>solitons</i> . See Majewski (2006; 2008).                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>rotational seismic waves</b>    | They are <i>rotational waves</i> excited in the <i>earthquake source</i> as a result of rotational vibrations, including rotational longitudinal waves, and rotational transverse waves. See Majewski (2006).                                                                                                                                                                                                                                                                                                                                                                     |
| <b>rotational sensor</b>           | A <i>rotational sensor</i> (often generically known as a <i>gyroscope</i> or <i>inertial angular sensor</i> ) that can measure angular displacement, velocity (often called “rate”), acceleration, or jerk (rate of change of acceleration). Such sensors may or may not have <i>response</i> down to zero frequency. See <i>rotational-motion sensor</i> .                                                                                                                                                                                                                       |
| <b>rotational velocity</b>         | The first-time derivative of <i>rotation angles</i> ; an acceptable synonym for <i>rotation rate</i> and spin, if the latter term refers to the antisymmetric part of the velocity gradient tensor. In the nonlinear case, rotational velocity is defined as vector $\boldsymbol{\omega}$ satisfying the equation $d\mathbf{P}/dt = \boldsymbol{\omega} \times \mathbf{P}$ , where $\mathbf{P}$ is the <i>rotation tensor</i> , $t$ is time.                                                                                                                                      |
| <b>rotational waves</b>            | Motions of a continuum, where the rotational dynamics is present. Term sometimes used in connection with <i>micropolar</i> media to describe rotation waves in a medium in which each point has degrees of freedom related to rotational motions. It is important not to confuse rotational waves with <i>rotational ground motions</i> appearing as the <i>curl</i> of the <i>wavefield</i> in linear elasticity with a symmetric <i>stress tensor</i> (e.g., <i>S waves</i> ).                                                                                                  |
| <b>ROV</b>                         | An acronym for Remotely Operated Vehicle. See Suyehiro and Mochizuki (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>RSAM</b>                        | An acronym for Real-time Seismic Amplitude Measurement; widely used by many volcano observatories, the RSAM system enables the continuous monitoring of total <i>seismic energy</i> release during rapidly escalating <i>volcanic unrest</i> , when conventional seismic-monitoring systems are saturated. See Endo and Murray (1991) for details. Both the RSAM and SSAM systems, together with associated PC-based software for data collection and analysis, form the core of an integrated “mobile volcano observatory” (Murray et al., 1996), which can be rapidly deployed. |
| <b>run-up</b>                      | The maximum height above a reference sea level of the water brought onshore by a <i>tsunami</i> . See <i>tsunami run-up height</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                              |

|                                            |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|--------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>rupture duration</b>                    | Time length of the rupture process.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>rupture front</b>                       | The instantaneous boundary between the <i>slipping</i> and <i>locked</i> parts of a <i>fault</i> during an earthquake. Rupture front propagation along the <i>fault length</i> of a <i>rectangular fault</i> from one end to the other is referred to as unilateral. If it starts in the middle of the fault and propagates toward its two ends, it is called bilateral. Rupture may expand outward in a circular or an elliptic form from its center. See <i>rectangular fault</i> , <i>circular fault</i> , <i>elliptic fault</i> , and Figures 9 and 10 in Ben-Zion (2003). |
| <b>rupture velocity</b>                    | The velocity $v_R$ at which a <i>rupture front</i> moves along the surface of the <i>fault</i> during an earthquake. Depending also on <i>stress drop</i> , $v_R$ may vary in a wide range between some 20% and 95% of the <i>shear-wave velocity</i> $v_S$ , yet being $> 60\%$ $v_S$ for most of the large earthquakes. Occasionally, $v_R > v_S$ has been reported from so-called “super shear” earthquakes, whereas for <i>tsunami</i> earthquakes $v_R$ may become $< 30\%$ of $v_S$ . See Kanamori and Brodsky (2004).                                                   |
| <b>Rytov method</b>                        | A method to find an approximate solution for the logarithm of the <i>wave field</i> (complex phase function) for waves in a smoothly varying <i>inhomogeneous</i> medium. In the Rytov approximation the second order term of the <i>gradient</i> of the complex phase function is often neglected. See Sato et al. (2002).                                                                                                                                                                                                                                                    |
| <b>S</b>                                   |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>S</b>                                   | General symbol for an <i>S-wave</i> . For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).                                                                                                                                                                                                                                                                                                                                                                                                        |
| $\bar{S}$                                  | See <i>Sg</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Sagnac effect (Sagnac interference)</b> | An interferometric phenomenon encountered in rotating systems, as in ring-laser and fiber-optic <i>gyroscopes</i> . In seismology most often described and measured in the <i>frequency domain</i> (a beat <i>frequency</i> ) rather than in the spatial domain (interference fringes).                                                                                                                                                                                                                                                                                        |
| <b>sag pond</b>                            | A <i>fault sag</i> that contains water.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>sand boil</b>                           | Sand and water ejected to the ground surface during strong earthquake shaking as a result of <i>liquefaction</i> at shallow depth; the conical sediment deposit remains as evidence of liquefaction. See Bardet (2003).                                                                                                                                                                                                                                                                                                                                                        |
| <b>satellite interferometry (InSAR)</b>    | Modern technology to measure areal <i>crustal deformations</i> and to derive seismic <i>fault plane</i> solutions after                                                                                                                                                                                                                                                                                                                                                                                                                                                        |

strong earthquakes by interferometric superposition of images taken from synthetic aperture radar (SAR) satellites before and after the earthquake. See *interferometry*, Feigl (2002) and color plate 14 in Lee et al. (2002); also Fig. 3.14 and related text in Chapter 3 of this Manual.

**SAR** Abbreviation for Synthetic Aperture Radar. See Feigl (2002) and *satellite interferometry (InSAR)*.

**saturation (of a magnitude scale)** Above a certain *magnitude* level, the magnitude determined from the *amplitudes* or amplitude/*period* ratios of specified *seismic waves* recorded by a specific *seismograph* increases only slowly or does not further increase as the physical size of the earthquake (such as measured by the *moment magnitude*  $M_w$ ) increases. This behavior, sometimes called saturation, is due to the shape of the *displacement* spectrum of a typical *seismic signal*, which is characterized by a *corner frequency* above which the displacement amplitude decreases rapidly. As the size of the earthquake increases, the corner frequency moves to lower frequencies. When the frequency at which a specific type of magnitude is calculated falls above this corner frequency, that magnitude scale tends to underestimate systematically the event magnitude. Saturation sets in at smaller magnitudes for scales based on records from short-period seismographs. Accordingly, magnitudes  $mb > 6.5$ ,  $M_L > 7.0$ ,  $mB > 7.5$ , or  $M_s > 8.0$  are rarely reported. The new IASPEI (2005 and 2011) standards delay magnitude saturation for  $mb$  and  $mB$  to values above 7.0 and 8.0, respectively. The new IASPEI (2005 and 2013) standards delay magnitude saturation for  $mb$  and  $mB$  to values above 7.0 and 8.0, respectively. No saturation occurs for  $M_w$  or *mantle magnitude*  $M_m$ , because these scales (if properly applied) always use frequencies well below the corner frequency. See Chapter 3 of this Manual as well as Bormann et al. (2009) and Bormann (2011).

**scalar seismic moment** See *seismic moment*.

**scale invariance** Property of a phenomenon which appears identical at a variety of scales. See Turcotte and Malamud (2002).

**scaling law of seismic spectrum** The law governing the *frequency*-dependent growth of the seismic spectrum with the increase of earthquake *magnitude*. The *saturation of a magnitude scale* is one of its consequences. It is also important for (roughly) estimating strong *ground motion* for a large earthquake based on the observed data for smaller ones. However, depending on the actual *stress drop*, the *corner frequencies* may vary for equal *seismic moment* by about one order. This explains the sometimes large differences between *seismic moment magnitude*  $M_w$  and *energy magnitude*  $M_e$ . See Aki and

Richards (2002 or 2009, p. 513-516), Chapter 3 in this Manual, Madariaga and Olsen (2002, p. 185-186), and Eq. (6.1) in Ben-Zion (2003), as well as Fig. 3.5 in this Manual or Fig. 1 in Bormann et al. (2009) and Fig. 2 in Bormann and Di Giacomo (2011).

|                                       |                                                                                                                                                                                                                                                                                                                                                                                         |
|---------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>scattering</b>                     | In physics, it refers to an alteration in direction of motion of a particle because of collision with another particle. In seismology, the particle is replaced by a wave, and it represents the change in wave properties such as <i>waveforms</i> and <i>ray</i> directions by inclusions or <i>heterogeneities</i> in the material properties of the medium. See Sato et al. (2002). |
| <b>scattering coefficient</b>         | Scattering power per unit volume of <i>heterogeneous</i> media having a dimension of the reciprocal of length. This quantity characterizes the <i>coda</i> excitation of a <i>local earthquake</i> and the strength of <i>scattering attenuation</i> in heterogeneous media. See Sato et al. (2002).                                                                                    |
| <b>scattering theory</b>              | Theory which develops models and algorithms suitable for synthesizing and (by inversion) interpreting observed scattering features in particle and wave physics. For seismic <i>scattering</i> , see Sato and Fehler (1998), and Sato et al. (2002).                                                                                                                                    |
| <b>scoria</b>                         | A porous glassy <i>pyroclast</i> , often containing more than 50 % void space, between 2 and 64 mm in size; it is generally darker in color and denser than <i>pumice</i> .                                                                                                                                                                                                             |
| <b>screw dislocation</b>              | See <i>dislocation</i> .                                                                                                                                                                                                                                                                                                                                                                |
| <b>ScS</b>                            | A symbol for <i>S</i> waves traveling through the <i>mantle</i> until reaching the <i>core-mantle boundary</i> and reflected there back to the surface. See Kulhanek (2002, p. 341), Cara (2002). For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual, and Storchak et al. (2003 and 2011).                                                    |
| <b>sea-floor spreading</b>            | A term introduced by Dietz (1961). It is now taken to be the creation of new oceanic <i>lithosphere</i> as <i>plates</i> separate at <i>mid-ocean ridges</i> , where upwelling <i>mantle</i> materials fills the growing gaps between the moving plates. See Uyeda (2002), and <i>magnetic reversal</i> .                                                                               |
| <b>sea-floor spreading hypothesis</b> | A hypothesis proposed by Dietz (1961) and Hess (1962) for the evolution of continents and oceans by the spreading of sea floor from <i>mantle convection</i> . See <i>sea-floor spreading</i> .                                                                                                                                                                                         |
| <b>second (s)</b>                     | The base unit of time in the <i>international system of units (SI)</i> . The official definition is “The second is the duration of 9 192 631 770 periods of the radiation corresponding to the transition between the two hyperfine levels of the ground state of the cesium-133 atom.” See NBS (1981).                                                                                 |

|                                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|--------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>sector collapse</b>               | The sudden failure of a significant part of the summit and (or) flank of a volcano, typically triggered by intrusion of new <i>magma</i> , <i>phreatic</i> explosions, or an earthquake. Also called flank failure or edifice collapse, such processes result in associated <i>landsliding</i> and <i>debris avalanches</i> , often creating deposits with distinctive hummocky topography (e.g., at Mount St. Helens in May 1980). At island volcanoes, large sector collapses can produce destructive <i>tsunamis</i> . See <i>hummocks</i> . |
| <b>secular</b>                       | Referring to long-term changes that take place slowly and imperceptibly. Commonly used to describe changes in elevation, tilt, and <i>stress</i> or <i>strain</i> rates that are related to long-term <i>tectonic deformation</i> . For example, a mountain that is growing is getting taller so slowly that we cannot see it happen, but if we measure the elevation in one year and then after, e.g., another time ten years later, we could see that it has grown taller.                                                                    |
| <b>sedimentary rock</b>              | A rock formed from the consolidation of sediment. The sediment can be physically deposited by wind, water or ice, or chemically precipitated from bodies of water. See Press and Siever (1994).                                                                                                                                                                                                                                                                                                                                                 |
| <b>segmentation</b>                  | The breaking up of a <i>fault</i> along its length into several smaller faults. This can happen as a result of other faults crossing it, topography changes, or bends in the <i>strike</i> of the faults. Segmentation can limit the length of faulting in a single earthquake to some fraction of the total <i>fault length</i> , thus also limiting the maximum possible size of an earthquake on that fault.                                                                                                                                 |
| <b>segment or segmentation model</b> | The assumption that earthquake <i>faults</i> are divided into discrete, identifiable sections that behave distinctively over multiple rupture cycles. See Grant (2002).                                                                                                                                                                                                                                                                                                                                                                         |
| <b>seiche</b>                        | A free oscillation ( <i>resonance</i> ) of the surface of an enclosed body of water, such as a lake, pond, or bay with a narrow entrance. They are sometimes excited by <i>earthquakes</i> and by <i>tsunamis</i> . The period of oscillation ranges from a few minutes to a few hours, and the oscillation may last for several hours to one or two days.                                                                                                                                                                                      |
| <b>seismic albedo</b>                | The ratio of the <i>scattering coefficient</i> to the total <i>attenuation</i> coefficient, i.e., the sum of absorption coefficient and scattering coefficient. This term derives from the original term “albedo” in astrophysics. See Sato et al. (2002).                                                                                                                                                                                                                                                                                      |
| <b>seismic anisotropy</b>            | A term used for effective material parameters which depend on the direction of propagation or <i>polarization</i> of <i>seismic waves</i> . It includes <i>isotropic</i> media with aligned small-scale                                                                                                                                                                                                                                                                                                                                         |

*heterogeneities* that may behave *anisotropic* for seismic *wavelengths* longer than the heterogeneity scale. See Curtis and Snieder (2002), and Cara (2002).

**seismic array**

An ordered arrangement of *seismometers* distributed over the surface or volume (with bore-hole sites) of the Earth whose outputs are transmitted to and recorded at a central *data acquisition* and *data processing* unit. Since the 1950's many arrays have been constructed worldwide for detecting underground nuclear tests and for fundamental research in seismology. See Chapter 9 of this Manual, Douglas (2002), *array analysis*, *array aperture*, and *array beam*.

**seismic belt**

An elongate earthquake zone, for example, the *Circum-Pacific*, Mediterranean, Rocky Mountains earthquake *belt*.

**seismic bulletin**

A compilation or listing of parametric seismic data from different stations (mainly *phase arrival times*, *amplitudes* and *periods* of waves) associated with specific *seismic events*. When enough and consistent data are available, event locations and *magnitudes* are also computed and listed. The most complete global seismic bulletin for 1964 to the present is the "Bulletin of the International Seismological Centre (ISC)". Between 1913 and 1964 the most relevant global bulletins are those of the International Seismological Summary (ISS) and the Bureau Central International de Seismologie (BCIS). See Engdahl and Villasenor (2002), and Schweitzer and Lee (2003).

**seismic catalog**

A compilation or listing of earthquake focal parameters. The basic focal parameters reported by most seismic catalogs are: earthquake *origin time*, latitude, longitude, depth, and some indicator of the size (e.g., *magnitude*). Unlike *seismic bulletins*, seismic catalogs do not include the primary data measured at the *seismic stations* (mostly *phase arrival times*, *amplitudes* and *periods*) that were used to obtain the *hypocenter* location or the other focal parameters. See *catalog*.

**seismic constant**

In building codes dealing with *earthquake hazard*, an arbitrarily set quantity of steady *acceleration* (in units of *gravity g*) that a building must withstand.

**seismic core phase**

A general term referring to a *seismic body-wave phase* that enters the *Earth's core*. See section 2.6.2 in Chapter 2, sections 11.5.2 and 11.5.3 in Chapter 11, IS 2.1, Data Sheet DS 11.3 and exercise EX 11.3 of this Manual; also Song (2002).

**seismic efficiency**

For an earthquake, the ratio of the energy radiated in the *seismic waves* to the total energy released due to *fault slip*, or, equivalently, the ratio of *apparent stress* to the average stress acting on the fault to cause slip. Savage and Wood (1971) suggested that the seismic efficiency of earthquakes should not



exceed 0.15 and a likely value would be about 0.07. The seismic efficiency is less than the *radiation efficiency*, because the latter neglects energy loss due to frictional heat and conversion of *strain* energy into acoustic energy. See Eq. (6.5) in Ben-Zion (2003); also Kanamori and Brodsky (2004) and Bormann and Di Giacomo (2011). See *radiation efficiency*.

**seismic energy**

That part of the potential energy released during an *earthquake*, an explosion or another type of *seismic source* that is converted into elastic *seismic waves* that propagate away from the source through the Earth and cause vibrating *ground motions*. The ratio between seismic energy  $E_s$  and *seismic moment*  $M_o$  is proportional to the *stress drop*  $\Delta\sigma$  in the seismic source. (See also *radiation efficiency*).

**seismic event**

A general term used for localized sources of different type which generate *seismic waves*, such as *earthquakes*, rock bursts, explosions (man-made or natural ones), impacts and cavity collapses.

**seismic exploration**

Seismic exploration uses *seismic waves* of different types (mostly *reflected*, *refracted* and *converted body waves*) of usually higher *frequencies* between some 1 to 100 Hz and thus shorter *wavelengths* and higher spatial resolution than commonly recorded and analyzed in earthquake seismology. These high-frequency seismic waves are typically generated by controlled *seismic sources* such as explosions, vibrators, and *airguns* and recorded by *geophones* or *hydrophones*. S. e. aims at detailed imaging of layered, folded and faulted structures in the Earth's crust. Commercial s. e. is usually termed *seismic prospection* and aims at finding structures with potential for exploitable natural deposits of oil, gas, aquifers etc. in the uppermost *crust*.

**seismic gap**

An area of a *seismic zone* in which large earthquakes are known to have occurred in the past, but in which none has occurred for a defined long time. Seismic activity is low compared to the neighboring areas of the same zone. Such an area is often considered to be the site of the next large earthquake following the idea of Fedotov (1965) who found that great earthquakes in the northern Pacific tended to occur in regions lacking earthquakes for several decades, and tended to rupture discrete segments which eventually fill the whole seismic zone without overlapping. See Kanamori (2003).

**seismic hazard**

"Hazard H is the probability of occurrence of an event with disaster potential within a defined area and interval of time". It is also used - without regard to a *loss* - to indicate the probable level of ground shaking occurring at a given point within a certain period of time. See *earthquake hazard*, Somerville and Moriwaki (2003), and Giardini et al. (2003).

|                                      |                                                                                                                                                                                                                                                                                                                                                                                                              |
|--------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>seismic hazard analysis (SHA)</b> | The calculation of the <i>seismic hazard</i> , expressed in probabilistic terms (see <i>probabilistic seismic hazard analysis</i> ), as contrasted with deterministic seismic hazard analysis, for a site or group of sites. The result is usually displayed as a <i>seismic hazard curve</i> or <i>seismic hazard map</i> . See Somerville and Moriawaki (2003), Giardini et al. (2003), and Reiter (1990). |
| <b>seismic hazard curve</b>          | A plot of <i>probabilistic seismic hazard</i> (usually specified in terms of annual probability of exceedance) or average return period versus a specified <i>ground-motion parameter</i> for a given site.                                                                                                                                                                                                  |
| <b>seismic hazard map</b>            | A map showing contours of a specified <i>ground-motion parameter</i> or <i>response spectrum</i> ordinate for a given <i>probabilistic seismic hazard</i> or average return period.                                                                                                                                                                                                                          |
| <b>seismic impedance</b>             | <i>Seismic wave velocity</i> multiplied by <i>density</i> of the medium. The contrast in seismic impedance across two media is the main factor governing the <i>reflection coefficient</i> at their boundary (see <i>impedance contrast</i> ).                                                                                                                                                               |
| <b>seismic intensity</b>             | The degree of ground shaking estimated from its effect on people, houses, construction, and natural objects. The intensity scales are essential for assigning <i>magnitude</i> to earthquakes for which instrumental data are not available. See Chapter 12 of this Manual, <i>intensity scales</i> , <i>intensity (macroseismic)</i> , and Musson and Cencić (2002).                                        |
| <b>seismicity</b>                    | A term (used already by Willis, 1923) to describe quantitatively the space, time and <i>magnitude</i> distribution of earthquake occurrences. Seismicity within a specific source zone or region is usually quantified in terms of a <i>Gutenberg-Richter relationship</i> . See Utsu (2002b).                                                                                                               |
| <b>seismic lid</b>                   | The <i>mantle</i> part of the seismic <i>lithosphere</i> overlying the <i>asthenosphere</i> . See Lay (2002).                                                                                                                                                                                                                                                                                                |
| <b>seismic line</b>                  | A set of <i>seismographs</i> usually lined up along the Earth's surface to record <i>seismic waves</i> generated by artificial or natural <i>seismic sources</i> for studying the internal structure of the Earth.                                                                                                                                                                                           |
| <b>seismic migration</b>             | In <i>exploration seismology</i> , a process which collapses <i>diffractions</i> and moves <i>reflections</i> to their proper location in the seismic image. See Sheriff and Geldart (1995).                                                                                                                                                                                                                 |
| <b>seismic moment</b>                | The magnitude of the component couple of the <i>double couple</i> that is the point force system equivalent to a <i>fault slip</i> in an <i>isotropic</i> elastic body. It is equal to <i>rigidity</i> times the fault slip integrated over the <i>fault plane</i> . It can be estimated from the <i>far-field</i> seismic spectrum or <i>waveform</i> fitting at <i>wavelengths</i> much                    |

|                               |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
|-------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                               | longer than the source size. It can also be estimated from the <i>near-field</i> seismic, geologic and geodetic data, and the consistency among various observations. This was first demonstrated for the Niigata, Japan, earthquake of 1964, supporting quantitatively the validity of the fault origin of an earthquake. Also called “ <i>scalar seismic moment</i> ” to distinguish it from <i>moment tensor</i> . See <i>slip moment</i> and <i>geodetic moment</i> . See section 3.5 of Chapter 3 and exercises 3.4 and 3.5 in this Manual, Aki and Richards (2002 or 2009, p. 48-49), and section 2 in Ben-Zion (2003).                                                      |
| <b>seismic multiples</b>      | Seismic <i>arrivals</i> arising out of <i>reflections</i> between subsurface layers.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>seismic network</b>        | A network of <i>seismographs</i> deployed for a specific purpose. Modern means of regional and global data communication permit to design task-tailored temporary virtual seismic networks. See Chapters 8, 9 and 11 in this Manual; also Lahr and van Eck (2003).                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>seismic noise</b>          | See <i>microseisms</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>seismic phase</b>          | A pulse, <i>wavelet</i> or group of seismic <i>body-</i> or <i>surface-wave</i> energy arriving at a definite time, which passed the Earth on a specific path (see also Chapters 2 and 11, IS 2.1, DS 11.1 to 11.3, Storchak et al. (2003 and 2011) and <i>phase nomenclature</i> .                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>seismic potency</b>        | A quantity (area times <i>slip</i> ) that characterizes the inelastic <i>strain</i> in a faulting region. See Ben-Menahem et al. (1965), and section 2 in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>seismic prospection</b>    | See <i>seismic exploration</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>seismic rays and tubes</b> | A seismic ray is a path in a 3D space along which <i>seismic energy</i> can be considered to propagate like a particle. For <i>high-frequency seismic signals</i> , this is often a useful approximation and simplification. The ray vector is oriented perpendicular to the <i>wavefront</i> , pointing into the direction of wave propagation, and marking behind it the “ray trace”. The propagation of <i>seismic waves</i> can be easily modeled as the propagation of seismic rays following <i>Snell’s Law</i> (see section 2.5.2 in Chapter 2 of this Manual). Neighboring rays form a ray tube and the energy flux is confined within its walls. See also Chapman (2002). |
| <b>seismic recorder</b>       | A device for recording seismic <i>ground motions</i> in analog or digital form (see <i>seismograph</i> ), usually as a <i>frequency-filtered</i> and amplified equivalent electronic <i>signal</i> (see <i>transducer</i> and <i>data logger</i> ). See Chapter 6 of the Manual.                                                                                                                                                                                                                                                                                                                                                                                                   |

**seismic refraction/wide-angle reflection profiles** Also referred to as Deep Seismic Sounding (DSS) profiles. *Seismic velocity* structures in the *lithosphere* derived from data recorded with long offsets (100-300 km) between sources and receivers consisting of widely-spaced *geophones* (100-5,000 m). See Mooney et al. (2002).

**Seismic Research Observatories (SRO)** Name and acronym of a former global U.S. *seismic network*, operating between 1974 and 1993. The long-period SRO *responses* were very specific: a narrow 20 s bandpass with a 6 s noise notch-filter (see in Chapter 11 of this Manual Fig. 11.38 and Tab. 11.3 for the poles and zeros). Many record examples, in comparison with those from other *seismographs*, are given in Chapter 11, DS 11.2 and 11.3 of this Manual. The SRO-LP response has advantages for measuring the IASPEI standard *Ms\_20 magnitude*. See Chapter 3 and IS 3.3 of this Manual, and Hutt et al. (2002).

**seismic risk** The risk to life and property from earthquakes. In probabilistic risk analyses, the probability that a specified loss will exceed some quantifiable level during a given exposure time. See *earthquake risk*, and Somerville and Moriawaki (2003).

**seismic risk analysis (SRA)** The calculation of *seismic risk* for a given property or portfolio of properties, usually performed in a probabilistic framework and displayed as a *seismic risk curve* or *seismic hazard map*. See Somerville and Moriawaki (2003).

**seismic risk curve** A plot of seismic risk (usually specified in terms of annual probability of exceedance or return period) versus a specified loss for a given property or portfolio of properties.

**seismic sensor** There are two basic types of seismic sensors: (1) *inertial seismometers* which measure *ground motion* relative to an inertial reference (a suspended mass), and (2) *strainmeters* or *extensometers* which measure the motion of one point of the ground relative to another. Inertial seismometers are generally more sensitive to earthquake *signals*, whereas strainmeters may outperform inertial seismometers when observing very long-period *free oscillations* of the Earth, tidal motions, and quasi-static *deformations* when it becomes increasingly difficult to maintain an inertial reference. See Chapter 5, DS 5.1 and IS 5.1 of this Manual.

**seismic sequence** Unconformity-bounded sequence of strata recognizable on seismic cross sections.

**seismic signal** A *coherent* transient *waveform* radiated from a definite, localized *seismic source* that is usually considered as a useful signal for the location of the source, the analysis of the source process and/or of the propagation medium (in contrast to *noise*). See *signal*.

|                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|--------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>seismic source</b>          | A localized area or volume generating <i>coherent</i> , usually transient seismic <i>waveforms</i> , such as an <i>earthquake</i> , <i>explosion</i> , vibrator etc. See Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>seismic source function</b> | See <i>source function</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>seismic spin solitons</b>   | A kind of <i>rotational seismic waves</i> in a form of a finite number of spin pulses excited in the earthquake source as a result of spin vibrations. See Majewski (2006; 2008).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>seismic station</b>         | A facility devoted to the measurement and recording of earthquake motions. A modern seismic station has a weak motion sensor (e.g., a broadband <i>seismometer</i> ), often a strong motion sensor (e.g., an <i>accelerometer</i> ), and a <i>data logger</i> with GPS timing, onboard storage, and real-time communication devices. See Chapter 7 of this Manual, Lee (2002) and Hauksson et al. (2003). See also <i>station</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>seismic tomography</b>      | An imaging method using a set of seismic <i>travel-time</i> , <i>amplitude</i> or <i>waveform</i> data obtained from <i>seismographs</i> distributed over the Earth's surface to deduce the 3D <i>velocity</i> and/or <i>attenuation/scattering</i> structure of the Earth's interior. In most cases, velocity structures are obtained from the travel-time data. Various computational techniques have been developed for data from different types of <i>seismic sources</i> including natural <i>earthquakes</i> , as well as controlled sources like explosions, <i>airguns</i> and bore-hole sources. However, it can be shown that collocated measurements of the ratio of <i>rotations</i> (or <i>strain</i> ) and corresponding components of <i>translational motions</i> contain also information on the subsurface velocity structure (e.g., Fichtner and Igel, 2009). See also Curtis and Snieder (2002). |
| <b>seismic trace</b>           | See <i>trace</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>seismic twist solitons</b>  | A kind of <i>rotational seismic waves</i> in a form of a finite number of twist pulses excited in the earthquake source as a result of twist vibrations. See Majewski (2006; 2008).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>seismic wave</b>            | A general term for waves generated by <i>earthquakes</i> , explosions or other types of <i>seismic sources</i> . There are many types of seismic waves. The principle ones are <i>body waves</i> , <i>surface waves</i> , and <i>coda waves</i> . See, e.g., Chapter 2 and IS 2.1 in this Manual, and Section 3 in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>seismic zonation</b>        | Geographic delineation of areas having different potentials for hazardous effects from future earthquakes. Seismic zonation can be done at any scale – national, regional, local or site, the latter two often referred to as <i>microzonation</i> . See <i>seismic zoning map</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |

|                                |                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|--------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>seismic zone</b>            | An area of <i>seismicity</i> probably sharing a common cause.                                                                                                                                                                                                                                                                                                                                                                             |
| <b>seismic zoning map</b>      | A map used to portray <i>seismic hazard</i> or seismic design variables. For example, maps used in building codes to identify areas of uniform seismic design requirements.                                                                                                                                                                                                                                                               |
| <b>seismite</b>                | Sedimentary units or features resulting from earthquake shaking. See Grant (2002).                                                                                                                                                                                                                                                                                                                                                        |
| <b>seismoelectric</b>          | Electric fields generated by <i>crustal stress</i> changes during earthquakes. See Johnston (2002).                                                                                                                                                                                                                                                                                                                                       |
| <b>seismogenic</b>             | Capable of generating earthquakes.                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>seismogenic thickness</b>   | The thickness of the <i>brittle</i> layer within the <i>lithosphere</i> in which earthquakes occur. On the continents this thickness is typically 10 to 15 km, corresponding to the region at temperature lower than 300-400 degrees C. See Jackson (2002).                                                                                                                                                                               |
| <b>seismogenic zone</b>        | That portion of the <i>crust</i> or <i>lithosphere</i> capable of generating earthquakes. See Sibson (2002).                                                                                                                                                                                                                                                                                                                              |
| <b>seismogeodetic method</b>   | A term introduced by Bodin et al. (1997) to describe the method used by Spudich et al. (1995) to infer the spatially uniform strain and rigid body <i>rotation</i> as functions of time that best fit ground translations recorded by a <i>seismic array</i> , essentially doing a geodetic inversion at every instant of time. The seismogeodetic method uses a generalized inversion of the data and accounts for the data covariances. |
| <b>seismogram</b>              | A record written by a <i>seismograph</i> in response to <i>ground motions</i> produced by an earthquake, explosion, or any other <i>seismic sources</i> that generate oscillating <i>ground motions</i> . Such a record may be analog or digital. See diverse figures in Chapter 2 and in DS 11.1 to 11.4 of this Manual, Lee (2002) and Kulhanek (2002).                                                                                 |
| <b>seismograph</b>             | Instrument that detects and records <i>ground motions</i> (and especially vibrations due to earthquakes) along with timing information. It consists of a <i>seismometer</i> , a precise timing element, and a recording device or <i>data logger</i> . See Chapters 5 in the Manual and Wielandt (2002).                                                                                                                                  |
| <b>seismological parameter</b> | Parameter used to characterize a seismological property of the source, propagation path, or site-response. See, e.g., Campbell (2003).                                                                                                                                                                                                                                                                                                    |
| <b>seismology</b>              | A branch of geophysics dealing with mechanical vibrations of the Earth caused by natural sources like earthquakes and                                                                                                                                                                                                                                                                                                                     |

|                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|--------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                          | volcanic <i>eruptions</i> , and controlled sources like underground explosions or any other <i>seismic source</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>seismomagnetic</b>    | Magnetic fields generated by <i>crustal stress</i> changes during earthquakes. See Johnston (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>seismometer</b>       | A sensor that responds to point <i>ground motions</i> or structural <i>acceleration</i> and usually produces a <i>signal</i> that can be recorded in terms of <i>displacement</i> , <i>velocity</i> , or acceleration of the seismometer mass with respect to the ground on which the seismometer is installed. A seismometer is usually a damped oscillating mass that is connected with a ground-fixed base and frame via a suspension, e.g., a spring (see <i>pendulum</i> with Figure 1). Thus the mass serves as an inertial reference for detecting and measuring seismic ground motion relative to the suspended mass. These relative motions are commonly directly recorded or transformed into an electrical voltage (see <i>transducer</i> ), which was traditionally recorded analog on paper, film or magnetic tape. Nowadays, digital recording media are almost exclusively used. The seismometer record can mathematically be converted to a record of the absolute ground motion. For <i>inertial seismometers (sensors)</i> , see Chapter 5 and DS 5.1 of this Manual. Another, non-inertial type of seismometers is <i>strainmeters</i> (see IS 5.1 in this Manual). Until recently, “seismometer” meant only a translational sensor that senses one of the three components of <i>translational motions</i> . More recently, a <i>rotational seismometer</i> may also be deployed to complement such translational seismometers and allow for measuring all six components of ground motion. |
| <b>seismometer array</b> | See <i>seismic array</i> and Chapter 9 in this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>seismometry</b>       | The science of preparing for, making and understanding the effects instrumental seismic measurements. See Chapter 5 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>seismoscope</b>       | A device that indicates some feature of <i>ground motion</i> during an earthquake, but without producing the time-dependent record. Seismoscopes are forerunners of <i>seismographs</i> and were in use until about 1910.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>self noise</b>        | Also termed <i>instrumental noise</i> , is the <i>frequency-dependent</i> electronic and mechanical noise at the output of a <i>seismometer</i> , <i>seismic recorder</i> or <i>seismograph</i> in the presence of zero input signal (Lee et al., 1982). The self noise of a seismographic system is typically measured at a seismically quiet site using a single sensor (if the sensor noise is higher than the seismic noise) or two or three sensors (if the sensors’ self noise is below the seismic noise). Weak-motion seismometers require two- or three-sensor methods to isolate self-noise from <i>ambient Earth noise</i> . Evans et al. (2010) provide a detailed definition and                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |

methods for measurement and analysis of *accelerometer* self noise and sections 5.5.5 to 5.5.7 in Chapter 5 of this Manual for different weak-motion seismometers. See also Sleeman et al. (2006).

- self-organized criticality** A system in a marginally steady state (at a critical point) in which the input is continuous and the output is a series of “avalanches” satisfying a *power-law (fractal) frequency-size* distribution. See Bak (1996) or Banzhaf (2002) or, for a general review, Wolfram (2002), and Turcotte and Malamud (2002) for application in seismology.
- sensitivity (of seismic sensors)** The scalar value used to convert the measured output *signal* (usually in voltage) to motion units. It is equivalent to the sensitivity scale factor of a normalized Laplace *transfer function*. For a sensor with a flat *frequency response*, the sensitivity corresponds to some mid-frequency level of the frequency-response *amplitude* for a specified frequency, often 1 Hz. Sensitivity has units of volts per input unit (e.g., volt per g for strong-motion *accelerometers*, volt per m/s for velocity meters or volt per rad/s for *rotational sensors*).
- sensitivity kernels** Term often used in connection with *seismic tomography*. Sensitivity kernels contain information about the possible resolution of a tomographic image. They are calculated as functions of space quantifying the influence of points in space on observable quantities (*arrival times*, translational or rotational *amplitudes*, or combinations of these) (or vice versa as functions of observable quantities vs. points in space).
- separation** A term in earthquake geology referring to the distance between any two parts of a reference plane (for example, a sedimentary bed or a geomorphic surface) offset by a *fault*, measured in any plane. Separation is the apparent amount of fault *displacement* and is nearly always less than the actual *slip*.
- serpentinization** A process by which the *upper mantle* reacts chemically with seawater to generate serpentine minerals. This process can cause substantial reductions in both *P*- and *S*-wave *velocities*. See Minshull (2002).
- Sg** *S waves* propagating through the *crust* like *Pg*. An older but not longer standard nomenclature for this phase is  $\bar{S}$ . At greater distances the *travel-time curve* of Sg is becoming (due to multiple paths all trapped in the crust) the *onset* of the *Lg-wave* group. Then both names are often alternatively used for the same seismic *onset*. For this and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).



|                               |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|-------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>SH, SV</b>                 | <p>S-wave components with <i>polarization</i> direction in the horizontal and vertical planes with respect to the propagation direction of the S-wave <i>ray</i>. The <i>amplitude</i> ratio between the SH and SV component of the S-wave particle motion recorded at a <i>seismic station</i> depends on the <i>slip</i> direction in the <i>double couple source</i> with respect to the <i>take-off angle</i> from the source of the <i>seismic ray</i> arriving at the station. For a vertically <i>heterogeneous</i> medium, the SV part of an S wave can be converted into a P wave and vice versa, The SH part of an S wave cannot interact with P-waves and is usually simpler than the SV part. In an <i>anisotropic</i> medium in which <i>shear-wave splitting</i> occurs, SH and SV are coupled, making this classification less meaningful or even useless.</p> |
| <b>shadow zone</b>            | <p>Distance range along the Earth's surface in which a <i>direct phase</i> is not observed as a result of downward deflection of the <i>ray</i> path by a <i>low velocity zone</i> at depth. See section 2.5.3.3 and Figure 2.33 in Chapter 2 of this Manual, and Lay (2002).</p>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>shake map</b>              | <p>A map of earthquake <i>ground motion</i> showing the geographical distribution of ground shaking as described by a specific parameter such as <i>peak acceleration</i>. See Hauksson et al. (2003).</p>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>shallow water wave</b>     | <p>A kind of oceanic gravity wave. When the <i>wavelength</i> is much larger than the water depth, it is called a long wave or a shallow water wave. Most <i>tsunamis</i> can be treated as shallow water waves. See Satake (2002).</p>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>Shanidar</b>               | <p>A cave in northern Iraq in which several Neanderthal skeletons were uncovered, dating back to 50,000 BP, presumably crushed by roof collapse caused by repeated earthquakes. They may be the oldest known casualties of earthquakes on Earth. See Nur (2002).</p>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>shear connection</b>       | <p>A connection between a beam and a column designed to transmit only the shear force from the beam to the column and not the moment. In a steel frame building shear connections typically have the web of the beam welded or bolted to the column, but show gaps between the flanges of the beam and the surface of the column. See Dowrick (1977), and Salmon and Johnson (1996).</p>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>shear-coupled PL-waves</b> | <p>This is a long-<i>period</i> wavetrain that follows S for distances up to about 80°. It has been explained as being due to the coupling of S waves with a <i>leaking mode</i> of the <i>crust-mantle</i> waveguide, i.e., PL-waves. The coupling of PL-waves with SS and SSS has also been observed.</p>                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |

|                                                            |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
|------------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>shear modulus</b>                                       | The ratio of <i>shear stress</i> to shear <i>strain</i> of a material during simple shear (see <i>rigidity</i> , Fig. 2.2 and explanations in Chapter 2 of this Manual, and Birch, 1966).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>shear stress</b>                                        | Force per unit area exerted upon an infinitesimal surface in the direction tangential to that surface. There is some confusion over the usage of this term (see text for discussion in Ruff, 2002). In three-dimensions, there are two independent force components tangential to an area element, thus further specification is required. Another definition of shear stress cites the difference between two <i>principal stresses</i> , while still another definition cites half the difference between two principal stresses. The latter definition is compatible with the Mohr's Circle representation (see <i>Mohr diagram</i> ) of the stress state. See <i>normal stress</i> ; also Ruff (2002), and Figure 1 in Ben-Zion (2003). |
| <b>shear wave</b>                                          | Another name for <i>S wave</i> . See Section 2.2.3 in Chapter 2 of this Manual, and Eq. (1.9b) in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>shear-wave splitting</b>                                | Behavior of <i>S waves</i> in an anisotropic medium which shows a split into two <i>shear waves</i> with mutually perpendicular <i>polarization</i> directions and different propagation <i>velocities</i> . See Cara (2002), IS 11.5 (p. 16) and Fig. 2.8 of Chapter 2 in this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>shield volcano</b>                                      | Volcanic landform with broad, gentle slopes built by countless <i>eruptions</i> of fluid <i>lava</i> ; classic example is Mauna Loa Volcano, Island of Hawaii. See <i>lava shield</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>short-period</b>                                        | In traditional <i>seismometry</i> with limited <i>dynamic range</i> , <i>seismographs</i> were naturally divided into two types; long-period seismographs for <i>periods</i> above the spectral peak (periods around 5 seconds) of the <i>microseismic noise</i> generated by ocean waves and short-period <i>seismographs</i> for periods below it. A short-period seismograph <i>response</i> usually has a bandwidth from about 1 to 10 Hz. See Fig. 3.20 in Chapter 3, Section 4.5 in Chapter 4 of this Manual, and Wielandt (2002).                                                                                                                                                                                                    |
| <b>short-period body-wave magnitude (<math>m_b</math>)</b> | See <i>body-wave magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>shutter ridge (also shutteridge)</b>                    | A ridge that has been displaced by <i>strike-slip</i> or <i>oblique-slip faulting</i> , thereby shutting off the drainage channels of streams crossing the <i>fault</i> (after Buwalda, 1937 and Sharp, 1954).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>SI</b>                                                  | An acronym for <i>Système International d'Unités</i> . See <i>international system of units</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |

|                                    |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Sigma-Delta modulation</b>      | A technique for <i>analog-to-digital conversion</i> commonly used in many modern seismic acquisition systems. See Chapter 6 in this Manual, and Scherbaum (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>signal</b>                      | The wanted part of given data. The remaining part is called <i>noise</i> . See <i>seismic signal</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>signal association</b>          | A major problem in detection and location of <i>seismic events</i> using a <i>seismic network</i> is to sort out lists of <i>arrival times</i> from each station and correctly identify the sources of overlapping events. This work is called signal or <i>event association</i> . It is challenging particularly for global networks in producing reliable <i>seismic bulletins</i> , which commonly locate more than 100 events per day. See Richards (2002).                                                                                                                                                                                                                                          |
| <b>signal detection</b>            | The detection of weak <i>signals</i> imposed upon <i>noise</i> can be accomplished by noting significant changes in <i>amplitude</i> , <i>frequency</i> content, <i>polarization</i> direction, signal shape, or wave propagation characteristics ( <i>backazimuth</i> and <i>slowness</i> ). If the signal to be detected has a <i>waveform</i> or a horizontal wave-propagation <i>velocity</i> that is known beforehand, for example, from a prior event in the same location and recorded at the same station, then <i>correlation</i> methods, matched filtering or velocity filtering can be applied for detection. See Chapter 9 and Section 4.4 in Chapter 4 of this Manual; also Richards (2002) |
| <b>signal-to-noise improvement</b> | A measure of how well a wanted <i>signal</i> is enhanced by processing, relative to the <i>ambient noise</i> and unwanted signals. Assuming that the processing leaves the signal unchanged, the signal-to-noise improvement is often expressed as the square root of the ratio of the noise power (or its average over channels for an <i>array</i> ) before processing to the power after processing. See Chapters 4 and 9 in this Manual, and Douglas (2002).                                                                                                                                                                                                                                          |
| <b>signal-to-noise ratio</b>       | The comparison between the <i>amplitude</i> of the <i>seismic signal</i> and the amplitude of the <i>noise</i> ; abbreviated as SNR. Often quantified as the ratio of the power (variance) of the signal to that of the noise for a given <i>frequency</i> band and measured in decibels (dB). See Chapters 4 and 9 in this Manual, Webb (2002), and Suyehiro and Mochizuki (2002).                                                                                                                                                                                                                                                                                                                       |
| <b>signature</b>                   | The appearance of a <i>seismic signal</i> that is more or less unique to the kind of <i>seismic source</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                              |
| <b>silent earthquake</b>           | <i>Fault</i> movement not accompanied by seismic radiation. See Johnston and Linde (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>similarity (of earthquakes)</b> | Similarity of earthquake <i>waveforms</i> is a phenomenological property with a great latitude of definitions. High similarity is interpreted as the result of closely spaced <i>seismic sources</i> ( <i>hypocenters</i> ), reflecting the similarity of the <i>Green's functions</i>                                                                                                                                                                                                                                                                                                                                                                                                                    |

that characterize the source-receiver paths, and very similar source-radiation patterns. See, e.g., Geller and Mueller (1980), and Baisch et al. (2008) and references therein. Two similar events are called doublets, while a group of more than two similar events is known as a multiplet. A common way to investigate and express the similarity of earthquakes is through the application of waveform cross-correlation or matched filtering techniques, the correlation coefficient being a measure of the level of similarity between waveforms. See *correlation*, *correlation coefficient*, *cross-correlation*, and Chapter 9 in this Manual.

- simulated annealing** A trick used to solve optimization problems, taking its roots in a thermodynamical analogy. Simulated annealing is based on the *Metropolis* algorithm. See Mosegaard and Tarantola (2002) and Tarantola (2005).
- single-degree-of-freedom (SDOF) system** An *oscillator* or structure composed of a single mass attached to a spring and dashpot. This system has a single mode or *period* of vibration whose *response* is described by a single variable. See Chapter 5 and Fig. 5.5 in this Manual, and Campbell (2003).
- single frequency microseisms** A secondary peak near 14-second *period* in *seismic noise* related to the action of ocean waves which is usually 20-40 dB smaller than the main peak due to the *double frequency microseism*. Called the single frequency peak because the seismic noise is at the same *frequency* as the ocean waves. See Chapter 4 in this Manual, and Webb (2002).
- sinistral fault** See *left-lateral fault*.
- site amplification** See *site response*.
- site category** Category of site geologic conditions affecting earthquake *ground motions* based on descriptions of the geology, or measurements of the *S-wave velocity* standard penetration test, shear strength, or other properties of the subsurface. For example, in the USA, geologic site conditions were classified by NEHRP into 6 categories ranging from A (hard rock) to F (very soft soil) and different amplification factors were assigned for them. See Kawase (2003).
- site classification** The process of assigning a site category to a site by means of geological properties (e.g., crystalline rock, *Quaternary* deposits, etc.), or by means of a geotechnical characterization of the soil profile, e.g., standard penetration test and *S-wave velocity*. See Borchardt et al. (2003).
- site effect** The effect of local geologic and topographic conditions at a recording site on *ground motions*. It is implicitly assumed that

|                               |                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|-------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                               | the source, <i>path</i> and site <i>effects</i> on ground motions are separable. See Chapter 14 in this Manual, and Kawase (2003).                                                                                                                                                                                                                                                                                                               |
| <b>site response</b>          | The modification of earthquake <i>ground motion amplitude</i> , <i>phase</i> and shape in the <i>time</i> or <i>frequency domain</i> caused by local site conditions. See Chapter 14 in this Manual, and Campbell (2003).                                                                                                                                                                                                                        |
| <b>Skempton's coefficient</b> | A poroelastic constant $B$ relating changes in mean <i>stress</i> $\sigma_m$ to <i>pore fluid pressure</i> $p$ , $B = dp/d\sigma_m$ . See Lockner and Beeler (2002, p. 526).                                                                                                                                                                                                                                                                     |
| <b>skewness</b>               | A quantitative measure of deviation of a random value distribution from the “normal” symmetric <i>Gaussian distribution</i> . The skewness is zero for symmetric distributions, it becomes negative (positive) if the distribution contains outliers to the left (right). See <i>higher order statistics</i> .                                                                                                                                   |
| <b>SKS</b>                    | <i>Mantle shear wave</i> transmitted through the outer <i>core</i> (K) as a <i>P wave</i> . See Figs. 2.8 and 2.52 in Chapter 2 and Figs. 11.48, 11.68, and 11.69 in Chapter 11 of this Manual, as well as Kulhanek (2002, p. 341-343), Cara (2002), and Aki and Richards (2002 or 2009, Box 5.9, p. 181-182). For this and other such symbols used in seismic <i>phase names</i> see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). |
| <b>slab</b>                   | Subducted oceanic <i>lithosphere plate</i> that underthrusts the continental plate in a <i>subduction zone</i> and is consumed by the Earth's <i>mantle</i> . Typically a dipping tabular region of relatively high seismic <i>velocity</i> , high $Q$ and low temperature. Deep earthquakes only occur within subducted slabs, and such <i>seismicity</i> extends to depths down to 700 km. See Lay (2002).                                     |
| <b>slab pull</b>              | The force of gravity causing the cooler and denser oceanic <i>slab</i> to sink into the hotter and less dense <i>mantle</i> material. The downdip component of this force leads to downdip extensional <i>stress</i> in the slab and may produce earthquakes within the subducted slab. Slab pull may also contribute to stress on the <i>subduction thrust fault</i> if the <i>fault</i> is <i>locked</i> .                                     |
| <b>slickensides</b>           | A polished and smoothly striated surface that results from <i>slip</i> along a <i>fault</i> surface. The striations themselves are slickenlines.                                                                                                                                                                                                                                                                                                 |
| <b>slip</b>                   | The relative <i>displacement</i> of formerly adjacent points on opposite sides of a <i>fault</i> . See <i>slip vector</i> , <i>fault slip</i> .                                                                                                                                                                                                                                                                                                  |
| <b>slip model</b>             | A kinematic model that describes the amount, distribution, and timing of a <i>slip</i> associated with an earthquake.                                                                                                                                                                                                                                                                                                                            |
| <b>slip moment</b>            | A geological term for <i>seismic moment</i> . See Chapter 36 in Lee et al. (2002).                                                                                                                                                                                                                                                                                                                                                               |

|                                                |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|------------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>slip rate</b>                               | How fast the two sides of a <i>fault</i> are slipping relative to one another, as derived from seismic records in case of an earthquake or determined, as a long-term average, from geodetic measurements, from offset man-made structures, or from offset geologic features whose age can be estimated. It is measured parallel to the predominant <i>slip</i> direction or estimated from the vertical or horizontal offset of geologic markers. See also <i>fault slip rate</i> .                                                                                                                                                                                                                                                                                 |
| <b>slip vector</b>                             | The direction and amount of <i>slip</i> in a <i>fault plane</i> , showing the relative motion between the two sides of the <i>fault</i> . See Figs. 3.30 and 3.31 in Chapter 3 and IS 3.1 in this Manual, Fig (13) in Ben-Zion (2003), Jackson (2002), and Aki and Richards (2002 or 2009, Fig. 4.13, p. 101).                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>slip waves</b>                              | See <i>creep waves</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>slope-intercept method</b>                  | A graphical technique used to infer seismic <i>velocities</i> and layer thicknesses from seismic <i>travel-time</i> data, based on the assumption that the <i>crust</i> is built up of uniform layers bounded by horizontal <i>discontinuities</i> . See Fig. 2.36 and section 2.5.3.5 in this Manual, and Minshull (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>slow earthquake</b>                         | <i>Fault</i> failure occurring without significant seismic radiation due to very low <i>rupture velocity</i> and/or <i>stress drop</i> . See Johnston and Linde (2002). See <i>creep</i> , <i>creep event</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>slowness</b>                                | The inverse of <i>velocity</i> $v$ , often annotated as scalar $s$ , related to vector $\mathbf{k}$ (the <i>wavenumber vector</i> ) and given in units seconds/degree or seconds/km.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>slowness vector <math>\mathbf{s}</math></b> | The slowness vector $\mathbf{s}$ points into the direction of wave propagation and has the absolute value $s = 1/v$ with $v$ = wave propagation <i>velocity</i> . See also <i>wavenumber vector</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>slow-slip event</b>                         | <i>Aseismic slip</i> (no <i>seismic waves</i> are generated) on the <i>subduction interface</i> downdip from the seismogenic zone. Slow slip-events are detected by measuring crustal <i>deformation</i> with dense GPS or <i>tiltmeter</i> networks. Estimates of the time span for these events range from 6 to 15 days in the Cascadia subduction zone (Dragert et al. 2001). Recurrence rate for these events ranges from 13 to 16 months (Cascadia) to approximately six month in southwest Japan (Obara et al., 2004). Slow-slip events appear to coincide with the occurrence of major non-volcanic tremor activities. See also <i>creep</i> , <i>creep event</i> , <i>silent earthquake</i> , <i>non-volcanic tremor</i> , <i>episodic tremor and slip</i> . |
| <b>small earthquake</b>                        | An earthquake with magnitude that ranges from 3 to 5 is often called “small”. See <i>magnitude</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |

|                                     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|-------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Sn</b>                           | Early use of the designation Sn was in reference to short- <i>period</i> S waves that were presumed to propagate as <i>head waves</i> along the top of the <i>mantle</i> . The new <i>IASPEI</i> standard <i>phase names</i> define Sn as any S wave bottoming in the uppermost mantle or an upgoing S wave from a source in the uppermost mantle; see also the discussion about the analogues defined phase <i>Pn</i> . Quite commonly, the term is now also applied to a prominent <i>arrival</i> of short- <i>period shear waves</i> that may be observed (with a straight-line <i>travel-time</i> curve) at <i>epicentral distances</i> as great as 40°. Although Sn arrives at regional distances larger than 100 km usually as the first S-wave arrival, its <i>amplitudes</i> are, as compared to the later <i>Sg</i> , <i>SmS</i> , <i>Lg</i> arrivals, relatively small and often covered in the signal generated noise after <i>Pg</i> . For record examples see, e.g., Figs. 1a, 3a, 4a, 6a, and 6c in DS 11.1 of this Manual and Kulháněk (2002). For this and other such symbols used in seismic phase names see IS 2.1 in this Manual and Storchak et al. (2003 and 2011). |
| <b>Snell's law</b>                  | In seismology, the law describing the angles of <i>seismic rays</i> reflected and transmitted (i.e., refracted) at an <i>interface</i> in the Earth. It is derived from the conservation of wave <i>slowness</i> parallel to the interface. This law was originally discovered in optics by the Dutch physicist Willebrord Snell in about 1621. See section 2.5.2 in Chapter 2 of this Manual, Chapman (2002), and Bleistein (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>SOFAR channel</b>                | A depth region of low acoustic <i>velocity</i> within the ocean water column which allows sound waves to propagate over very large distances with very low <i>attenuation</i> . It is a very efficient waveguide for the seismic <i>T phase</i> from earthquake sources. See Kulhanek (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>soil</b>                         | (1) In geotechnical engineering, all unconsolidable material above the <i>bedrock</i> . (2) In soil science, naturally occurring layers of mineral and (or) organic constituents that differ from the underlying parent material in their physical, chemical, mineralogical, and morphological character because of <i>pedogenic</i> processes.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>soil profile</b>                 | Vertical arrangement of soil horizons down to the parent material or to <i>bedrock</i> . Commonly subdivided into A, B and C horizons.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>soil-structure interaction</b>   | A term applied to the consequences of the <i>deformation</i> and forces induced into the soil by the movement of a structure. The common fixed-base assumption in the analysis of structures implies no soil-structure interaction.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>solitons (or solitary waves)</b> | A special kind of localized waves that propagate undistorted in shape. They are essentially nonlinear waves, which can be treated as non-dispersive localized packets of energy moving                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |

|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
|----------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                  | with uniform <i>velocity</i> . Solitons are exact solutions of nonlinear wave equations. See Majewski (2006).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>sound waves</b>               | See <i>air wave</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>source</b>                    | See <i>seismic source</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>source depth</b>              | Depth of an earthquake <i>hypocenter</i> (see <i>focal depth</i> ) or of a buried explosion, mining collapse, or any other type of <i>source</i> that generates <i>seismic waves</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>source directivity effect</b> | The effect of the earthquake source on the <i>amplitude</i> , <i>frequency</i> content and duration of the <i>seismic waves</i> propagating in different directions, resulting from the propagation of <i>fault rupture</i> in one or more directions with finite speeds. Directivity effects are comparable to, but not exactly the same as the <i>Doppler effect</i> for a moving <i>oscillator</i> . See Faccioli and Pessina (2003), and material below Eq. (3.8), Figure 6 in Ben-Zion (2003) and Fig. 3.112 in Chapter 3 of this Manual                                                                                                                                                                                                                                                                  |
| <b>source duration</b>           | Time length of <i>seismic energy</i> radiation from the <i>seismic source</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>source effect</b>             | The effect of the earthquake source on seismic motions. It is implicitly assumed that the source, <i>path</i> and <i>site effects</i> on <i>ground motions</i> are separable. See, e.g., Aki (2002), and Eq. (3.5) in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>source function</b>           | The <i>ground motion</i> generated at the <i>fault</i> during rupture, usually as predicted by a theoretical model and represented by a time history or <i>spectrum</i> . The terms Brune spectrum, Aki spectrum, and <i>Haskell model</i> refer to varying representations of the source function, each based on different assumptions, as devised by the scientist for which the model is named. Strictly defined, the source function is a compact space-time history of the earthquake source process that will give the observed <i>displacement waveforms</i> as its convolution with the <i>Green function</i> for the wave propagation in the Earth and the <i>instrument response</i> of the recording instrument. See Aki and Richards (1980), Ben-Menahem and Singh (1981), Das and Kostrov (1988). |
| <b>source-to-site distance</b>   | Shortest distance between an observation point and the source of an earthquake <i>hypocenter</i> that is represented as either a point (point-source distance measure) or a ruptured area (finite-source distance measure). See Campbell (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>source zone</b>               | An area in which an earthquake is expected to originate. More specifically, an area considered to have a uniform rate of <i>seismicity</i> or a single probability distribution for purposes of a <i>seismic hazard</i> or <i>seismic risk</i> analysis.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |



|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|----------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>space-based geodesy</b>       | Space-based technologies to measure the positions of geodetic monuments on the Earth. A fundamental departure from the traditional geodesy is its reference to an inertial reference frame rather than those fixed to the Earth. Current accuracy of better than a centimeter for sites thousands of kilometers apart is good enough for measuring the present-day <i>plate motion</i> . See Stein and Klosko (2002), and Feigl (2002).                                                                                                                                                                                                                                       |
| <b>spasmodic burst</b>           | A rapid-fire sequence of high- <i>frequency</i> earthquakes often observed in volcanic or geothermal areas presumably due to a cascading <i>brittle-failure</i> sequence along subjacent <i>faults</i> in a fracture mesh driven by transient increases in local fluid <i>pressures</i> . See Hill et al. (1990).                                                                                                                                                                                                                                                                                                                                                             |
| <b>specific barrier model</b>    | A <i>fault</i> model consisting of a <i>rectangular fault plane</i> filled with circular-crack sub-events developed by Papageorgiou and Aki (1983) for interpreting the observed power spectra of strong <i>ground motion</i> records. It is a hybrid of deterministic and stochastic fault model, in which the sub-event ruptures statistically independently from each other as the <i>rupture front</i> sweeps along the <i>fault length</i> . See Papageorgiou (2003) for its latest review.                                                                                                                                                                              |
| <b>spectral acceleration</b>     | Commonly refers to either the <i>Fourier amplitude spectrum</i> of ground <i>acceleration</i> or the pseudo relative velocity response spectrum (PSRV); abbreviated as SA.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>spectral amplification</b>    | A measure of the relative shaking response of different geologic materials depending on the <i>frequency</i> of excitation; the ratio of the <i>Fourier amplitude spectrum</i> of a <i>seismogram</i> recorded on one material to that computed from a seismogram recorded on another material for the same <i>earthquake</i> or explosion.                                                                                                                                                                                                                                                                                                                                   |
| <b>spectral ratio</b>            | The ratio of two <i>amplitude</i> (or <i>power</i> ) <i>spectra</i> . In classical linear theory, the ratio between transverse <i>acceleration</i> and <i>rotation rate</i> for time windows containing <i>surface waves</i> is proportional to the <i>dispersion</i> curves, the <i>frequency-dependent</i> local <i>phase velocities</i> (see Lee and Trifunac 2009). The spectral ratio H/V of horizontal to vertical component records of ambient <i>seismic noise</i> (so-called <i>Nakamura method</i> ) is used to estimate the fundamental <i>resonance</i> frequency of the sedimentary cover layers in <i>microzonation</i> studies (see Chapter 14 of the Manual). |
| <b>spectrum</b>                  | Curves showing <i>amplitude</i> or <i>phase</i> of a <i>time-history</i> as a function of <i>frequency</i> or <i>period</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>spectrum intensity (S.I.)</b> | A broadband measure of the intensity of strong <i>ground motion</i> defined by the area under a <i>response spectrum</i> between two selected <i>natural periods</i> or <i>frequencies</i> . First introduced by                                                                                                                                                                                                                                                                                                                                                                                                                                                              |

Housner (1952) as the area under the 20 per cent damped response spectrum between periods of 0.1 and 2.5 seconds.

**spherical surface harmonics** Analog of the two-dimensional Fourier series for the spherical surface. It combines Fourier harmonics and Legendre polynomials and is used to describe the Earth's *normal modes* of oscillation (among many other phenomena). See Aki and Richards (2002 or 2009, Box 8.1, p. 334-339).

**spherical wave** A wave for which the constant-phase surfaces are spheres perpendicular to the direction of propagation. See *plane wave*; also Aki and Richards (2002 or 2009, p. 190-217).

**spheroidal** A mode of vector field **u** in spherical coordinates for which the radial component of **rot u** is zero. See *toroidal*; also Aki and Richards (2002, p. 341).

**spinor** A geometrical object similar to a tensor that describes a rotational state. Some authors believe that spinors are more fundamental objects than tensors. See Majewski (2008).

**spread** The layout of *seismometer/geophone* groups from which data from artificial *seismic sources* (such as explosions, vibrators, *airguns*) are recorded simultaneously.

**spread footing** A support structure for a wall, column or pier that is essentially a horizontal mat, usually of reinforced concrete, lacking piles or caissons. The footing works by spreading the load over an area wide enough to reduce the bearing *stresses* to permissible levels. See Park and Paulay (1975).

**SRSS (Square Root of the Sum of Squares)** A technique used in earthquake engineering analysis and design (introduced by Rosenblueth, 1951), wherein the maximum contributions of different modes to a variable of interest, determined from a *response spectrum* or spectra, are combined by squaring them, adding, and taking the square root of the sum. The result gives an estimate of how the modal responses, whose time of maximum occurrence is not known, combine to produce the maximum total response. See also the biography of Emilio Rosenblueth in Howell (2003).

**SSAM** An acronym for real-time Seismic Spectral Amplitude Measurement; the SSAM system aids in the rapid identification of precursory *long-period (LP) volcanic seismicity* during rapidly escalating *volcanic unrest* (see Stephens et al., 1994, for details).

**stable tectonic regime** A term that refers to a region where *tectonic deformation* is relatively low and earthquakes are relatively infrequent, usually far from plate *boundaries*.

|                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|----------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>stacking process</b>          | In <i>reflection seismology</i> , the process of summing several seismic traces that have been corrected for normal <i>moveout</i> . See Sheriff and Geldart (1995). See <i>beamforming</i> and <i>delay-and-sum processing</i> .                                                                                                                                                                                                                                                                                                      |
| <b>standard deviation</b>        | A measure of how much a set of data is different from the curve it should make when plotted on a graph. Or, the square root of the average of the squares of deviations about the mean of a set of data. Standard deviation is a statistical measure of spread or variability.                                                                                                                                                                                                                                                         |
| <b>state of tectonic stress</b>  | The magnitude and orientation of <i>principal stresses</i> in the Earth. Principal stresses are usually aligned with horizontal and vertical. See Zoback and Zoback (2002).                                                                                                                                                                                                                                                                                                                                                            |
| <b>static fatigue</b>            | A laboratory observed phenomenon in mechanical engineering, material science and rock mechanics, consisting of an inherent time delay between the application of a <i>stress</i> increase and the occurrence of <i>brittle</i> failure ( <i>stress drop</i> ) induced by the stress change. See Lockner and Beeler (2002, p. 523).                                                                                                                                                                                                     |
| <b>static stress drop</b>        | A theoretically well-defined quantity that compares the “before” and “after” <i>stress-strain</i> states of the elastic volume that surrounds the <i>fault</i> . The definition of the overall static stress drop reduces to a weighted integral over just the fault portion that slips during the earthquake. See <i>dynamic stress drop</i> ; also Figure 10 in IS 3.1 of this Manual, Ruff (2002, p. 548-550), Brune and Thatcher (2002, p. 580-581), and Figure 14 in Ben-Zion (2003).                                             |
| <b>station</b>                   | In seismology, the site where geophysical instruments, e.g., <i>seismographs</i> , have been installed for observations. Stations can either be single sites or specially grouped <i>arrays</i> .                                                                                                                                                                                                                                                                                                                                      |
| <b>stationary random process</b> | An ensemble of time series for which the ensemble-averaged statistical properties such as mean and variance do not vary with time. If the statistical averages taken over time for a single member of the ensemble are the same as the corresponding averages taken across the ensemble, the stationary random process is also ergodic.                                                                                                                                                                                                |
| <b>stationary linear system</b>  | See <i>linear system</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>station calibration</b>       | For using <i>station</i> observations from all over the world in a common data center, these station observations should be corrected (if possible) for all very anomalous local effects such as significant azimuth- and distance-dependent <i>travel-time residuals</i> . The same applies to anomalously large or small <i>amplitudes</i> of relevant <i>seismic phases</i> due to strong <i>site effects</i> which systematically bias station <i>magnitude</i> readings with respect to the network average. See Richards (2002). |

|                               |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|-------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>steady-state strength</b>  | The shear resistance of contractive soil in the liquefied state. With respect to <i>liquefaction</i> , the steady-state strength and the residual strength are used interchangeably. See Youd (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>steam-blast</b>            | See <i>phreatic</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>step-out</b>               | The time between two seismic <i>phases</i> , such as pP and P, at a specific <i>epicentral distance</i> from a <i>station</i> . The step-out, if it increases or decreases as the distance increases, can be a characteristic determinant for phase identification.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>stepover</b>               | Region where one <i>fault</i> ends, and another <i>en echelon</i> fault of the same orientation begins; described as being either right or left, depending on whether the bend or step is to the right or left as one progresses along the fault.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>stick-slip</b>             | A periodic or quasi-periodic faulting cycle consisting of a relatively long period of static, predominately elastic loading followed by rapid <i>fault slip</i> seen on laboratory fault surfaces loaded at a constant rate of shear stressing; stick-slip is considered the laboratory time scale equivalent of periodic earthquake recurrence. See Lockner and Beeler (2002, p. 518).                                                                                                                                                                                                                                                                                                                                                                       |
| <b>stochastic process</b>     | A physical process with some random or statistical element in its structure. See Turcotte and Malamud (2002), and Vere-Jones and Ogata (2003, p. 1577).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>Stoneley wave</b>          | Wave trapped at a plane <i>interface</i> of two elastic media. It is always possible at a solid-fluid interface with the <i>phase velocity</i> lower than the compressional velocity of the fluid. It can exist at a solid-solid interface only in restrictive cases. See Aki and Richards (2002, p. 156-157), as well as Webb (2002), and the biography of Robert Stoneley in Howell (2003).                                                                                                                                                                                                                                                                                                                                                                 |
| <b>strain (strain tensor)</b> | Change in shape and/or size of a infinitesimal material element per unit length or volume. In elasticity theory, the strain is generally referred to a reference state in which the material has zero <i>stresses</i> (natural configuration). For small <i>deformation</i> , the complete description of strain at a point in three dimensions requires the specification of the extensions (change in length/length) in the directions of each of the reference axes and the change in the angle between lines which were parallel to each pair of reference axes in the reference state. Diagonal components of the strain tensor describe normalized changes in length; off-diagonal components describe changes in angle. See Johnston and Linde (2002). |
| <b>strain events</b>          | Episodic <i>strain</i> transients recorded by <i>strainmeters</i> . See Johnston and Linde (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |

|                           |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
|---------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>strain hardening</b>   | The property of a material to resist post-yield <i>strains</i> with increasing <i>stress</i> , as opposed to the constant post-yield stress of a perfectly plastic material.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>strainmeter</b>        | Strainmeters or extensometers are another basic type of <i>seismic sensor</i> . They measure the motion of one point of the ground relative to another, in contrast to inertial <i>seismometers</i> which measure the <i>ground motion</i> relative to an inertial reference such as a suspended mass. See IS 5.1 in this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>strain rate</b>        | <i>Strain</i> measurements are computed from observed changes in length on the Earth's surface, commonly along multiple paths. Because the changes in length are observed over varying time periods and path lengths, they are expressed as the change in length divided by the measurement distance divided by the time period of measurement. This number, termed the strain rate, is expressed as the change in length per unit length per unit time. Strain rates vary widely in different <i>tectonic</i> environments, ranging from about $10^{-8}$ /year at highly active <i>plate boundaries</i> and $10^{-11}$ to $10^{-12}$ within stable continental platforms. These differences in strain rates control the average return period of earthquakes.                                                                                                                                                              |
| <b>strain steps</b>       | Step-like changes in <i>strain</i> recorded by <i>strainmeters</i> . See Johnston and Linde (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>stratigraphy</b>       | The study of the character, form and sequence of layered rocks.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             |
| <b>stratovolcano</b>      | Volcanic landform produced by deposits from explosive and non-explosive <i>eruptions</i> , resulting in a volcanic edifice built of interbedded layers of <i>lavas</i> and <i>pyroclastic</i> materials.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>strength (of rock)</b> | The ability of rock to withstand an applied <i>stress</i> before failure. It is measured in units of stress by means of the uniaxial and triaxial compression test. See Handin (1966), and Scholz (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>stress</b>             | Force, resolved into normal and tangential components, exerted per unit area on an infinitesimal surface. A complete description of stress (i.e., the <i>stress tensor</i> ) at a point in a three dimensional reference frame requires the specification of the three components of stress on each of three surface elements orthogonal to one of the reference axes. Hence there are nine components of stress. However, in a <i>homogeneous, isotropic</i> medium (i.e., classical continua, consisting of point masses), the balance of <i>torque</i> implies the symmetry of the stress tensor and therefore reduces the stress description to six independent components: three normal components and three shear components. In <i>micropolar</i> theory, stresses may be non-symmetric. In contrast, <i>traction</i> is a force vector defined for a particular surface. See IS 3.1 in this Manual, as well as Ruff |

|                             |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|-----------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                             | (2002), Zoback and Zoback (2002), Johnston and Linde (2002), and Figure 1 in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>stress corrosion</b>     | A <i>stress</i> degradation process that is related to the action of some chemical agents. See Teisseyre and Majewski (2002).                                                                                                                                                                                                                                                                                                                                                        |
| <b>stress drop</b>          | The difference between the <i>stress</i> on a <i>fault</i> before and after an earthquake. A parameter in many earthquake source models that has a bearing on the level of high- <i>frequency</i> shaking radiated by the earthquake. Commonly stated in units termed bars or megapascals (1 megapascal (MPa) = $10^6$ N/m <sup>2</sup> = 10 bars). See <i>static stress drop</i> , <i>dynamic stress drop</i> , Fig. 10 in IS 3.1 of this Manual, and Figure 14 in Ben-Zion (2003). |
| <b>stress map</b>           | Map showing the orientation and relative magnitude of horizontal <i>principal stress</i> orientations. See Zoback and Zoback (2002).                                                                                                                                                                                                                                                                                                                                                 |
| <b>stress shadow</b>        | A region where earthquakes are temporarily inhibited because the ambient <i>stress</i> has been decreased by a nearby earthquake. A <i>fault</i> can remain in a stress shadow until stress accumulation recovers the pre-earthquake stress state. See Harris (2003).                                                                                                                                                                                                                |
| <b>stress tensor</b>        | See IS 3.1 in this Manual and <i>stress</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>stretch modulus</b>      | Also termed <i>Young's modulus</i> , describes the behavior of a cylinder of length L that is pulled on both ends.                                                                                                                                                                                                                                                                                                                                                                   |
| <b>strike</b>               | Trend or bearing, relative to North, of the line defined by the intersection of a planar geologic surface (for example, a <i>fault</i> or a bed) and a horizontal surface such as the ground. See <i>fault strike</i> , Fig. 3.107 in Chapter 3 of this Manual, and Figure 13 in Ben-Zion (2003).                                                                                                                                                                                    |
| <b>strike-slip</b>          | See <i>fault movement</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
| <b>strike-slip fault</b>    | A <i>fault</i> on which the movement is parallel to the <i>strike</i> of the fault, e.g., the San Andreas fault. The maximum and minimum <i>principal stresses</i> are both horizontal. See Yeats et al., (1997, p.167-248).                                                                                                                                                                                                                                                         |
| <b>Strombolian eruption</b> | Slightly more violent than <i>Hawaiian eruptions</i> , this type of activity is characterized by the intermittent explosion or fountaining of <i>lava</i> from a single <i>vent</i> or <i>crater</i> . Such eruptions typically occur every few minutes or so, sometimes rhythmically and sometimes irregularly.                                                                                                                                                                     |
| <b>strong ground motion</b> | <i>Ground motion</i> of sufficient <i>amplitude</i> and duration to be potentially damaging to a building's structural components, architectural features, or to its content. One common practical designation of strong ground motion is a <i>peak ground</i>                                                                                                                                                                                                                       |

*acceleration* of 0.05 *g* or larger. See Anderson (2003), Bolt and Abrahamson (2003), and Campbell (2003).

**strong motion** See *strong ground motion*.

**strong-motion accelerograph** An *accelerograph* designed to record accurately the *strong ground motion* generated by an *earthquake*. They were originally developed by earthquake engineers, because the traditional *seismographs* designed by seismologists to record weak ground motions from earthquakes were too fragile and did not have sufficient *dynamic range* to record strong ground motions. A typical strong-motion accelerograph has a tri-axial *accelerometer*, which records *acceleration* up to 2 *g* on scale. See Anderson (2003), and Kinoshita (2003).

**strong-motion instrument** See *strong-motion accelerograph*.

**strong-motion parameter** A parameter characterizing the *amplitude* of *strong ground motion* in the *time domain* (time-domain parameter), or *frequency domain* (frequency-domain parameter). See Campbell (2003).

**strong-motion seismograph** See *strong-motion accelerograph*.

**structural geology** The study of geologic structures and their formation processes. Applied to earthquakes, it includes the relation between *faults*, folds and *deformation* of rocks on all scales. See Jackson (2002).

**subduction** A *plate tectonics* term for the process whereby the oceanic *lithosphere* collides with and descends beneath the continental lithosphere.

**subduction thrust fault** The *fault* that accommodates the differential motion between the downgoing oceanic crustal *plate* and the continental plate. This fault is the contact between the top of the oceanic plate and the bottom of the newly formed continental *accretionary wedge*. Also alternately referred to as the *plate-boundary* thrust fault, the thrust *interface* fault or the megathrust fault.

**subduction zone** A zone of convergence of two *lithospheric plates* characterized by thrusting of one plate into the Earth's *mantle* beneath the other. Processes within the subduction zone bring about melt generation in the mantle wedge and cause buildup of the overlying *volcanic arc*. Subduction zones (where most of the world's greatest earthquakes occurred) are recognized from the systematic distribution of *hypocenters* of deep earthquakes called *Wadati-Benioff zones*. See Uyeda (2002, p. 62-64), and biographies of Hugo Benioff and Kiyoo Wadati in Howell (2003).

- subductology, comparative** Comparative study of *subduction zones* emphasizing the differences between the Mariana-type and Chilean-type subduction zones, caused by difference in the degree of *plate coupling*. See Uyeda (1982) and Uyeda (2002).
- subplinian eruption** A small-scale *Plinian eruption* - intermediate in size and energy release between Strombolian and Plinian activity - that is characterized by *pumice* and *ash* deposits covering less than 500 km<sup>2</sup>.
- substructure** Generally, that portion of a structure that lies below the ground, for example, the basement parking in a tall building. In analyses, the substructure can include a portion of the surrounding ground as well as all or part of the structural foundation. See Hall (2003).
- summit (or central) eruptions** General terms referring to *eruptions* that take place from one or more *vents* (fissural or cylindrical) located at or near the summit, or center, of the volcanic edifice, in contrast to eruptive outbreaks on the flanks of volcanoes downslope from the summit area. Compare with *flank* or *lateral eruption*.
- superstructure** That portion of a structure that is above the surface of the ground or other reference elevation at which a significant change in the structure occurs. See Hall (2003).
- surface faulting** Faulting that reaches the Earth's surface; commonly accompanies moderate and large earthquakes having *focal depths* less than 20 km. The faulting is almost always *coseismic*, however it may also accompany *aseismic tectonic creep* or natural or man-induced subsidence. See Bonilla (1970).
- surface P wave** The *ray path* of surface P consists of two segments: a) an *S-wave* path from the source to the free surface with an apparent horizontal *velocity* equal to the *P-wave* velocity, and a *P-wave* path along the free surface to the receiver. Surface P-waves appear at the critical distance and can be a sharper *arrival* than the direct S waves, although they attenuate very rapidly with distance. In some respects they behave like *diffracted* or *head waves*.
- surface reflections/conversions** Waves leaving the source downwards and then undergoing one or more *reflections* or *conversions* (from P to S or vice versa) at the surface before arriving at the recording station, contrary to *depth phases*, which leave the source in an upward direction. Examples are *PP*, *PPP*, *SS*, *SSS*, *SP*, *PS*, *PPS*, *SP* etc. For these and other such symbols used in seismic *phase names* see IS 2.1 in this Manual and Storchak et al. (2003 and 2011).



- surface wave** *Seismic wave* that propagates along the surface of an elastic *halfspace* or a layered elastic halfspace. See *Love wave* and *Rayleigh wave*, Chapter 2 of this Manual and also Romanowicz (2002), and Kulhanek (2002).
- surface-wave magnitude ( $M_S$ )** An earthquake *magnitude* scale determined from surface wave records and introduced by Gutenberg (1945) uses maximum *amplitudes of surface waves* with *periods* around 20 seconds. In its IASPEI-recommended standard form, it is now termed  $M_{s_{20}}$  and uses a slightly different *magnitude calibration function* than Gutenberg and amplitude readings on seismic records which simulate the long-period WWSSN *response*. Another IASPEI-recommended surface wave magnitude standard is the broadband version  $M_{s_{BB}}$ . It allows magnitude determinations in a wider range of periods (2 to 60 s) and distances ( $2^\circ$  to  $160^\circ$ ), is better tuned to the use of the IASPEI  $M_s$  standard *calibration function* and correlates better with the *moment magnitude*  $M_w$ . See *magnitude saturation*, Chapters 3 and IS 3.3 in this Manual, Bormann et al. (2009), and IASPEI (2013).
- SV** See *SH*, *SV*.
- swarm (of earthquakes)** A sequence of earthquakes occurring closely clustered in space and time with no dominant *mainshock*. See McNutt (2002) and Utsu (2002b).
- swath bathymetry** Seafloor bathymetry obtained by a multi-narrow beam echo sounder. One sounding sequence can provide bathymetry information over a width of more than 7 times the water depth. See Suyehiro and Mochizuki (2002).
- S wave** *Elastic waves* producing shear and no volume change are called S waves (S standing for Latin *Secundae* or “secondary”). In a *homogeneous, isotropic*, linear elastic medium the *velocity* of S waves is equal to  $(\mu/\rho)^{-1/2}$ , where  $\mu$  and  $\rho$  are the *rigidity* and *density*, respectively. For a classical isotropic linear elastic medium (in an isotropic *far field*), the particle *displacement* associated with S waves is perpendicular to the direction of wave propagation, so they are sometimes called “transverse waves.” See Aki and Richards (2002, p. 122), and Ben-Zion (2003, Eq. (1.9b)). In an isotropic full space, only S waves generate *rotational motions*. For isotropic *Cosserat continua*, the S wave becomes a mixed shear-rotational wave. In anisotropic media, it couples also with *P waves*.
- synthetic (ground motion)** Time-history of a strong *ground motion* for engineering purposes calculated by a deterministic or stochastic simulation. See Boore (2003) for the latest review on simulation of ground motion using the stochastic model. See also Faccioli and Pessina (2003).

**synthetic seismograms** Computer-generated *seismograms* for models of *seismic source* and Earth's structure. Various methods are used for different purposes, encompassing the whole field of theoretical seismology. Those for the vertically *heterogeneous* one-dimensional Earth model have been well developed and described in many text books. Those for the 2D- and 3D-models are still under development. See Chapters 8 through 13 and 55 in Lee et al. (2002) and for examples Figs. 2.82, 2.83, and 2.85 of Chapter 2 and Figures 3 and 4 in IS 11.4.

## T

**take-off angle** The angle under which a *seismic ray* leaves the *seismic source*, measured from the plum line clockwise or counter-clockwise. It is 0° when taking off in the direction of the plum line, 90° when leaving the source in horizontal direction and 180° when taking off vertically upward. The take-off angles from the source have to be known (respectively calculated on the basis of a certain model) when reconstructing seismic *fault-plane solutions* from observed *first motion polarities*, *amplitude ratios* or *polarization analysis* via stereographic projections of the *focal sphere*. See section 3.4.2 of Chapter 3 and EX 3.1 in this Manual.

**tamped explosion** An explosion in which the explosive is buried under the ground firmly to avoid venting into the atmosphere. See Richards (2002).

**Taylor expansion** A mathematical technique commonly used to linearize an equation. It is based on the Taylor's theorem which describes approximating polynomials for rather general functions and provides estimates for errors. It is named after the English mathematician and philosopher Brook Taylor (1685-1731). See James and James (1976).

**tectonic** Of, pertaining to, or designating the rock structure and external forms resulting from deep-seated crustal and subcrustal forces in the Earth.

**tectonic earthquake** An earthquake caused by the release of *strain* that has accumulated in the Earth as a result of broad scale *tectonic deformations* (as contrasted with volcanic earthquakes or impact earthquakes).

**tectonic plates** Large, relatively *rigid plates* of the *lithosphere* that move relative to one another on the outer surface of the Earth.

**tectonic rotational solitons** Rotational (spin or twist) *solitons* generated by past earthquakes and propagating slowly (about 1 km/day) along a fractured

earthquake *fault* and triggering new earthquakes. See Majewski (2006).

|                                   |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|-----------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>tectonics</b>                  | The branch of Earth Science that deals with the structure, evolution, relative motion, and <i>deformation</i> of the outer part of the Earth, or the <i>lithosphere</i> , on a regional to global scale, for time scales ranging from very short up to thousands of millions of years. The term “active tectonics” refers to tectonic movements that occur on a time scale of up to thousands of years, and “neotectonics” to tectonic movements on a scale of thousands of years to a few million years. The lithosphere includes the Earth's <i>crust</i> and part of the Earth's upper mantle and averages about 100 km in thickness. Plate tectonics is a theory of global tectonics in which the Earth's lithosphere is divided into a number of spherical <i>plates</i> that undergo dominantly rigid-body translation on the sphere. The relative motions of these plates cause earthquakes and deformation along the plate boundaries and in adjacent regions. |
| <b>tectonoelectric</b>            | Electric fields generated by non-seismic crustal <i>stress</i> changes. See Johnston (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>tectonomagnetic</b>            | Magnetic fields generated by non-seismic crustal <i>stress</i> changes. See Johnston (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>tectonophysics</b>             | A branch of geophysics that deals with the <i>deformation</i> of the Earth.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>teleseism</b>                  | A <i>seismic source</i> at an <i>epicentral distance</i> larger than about 20° or 2200 km from the observation site.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>teleseismic waves</b>          | <i>Seismic waves</i> observed at distances more than 20° (or about 2200 km) away from the <i>epicenter</i> . For record and ray path examples see, e.g., Fig. 1.6 in Chapter 1, section 2.6.2 of Chapter 2, sections 11.5.2 to 11.5.4 of Chapter 11 as well as record examples in DS 11.2 and 11.3 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>tensile (tensional) stress</b> | The <i>stress</i> component normal to a given internal surface, such as a <i>fault plane</i> , that tends to pull the materials apart. See <i>compressive stress</i> , <i>normal stress</i> , Figure 3 in IS 3.1 of this Manual, and Figure 1 in Ben-Zion (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>tensor of inertia</b>          | Rotational analogue of mass. This is a symmetric (maximum six components), positively defined tensor that describes the distribution of mass in a rigid body. This tensor rotates together with the body if it moves, and its eigenvalues are called principal moments of inertia (diagonal components of the symmetric tensor). The tensor of inertia multiplied by the vector of <i>rotational velocity</i> is a part of the <i>kinetic moment</i> . In <i>micropolar</i> theories, inertial characteristics of each particle are density and density of tensor of inertia.                                                                                                                                                                                                                                                                                                                                                                                          |

|                                             |                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|---------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>tensor strainmeters</b>                  | Borehole <i>strainmeters</i> measuring three horizontal <i>strain</i> components from which the principal strain components can be derived. See Johnston and Linde (2002).                                                                                                                                                                                                                                                                               |
| <b>Tertiary</b>                             | A period in geologic time, from 66.4 to 1.6 million years ago. See <i>geologic time</i> .                                                                                                                                                                                                                                                                                                                                                                |
| <b>tephra</b>                               | Collective term for airborne <i>pyroclastic</i> material, regardless of fragment size or shape, produced by explosive volcanic <i>eruptions</i> . Tephra can be of juvenile or lithic origin. See <i>lithics</i> and <i>juvenile</i> .                                                                                                                                                                                                                   |
| <b>theoretical onset time</b>               | Time at which the <i>arrival</i> of a seismic <i>phase</i> is expected to appear on a <i>seismic record</i> , based on the known location and depth of the <i>seismic source</i> and related theoretical <i>travel-time</i> calculations based on an assumed Earth model.                                                                                                                                                                                |
| <b>thermal boundary layer</b>               | Layers at the top and bottom of a convecting fluid where heat advection and heat conduction are in balance.                                                                                                                                                                                                                                                                                                                                              |
| <b>thermal conductivity (of rock)</b>       | The rate of <i>heat flow</i> in rock per unit area of the internal surface normal to the temperature <i>gradient</i> vector is proportional to the magnitude of the temperature gradient. This proportionality constant is the thermal conductivity.                                                                                                                                                                                                     |
| <b>thermal fluid pressurization</b>         | Transient thermal pressurization of <i>fault zone</i> fluids as a consequence of heat generated during <i>seismic slip</i> . See Sibson (2002).                                                                                                                                                                                                                                                                                                          |
| <b>three-dimensional site amplification</b> | Traditionally, site <i>amplification</i> of <i>ground motion</i> has been modeled by means of a depth-dependent one-dimensional structure. Three-dimensional amplification considers a structure varying in all directions due to features such as irregular topography and sub-surface <i>interfaces</i> . The computational load increases dramatically as one proceeds from 1-D to 3-D through the intermediate 2-D approximation. See Kawase (2003). |
| <b>thrust fault</b>                         | A <i>reverse fault</i> that dips at shallow angles, typically less than about 30 degrees. See also <i>fault</i> and Jackson (2002).                                                                                                                                                                                                                                                                                                                      |
| <b>tilt</b>                                 | A term that to some authors seems to mean long-period <i>rotations</i> about a horizontal axis, to others only static rotations, and to yet others rotations at any <i>frequency</i> . The term “tilt” should be used only with full, graphically supported definitions inclusive of frequency band, according to Evans et al. (2010).                                                                                                                   |
| <b>tiltmeter</b>                            | An instrument designed to measure changes in the direction of (or angle to) the local surface normal. <i>Tilt</i> at the Earth’s surface is equivalent to the horizontal components of <i>rotation</i> .                                                                                                                                                                                                                                                 |

|                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                      |
|---------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                 | Mechanically based tiltmeters are in general sensitive to horizontal <i>acceleration</i> , according to the separation distance of axis and center of mass. At very long <i>periods</i> ( <i>Earth's eigenmodes</i> , <i>Earth tides</i> ) the horizontal components of <i>ground motion</i> recorded by <i>inertial sensors</i> may contain substantial contributions due to tilts (see section 5.3.3 in Chapter 5 of this Manual). |
| <b>tilt noise</b>               | A <i>noise</i> on a horizontal-component <i>seismogram</i> created by <i>rotation</i> (local <i>tilt</i> ) around a horizontal axis. Slight tilts produce large apparent <i>accelerations</i> as a component of the acceleration of <i>gravity</i> is coupled into the horizontal components. See section 5.3.3 with Figs. 5.8 and 5.9 in Chapter 5 of this Manual, and Webb (2002).                                                 |
| <b>time domain</b>              | A seismic record is usually presented in the time domain, i.e., as a display of varying <i>amplitudes</i> of <i>ground motion</i> as a function of time (in contrast to the equivalent representation in the <i>frequency domain</i> ). See also <i>Fourier analysis</i> .                                                                                                                                                           |
| <b>time history</b>             | The sequence of values of any time-varying quantity (such as a <i>ground motion</i> measurement) measured at a set of fixed times. Also termed time series. Examples are <i>accelerograms</i> , <i>seismograms</i> or recordings of the <i>displacement</i> of a point in a structure.                                                                                                                                               |
| <b>tomogram</b>                 | An image of (usually a slice through) the interior of a body (in this case, the Earth) formed using <i>tomography</i> . See Curtis and Snieder (2002).                                                                                                                                                                                                                                                                               |
| <b>tomography</b>               | The science and technological art of creating images of the interior of a body. See <i>seismic tomography</i> for imaging the Earth, and Curtis and Snieder (2002).                                                                                                                                                                                                                                                                  |
| <b>topographic site effects</b> | <i>Site effects</i> caused by surface irregularities such as canyons or mountains. In practice, it is difficult to separate topographic effects from effects caused by subsurface layering. See Kawase (2003).                                                                                                                                                                                                                       |
| <b>tornillo</b>                 | In <i>volcano seismology</i> , a term sometimes applied to <i>long-period (LP) volcanic earthquakes</i> with an especially monochromatic <i>coda</i> reminiscent of the profile of a screw (from the Spanish for “screw”). See McNutt (2002, Figure 4).                                                                                                                                                                              |
| <b>toroidal</b>                 | A mode of vector field $\mathbf{u}$ in spherical coordinates for which both the radial component of $\mathbf{u}$ and $\text{div } \mathbf{u}$ vanish. See <i>spheroidal</i> , and Aki and Richards (2002 or 2009, p. 341).                                                                                                                                                                                                           |
| <b>torque</b>                   | Torque is a cause of the <i>rotational motion</i> of a body (i.e., rotational analogue of force). Torque is equal to the <i>moment of force</i> plus proper torque, which does not depend upon forces.                                                                                                                                                                                                                               |

Independent (“proper”) torques were introduced for the first time by Leonhard Euler in the theory of rods. In *micropolar* theories, each elementary material volume is subjected to forces as well as to independent torques (which may be caused, for instance, by moment of forces on the next micro level, or may have another origin).

**torque waves**

Knopoff and Chen (2009) showed that frictional torques accumulated in a *fault zone* of finite width are relaxed through the development of *torque* or *rotation* waves radiated as *shear waves* during the fracture process near the tips of advancing cracks.

**torsion**

A term that seems to mean *rotations* or *strains* about the vertical axis of a structure (but with the same variance as “*tilt*” in the implied *frequency* band). The term “*torsion*” should be used only with full, graphically supported definitions inclusive of frequency band, according to Evans et al. (2010).

**torsion balance**

Also called a torsion pendulum, an instrument used to measure small forces. Invented by John Michell, it was used by Coulomb to establish the electrostatic force law and by Henry Cavendish to measure the Earth’s *density*. The Cavendish balance is also used to measure the universal gravitational (fundamental) constant, *G*. The restoring *torque* that acts on the boom of the balance depends upon the *shear modulus* of the fiber that supports it. An instrument similar to the classic torsion balance is the Wood-Anderson *seismometer*.

**total overburden stress**

The vertical *stress* in a soil layer due to the weight of overlying soil, water, and surface loads. See Youd (2003).

**T phase**

Tertiary wave (*T wave*) in seismology introduced by Linehan (1940), next to *P* (primary) and *S* (secondary) waves. It is generated by an earthquake in or near the oceans, propagated in the oceans as an acoustic wave guided in the *SOFAR channel* and converted back to *seismic wave* at the ocean-land boundary near the recording site. See Storchak et al. (2003 and 2011), Kulhanek (2002), Fig. 2.22 in Havskov and Ottemöller (2010), and Figs. 2.72 and 2.73 in Chapter 2 of this Manual.

**trace**

In observational seismology, a trace is a plot of the *time history* of seismic *waveforms* recorded by a channel of a *seismograph* or an *accelerometer*.

**traction**

Force exerted per unit area of a particular internal or external surface by the body on one side (containing the normal vector) of the surface to that on the other. This is a vector in contrast to the *stress tensor*. See *stress*.

|                              |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                          |
|------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>transducer</b>            | Any of various devices that transmit energy from one system to another, sometimes one that converts the energy in form. For recording seismic <i>ground motion</i> , the motion of the <i>seismometer</i> mass has to be transmitted either via a mechanical or optical lever system to the recorder (used in old classical <i>seismographs</i> ) or to be converted into an equivalent electronic <i>signal</i> (as used in all modern seismometers). Transducers use different physical principles and devices such as coil-magnet systems, inductive bridges, capacity half-bridges, piezo-electric effects, interferometric-optical devices, etc. Accordingly, the output signal of the transducer may be proportional to ground <i>displacement</i> , <i>velocity</i> or <i>acceleration</i> .                                                                                                                                                                                                                                                                                                                                                                      |
| <b>transfer function</b>     | In general terms the <i>Laplace</i> or <i>Fourier transform</i> of the <i>impulse response</i> of a linear system ( <i>filter</i> , sensor, etc.). It is equivalent to the Laplace or Fourier transform of a broadband output <i>signal</i> divided by that of the input signal. In seismology the transfer function of a seismic sensor-recorder system (= <i>seismograph</i> ), or of the Earth medium through which <i>seismic waves</i> propagate, describes the <i>frequency-dependent amplification</i> , <i>damping</i> and <i>phase</i> distortion of <i>seismic signals</i> by a specific sensor-recorder device (or the propagation medium). The amplitude (or absolute value or modulus) of the complex transfer function is termed the <i>amplitude-frequency response</i> function or <i>magnification curve</i> of a seismograph. The influence of the related phase response on the seismic wave forms is often ignored in the classical routine observatory practice of amplitude measurements, yet it has to be taken into account for correct simulation of the transfer function. See section 5.2.7 of Chapter 5 in this Manual and Scherbaum (2002). |
| <b>transform fault</b>       | A <i>strike-slip fault</i> that connects segments of convergent or divergent <i>plate boundary</i> features, such as ridge-to-ridge, ridge-to-trench, etc. This type of fault was proposed by Wilson (1965), and confirmed by Sykes (1967). See Uyeda (2002, p. 57), and the biography of J. Tuzo Wilson in Howell (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>transition zone</b>       | A term in the study of the Earth's interior, referring to the depth range from 410 to about 800 km, in which solid-state phase transitions produce strong seismic <i>velocity gradients</i> and sharp <i>discontinuities</i> . Sometimes used only for the depth range 410-660 km between the two major discontinuities at those depths. See the depth range of TZ in Fig. 2.79 of Chapter 2 of this Manual; see also Lay (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>translation</b>           | See <i>displacement</i> and <i>translational motions</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>translational motions</b> | The term "translational" is strongly preferred over "linear" because the latter conflicts with matters of linearity <i>versus</i> nonlinearity in soils, instruments, mathematics, and so forth. If                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |

*displacements, velocities, and accelerations* are not otherwise qualified, they are assumed to refer to translational motions. The in-line displacements are indicated  $u_x$ ,  $u_y$ , and  $u_z$  in Figure 1 of Evans et al. (2010).

|                                      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
|--------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>transmission coefficient</b>      | The ratio of <i>amplitude</i> of the transmitted wave relative to the incident wave at a sub-surface <i>discontinuity</i> . The coefficient is found by imposing boundary conditions that the seismic <i>displacement</i> and the <i>traction</i> acting on the discontinuity surface are continuous across the surface. These conditions physically mean a welded contact and absence of extra-seismic sources at the discontinuity. Coefficients have been defined for different normalizations and signs of the component waves, e.g., energy or displacement (Knott, 1899, and Zoeppritz, 1919). See <i>reflection coefficient</i> , Chapman (2002, p. 110-112), and Aki and Richards (2002 or 2009, p. 128-149). |
| <b>transmitting boundary</b>         | In analytical studies, a boundary to a region of interest that approximates the effects of media outside the boundary. Transmitting boundaries or boundary elements are often used in <i>finite element</i> studies to include the effects of far extents of solid or fluid domains.                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>transport equation</b>            | The ordinary differential equations describing the variation of the <i>amplitude</i> coefficients along a <i>ray</i> . See Chapman (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>transverse coherence function</b> | Measure of <i>coherency</i> of transmitted <i>wave field</i> as a function of the transverse-separation of receivers. See Sato et al. (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>transverse isotropy</b>           | A term introduced by Love (1927) to describe a medium with one axis of cylindrical symmetry. In the geophysical literature, this axis is often implicitly considered as vertical, so that the <i>velocity</i> and <i>polarization</i> are independent of the <i>azimuth</i> of propagation in the horizontal plane. See Cara (2002).                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>travel time</b>                   | The travel time is the time taken for the <i>seismic waves</i> to propagate from one point to another along the <i>ray</i> . See Chapter 2 in this Manual, and Chapman (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>travel-time curve</b>             | A graph of travel time of a considered <i>seismic phase</i> plotted against the epicentral distance of the recording <i>station</i> . Seismic <i>velocities</i> within the Earth can be computed from the slopes of the resulting curves. See Figs. 2.30, 2.32-2.33, 2.36/-2.37, and 2.59-2.61 in Chapter 2 of this Manual and the entries <i>travel time</i> , <i>epicentral distance</i> , <i>hodograph</i> and <i>Wiechert-Herglotz-Formula</i> .                                                                                                                                                                                                                                                                  |
| <b>tremor</b>                        | See <i>episodic tremor and slip (ETS)</i> , <i>non-volcanic tremor</i> and <i>volcanic tremor</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |



|                             |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|-----------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>trench</b>               | In global scale it means <i>oceanic trench</i> . In <i>earthquake geology</i> , it means a trench excavated across a <i>fault trace</i> for investigating of the <i>fault zone</i> structure.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>trench log</b>           | A map of an excavation or <i>trench</i> wall that exposes geologic formations or structures for detailed study. Trench logs constitute primary data records for many <i>paleoseismic</i> investigations. See Grant (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>triaxial seismometer</b> | A set of three <i>seismic sensors</i> whose sensitive axes are perpendicular to each other. See Wielandt (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>triggered earthquake</b> | An earthquake that results from <i>stress</i> changes (e.g., those from a distant earthquake) that are considered small compared to ambient stress levels at the earthquake source. See <i>induced earthquake</i> . and McGarr et al. (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>triggering</b>           | In seismometry, the turning on of an instrument, especially an <i>accelerograph</i> , by <i>ground motion</i> of a prescribed level. In soil mechanics, the onset of a liquefied condition. In earthquake seismology, an earthquake may be triggered due to <i>stress</i> redistribution in the wake or a preceding neighboring strong earthquake. A striking example for the latter has been the triggering of the November 12, 1999 M7.2 Düzce earthquake along a side branch of the North Anatolian Fault (NAF) in Turkey after the August 17, 1999 M7.6 Izmit earthquake on the main branch of the NAF. See also Youd (2003).                                                                                |
| <b>triple junction</b>      | See <i>plate motion</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
| <b>triplication</b>         | Behavior of the <i>travel-time curve</i> showing three <i>arrivals</i> for the same range of <i>epicentral distance</i> caused by a sharp increase of <i>velocity</i> with depth. This happens when a part of a <i>ray</i> path refracted from below the zone of velocity increase appears at shorter distances than that from above the zone. These two travel-time branches are connected by the so-called <i>receding branch</i> associated with ray paths through the zone. See Aki and Richards (2002, Fig. 9.19, p. 421). See <i>branch (of travel-time curve)</i> , <i>PKP triplication</i> , and P triplications in Figs. 2.30, 2.32, and 2.60 of Chapter 2 and Fig. 11.73 in Chapter 11 of this Manual. |
| <b>tromometer</b>           | Originally a simple-pendulum <i>seismoscope</i> , observed with a microscope, to measure background vibrations. Used through about 1900 as another term for <i>seismometer</i> . See Agnew (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>tsunami</b>              | In Japanese meaning “wave (running into the) harbour”. A train of gravity sea waves set up by a disturbance in the sea bed or water column, e.g., by a submarine earthquake, <i>landslide</i> , or explosion of a volcanic island. Tsunami <i>amplitudes</i> are typically small in the open ocean, but when running into                                                                                                                                                                                                                                                                                                                                                                                        |

shallow waters near the coast the tsunami wavefront sometimes piles up the waters 30 m or higher and sweeps up to several km into shallow land. The *tsunami run-up height* strongly depends on coastal profile and shape. Bays and cone-shaped river mouths increase run-up heights. See Benz et al. (2002).

**tsunami earthquakes** Earthquakes that generate much larger *tsunamis* than expected from their *magnitudes*. Tsunami earthquakes have much smaller than usual *rupture velocities* and *stress drops*. See Kanamori (1972), Satake (2002), Polet and Kanamori (2009), and Bormann and Di Giacomo (2011).

**tsunamigenic earthquake** An earthquake that causes a *tsunami*. Most large shallow earthquakes under the sea are tsunamigenic. See Satake (2002).

**tsunami magnitude ( $M_t$ )** The *magnitude* of an earthquake determined from the height of a *tsunami* wave at a given travel distance representing the strength of a tsunami source. The tsunami magnitude  $M_t$ , referenced to *moment magnitude*  $M_w$ , was introduced by Abe (1979). See Satake (2002), and Utsu (2002c).

**tsunami run-up height** The *tsunami* height on land above a reference sea level measured at the maximum inundation distance from the coast. See Satake (2002).

**tube waves in a borehole** In an empty cylindrical hole, a kind of *surface wave* can propagate along the axis of the hole with energy confined to the vicinity of the hole. They exhibit *dispersion* with *phase velocity* increasing with the *wavelength*. At wavelengths much shorter than the borehole radius, they approach *Rayleigh waves*. The phase velocity reaches the shear velocity at wavelengths of about three times the radius. Beyond this cutoff wavelength, they attenuate quickly by radiating *S waves*. In a fluid-filled cylindrical hole, in addition to a series of multi-reflected *conical waves* propagating in the fluid, tube waves exist without a cutoff for the entire *period* range. At short wavelengths, they approach *Stoneley waves* for the plane liquid-solid *interface*. For wavelengths longer than about 10 times the hole radius, the velocity of tube waves becomes constant, given in terms of the *bulk modulus*  $\kappa$  of the fluid and the *rigidity*  $\mu$  of the solid, by  $v = c / \sqrt{1 + \kappa/\mu}$ , where  $c$  is the acoustic velocity in the fluid.

**turbidite** Sea-bottom deposit formed by massive slope failures of large sedimentary deposits. These slopes fail in response to, e.g., earthquake shaking or excessive sedimentation load.

**T wave** See *T phase*.

**twistor** A geometrical object that consists of a pair of *spinor* fields. Twistors describe rotational states. See Majewski (2008).

**twist** A term that is used to describe a shear *deformation* caused by torsional moment. It is widely used in the European literature.

**two-dimensional site amplification** A site effect calculated for a geologic structure varying in the vertical direction and in only one horizontal direction. In a purely two-dimensional case, incident waves are assumed to be *homogeneous* along the same horizontal direction. The case, in which the incident wave field is allowed to be three-dimensional, such as *spherical waves*, is sometimes called 2.5-dimensional. See Kawase (2003).

## U

**ultra-long-period (ULP) earthquake** In volcano seismology, long-period earthquakes with *waveforms* having dominant *periods* of 100 seconds or longer. See Chouet (1996a).

**unconformity** Surface of erosion or non-deposition that separates younger strata from older rocks.

**underplating** Addition of igneous material to the base of the *crust*, a process thought to have occurred at volcanic rifted margins and beneath some ocean islands, or addition of material (including sediments) to the base of an *accretionary wedge* by thrusting at *subduction zones*. See Minshull (2002).

**uniform hazard (response) spectrum** In *probabilistic hazard analysis*, a *response spectrum* with ordinates having equal probability of being exceeded. See Campbell (2003).

**uniform shear beam** A beam with constant properties, but arbitrary cross-section, that can deform only in shear and whose response is described by a one-dimensional wave equation. A cantilevered shear beam is often used as a simplified model of a building. See Jennings (2003).

**unreinforced masonry** Construction that employs brick, stone, clay tile or similar materials, but lacks steel bars or other strengthening elements. See Jennings (2003).

**upper mantle** Uppermost part of the Earth's *mantle*, connected with mineralogical phase transformations and increase of seismic velocities at several *discontinuities*, located between the *Moho* and a depth of about 700 km.

**UTC** An acronym for the Universal Time Coordinated, a time scale defined by the International Time Bureau and agreed upon by international convention. Seismology is using UTC and the former Greenwich Mean Time (GMT) since more than 100

years as global time base for *seismic bulletins* and *seismic catalogues*.

## V

|                                          |                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
|------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>VBB</b>                               | An acronym for very broadband, in reference to the 0.1-second to 360-second band of <i>seismic signals</i> . See Hutt et al. (2002).                                                                                                                                                                                                                                                                                                             |
| <b>velocity</b>                          | In reference to earthquake shaking, velocity is the time rate of change of ground <i>displacement</i> of a reference point during the passage of earthquake <i>seismic waves</i> , commonly expressed in nanometer or micrometer per second. However, velocity may also refer to the speed of propagation of seismic waves through the Earth, commonly expressed in kilometers per second.                                                       |
| <b>velocity, apparent</b>                | See <i>apparent velocity</i> .                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>velocity structure</b>                | In seismology, a generalized local, regional or global model of the Earth that represents its structure in terms of the <i>velocities</i> of <i>P</i> and/or <i>S</i> waves as a function of depth and/or of its lateral distribution. See, e.g., Figs. 2.77, 2.79, and 2.80 in Chapter 2 of this Manual; Lay (2002), Mooney et al. (2002), and Minshull (2002).                                                                                 |
| <b>very-long-period (VLP) earthquake</b> | In volcano seismology, long-period earthquakes with <i>waveforms</i> having dominant <i>periods</i> of 10's of seconds. See Chouet (1996a).                                                                                                                                                                                                                                                                                                      |
| <b>vent</b>                              | See <i>volcanic vent</i> .                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>vesicle</b>                           | Cavity formed in a volcanic rock as a result of exsolution and expansion of gas phase when the rock was still in a molten state, either below the surface as <i>magma</i> or during the slow cooling and solidification of thick <i>lava flows</i> or <i>lava lakes</i> .                                                                                                                                                                        |
| <b>virtual seismic network</b>           | See <i>seismic network</i> .                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>viscoelasticity</b>                   | The property of a material that causes its <i>response</i> to an applied force to involve both elastic and viscous components. Several models of viscoelastic materials have been developed that depend on the initial and long-time response to a step in <i>stress</i> , such as Maxwell and Voigt models. The short-time response of a Maxwell material is that of an elastic solid, while the long-time response is that of a viscous fluid. |
| <b>viscous damping</b>                   | The most commonly used type of <i>damping</i> for analytical or numerical studies of earthquake <i>response</i> of a structure. A viscous damper generates a force proportional to the relative velocity of its two attachment points in a direction that opposes the motion. Viscous damping is one of the few forms of                                                                                                                         |

|                                         |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                            |
|-----------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                         | damping that can be described by linear equations. See Jennings (2003).                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>visible earthquake waves</b>         | Slow waves with long <i>period</i> and short <i>wavelengths</i> reported by eyewitnesses in the <i>epicentral area</i> of a <i>great earthquake</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>volatiles</b>                        | Chemical species or compounds that are dissolved in <i>magmas</i> at high <i>pressure</i> and that appear as low- <i>density</i> gases at low pressure; the most common ones in magmas are water, carbon dioxide and sulfur dioxide.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |
| <b>volcanic arc</b>                     | Long, arcuate chains of <i>volcanoes</i> that are created above <i>subduction zones</i> , and are currently active or were operative in the geologic past. It can form on land as volcanic mountains or in the sea as volcanic islands.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>volcanic ash</b>                     | <i>Pyroclastic</i> fragments of rocks, minerals, and <i>volcanic glass</i> smaller than 2 mm in size produced during explosive <i>eruptions</i> ; the pyroclastic deposit formed by the accumulation of such fragments.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                    |
| <b>volcanic block</b>                   | See <i>block, volcanic</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
| <b>volcanic bomb</b>                    | See <i>bomb, volcanic</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>volcanic chains</b>                  | Roughly linear alignments of a series of volcanoes, extinct, dormant, or active. Most volcanic chains are located along or near the boundaries of <i>tectonic plates</i> , but some can be of <i>intraplate</i> origin, the best-known example of which is the Hawaiian Ridge-Emperor Seamount Chain in the interior of the Pacific Plate.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
| <b>volcanic degassing</b>               | A collective term for the processes by which <i>volatiles</i> and <i>volcanic gases</i> escape from <i>magma</i> or <i>lava</i> . Degassing can occur continuously or discontinuously at highly variable rates before, during, and after <i>eruptions</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                |
| <b>Volcanic Explosivity Index (VEI)</b> | First proposed by Newhall and Self (1982), a now widely used, open-ended semi-quantitative classification scheme - based principally on the height of the <i>eruption column</i> and the erupted volume of <i>tephra</i> - to describe the size of explosive <i>eruptions</i> . The largest historical eruption (Tambora, Indonesia, in 1815) is assigned a VEI = 7; for comparison, the 18 May 1980 eruption of Mount St. Helens was ranked as VEI = 5. However, much larger explosive eruptions in the geologic past (e.g., voluminous <i>caldera</i> -forming events at Yellowstone volcanic system, Wyoming) can be assigned VEI = 8. Non-explosive eruptions are VEI = 0, regardless of size. As with earthquakes, small to moderate-size explosive volcanic events (< VEI = 5) occur much more frequently than large events (> VEI = 5); Simkin and Siebert (1994) have assigned VEI |

estimates for all the world's known eruptions during the last 10,000 years. In terms of the energy involved, the energy of the 1980 Mount St. Helens (VEI = 5) eruption has been estimated to be about  $2 \times 10^{16}$  J, which is roughly the amount of *seismic energy* released by an  $M_w = 8$  earthquake. The energy of the 1815 Tambora eruption (VEI = 7) has been estimated to be about  $1 \times 10^{19}$  J, and is thus comparable to the seismic energy of the 1960 Chilean earthquake ( $M_w = 9.5$ ), the largest instrumentally recorded earthquake so far.

**volcanic gases**

Are produced from exsolution of *volatiles*, which were originally dissolved in *magma* at great *pressure* (depth), during magma rise into, and storage within, lower-pressure regions. Such gases are released from the magmatic system either passively (when the volcano is quiescent) or abruptly upon rapid expansion and *eruption*. Explosively expanding volcanic gases initiate and propel eruptions.

**volcanic glass**

Quenched *lava* that contains no visible or only a few submicroscopic crystals.

**volcanic plume**

Synonymous with *eruption column*, the term is also sometimes used to describe the quiescent degassing of water vapor and other gases above a *volcanic vent*.

**volcanic tremor**

*Seismic signals* which are generated by volcanic activity and are highly variable in character, ranging from those indistinguishable from *tectonic earthquakes* to continuous vibrations with relatively low *frequencies* (0.5-5 Hz). The continuous vibrations are known as volcanic or harmonic tremors and are likely caused by the involvement of gas-fluid interaction. Harmonic tremors have a very uniform appearance, whereas spasmodic tremors are pulsating and consist of higher frequencies with a more irregular appearance. See McNutt (2002).

**volcanic unrest**

A general term that applies to a *volcano's* behavior when it departs from its usual or background level of activity, as evidenced by visible or instrumentally measurable changes - seismic, geodetic, or geochemical - in the state of the volcano from volcano-monitoring studies. However, not all episodes of volcanic unrest, which can last from a few days to a few decades, culminate in eruptive activity.

**volcanic vent**

Surface opening at and through which volcanic materials are erupted, explosively or non-explosively; such openings can be fissures or point sources, vary widely in form and complexity.

**volcano**

A mountain, hill, or plateau formed by the accumulation of materials erupted - effusively or explosively - through one or more openings (*volcanic vents*) in the Earth's surface; also

|                                         |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                        |
|-----------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
|                                         | refers to the vents themselves. However, some volcanoes have a negative topographic expression, such as <i>calderas</i> or <i>craters</i> .                                                                                                                                                                                                                                                                                                                                                                                            |
| <b>volcano hazards</b>                  | Potentially damaging volcano-related processes and products that occur during or following eruptions. In quantitative hazard assessments, the probability of a given area being affected by potentially destructive phenomena within a given period of time. At the start of the 21 <sup>st</sup> century, more than half a billion of the world's population is at risk from volcano hazards. For general discussions of volcano hazards and their monitoring and mitigation, see Scarpa and Tilling (1996) and Tilling (1989, 2002). |
| <b>volcano monitoring</b>               | The general term applied to the systematic surveillance of the physical and chemical changes in the state of a volcano and its associated hydrothermal system, before, during, and after <i>volcanic unrest</i> .                                                                                                                                                                                                                                                                                                                      |
| <b>volcano risk</b>                     | Relating to the adverse impacts of <i>volcano hazards</i> , generally involving the consideration of the general relation:<br>$\text{risk} = \text{hazard} \times \text{vulnerability} \times \text{value (what is at risk)}.$ In probabilistic risk assessments, the probability or likely magnitude of human and economic <i>loss</i> is calculated from the same relation.                                                                                                                                                          |
| <b>volcano seismology</b>               | Seismological study of the structures and processes beneath volcanoes. For reviews see McNutt (2002) and Chouet (1996b and 2003).                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>volcano-tectonic (VT) earthquake</b> | Often used to refer to ordinary (high-frequency or <i>brittle-failure</i> earthquakes) in a volcanic environment. This usage reflects an ambiguity whether the <i>stresses</i> leading to the earthquake are a result of regional <i>tectonic</i> processes or local magmatic processes, such as dike intrusion, or some combination thereof. See <i>A-type earthquake</i> .                                                                                                                                                           |
| <b>Volume (Vol. n)</b>                  | A defined stage of the operations of processing <i>strong-motion</i> data, typically consisting of Vol. 1 (raw <i>acceleration</i> data), Vol. 2 (corrected acceleration data), Vol. 3 ( <i>response spectra</i> ), and Vol. 4 ( <i>Fourier spectra</i> ). See Shakal et al. (2003).                                                                                                                                                                                                                                                   |
| <b>Vulcanian eruption</b>               | Term is taken from descriptions of <i>eruptions</i> on the Island of Vulcano (the home of the Roman god Vulcan), Italy, in the late 19 <sup>th</sup> century. Characterized by discrete, violent explosions lasting second to minutes in duration, Vulcanian activity is generally considered to be more energetic than Strombolian activity (see <i>Strombolian eruption</i> ).                                                                                                                                                       |
| <b>vulnerability V</b>                  | Expected degree of loss ( $0 < V < 1$ ; 0 - no loss/damage, 1 - total loss/damage) due to a disaster event such as strong earthquake                                                                                                                                                                                                                                                                                                                                                                                                   |

shaking (see *earthquake hazard*, *elements exposed to risk*, *loss* and *loss function*).

## W

|                            |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
|----------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Wadati-Benioff zone</b> | A dipping planar (flat) zone of earthquakes that is produced by the interaction of a downgoing oceanic crustal <i>plate</i> with a continental plate. These earthquakes can be produced by <i>slip</i> along the <i>subduction thrust fault</i> (thrust <i>interface</i> between the continental and the oceanic plate) or by slip on <i>faults</i> within the downgoing plate as a result of bending and extension as the plate is pulled into the <i>mantle</i> . Slip may also initiate between adjacent segments of downgoing plates. Wadati-Benioff zones are usually well developed along the <i>trenches</i> of the <i>Circum-Pacific belt</i> , dipping towards the continents. See also <i>subduction zone</i> . |
| <b>watt (W)</b>            | The <i>SI</i> derived unit for power, or radiant flux, expressed in meter <sup>2</sup> kilogram second <sup>-3</sup> (m <sup>2</sup> kg s <sup>-3</sup> ), or joule per second (J/s). It is named after James Watt (1736-1819) who invented the modern steam engine.                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>water table</b>         | The upper surface of a body of unconfined ground water at which the water <i>pressure</i> is equal to the atmospheric pressure.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                           |
| <b>wave field</b>          | A general term used to describe the whole group of different propagating <i>seismic waves</i> radiated from one <i>seismic source</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                   |
| <b>waveform (data)</b>     | The complete analog or sufficiently dense sampled digital representation (i.e., without <i>aliasing</i> ) of a continuous wave group (e.g., of a seismic <i>phase</i> ) or of a whole wave train ( <i>seismogram</i> ) in contrast to <i>parameter data</i> . Accordingly, waveform data allow to reconstructing and analyzing the whole seismic phase or earthquake record both in the <i>time</i> and <i>frequency domain</i> whereas <i>parameter data</i> describe the <i>signal</i> only by a very limited number of more or less representative measurements such as <i>onset time</i> , maximum signal <i>amplitude</i> and related <i>period</i> .                                                                |
| <b>waveform inversion</b>  | A term used both in Earth structure and earthquake source studies, indicating the use of the whole <i>waveform data</i> instead of their particular parameter characteristics. Various inversion procedures can be used for estimating the model parameters of Earth structure and earthquake source, by minimizing the misfit between the observed and <i>synthetic seismograms</i> based on trial parameters. See Jackson (2002).                                                                                                                                                                                                                                                                                       |
| <b>wavefront</b>           | A wavefront is a 3D surface in space with the same travel time from a common source point. See <i>travel time</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |



|                                                  |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |
|--------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>wavelength</b>                                | The spatial distance between adjacent points of equal <i>phase</i> of wave motion (e.g., crest to crest or trough to trough). See Fig. 2.5 in Chapter 5 of this Manual.                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |
| <b>wavelet</b>                                   | A “little wave”, wave ripple, or short wave group related to a distinct <i>seismic phase</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                  |
| <b>wavelet transform</b>                         | The wavelet transform can be used to decompose time-series. It gives information in both the <i>time</i> and <i>frequency domains</i> , and is therefore useful for describing non-stationary <i>signals</i> like <i>seismograms</i> . The wavelet transform decomposes the signal at different scales, thus characterizing its components at different resolutions. Wavelet coefficients at high resolutions show the fine structure of the time series, and those at low resolutions characterize its coarse features. Possible applications for wavelet transform are de-noising of seismograms (e.g., Galiana-Merino et al., 2003) or automatic phase detection (e.g., Zhang et al., 2003). |
| <b>wavenumber</b>                                | The ratio between angular frequency $\omega = 2\pi f$ and the wave propagation <i>velocity</i> $v$ , i.e., $k = \omega/v$ . See <i>wavenumber vector</i> .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                      |
| <b>wavenumber filtering</b>                      | A technique for processing recordings from an <i>array</i> of <i>seismometers</i> to enhance <i>signals</i> with a given range of angular frequency $\omega$ and of the radial component $\mathbf{k}_R$ of the <i>wavenumber vector</i> $\mathbf{k}$ in the horizontal (azimuthal) direction of wave propagation. The <i>wave field</i> is then assumed to be composed of plane harmonic waves with frequency $\omega$ propagating along the surface with horizontal <i>slowness vector</i> $\mathbf{k}_R/\omega$ . The reciprocal of the absolute value of $\mathbf{k}_R/\omega$ is the <i>apparent velocity</i> $v_{app}$ . See Chapter 9 of this Manual, and Douglas (2002).                 |
| <b>wavenumber vector <math>\mathbf{k}</math></b> | $\mathbf{k}$ is the vector in wave propagation direction, i.e., perpendicular to the propagating <i>wavefront</i> , with the scalar amount of the <i>wavenumber</i> $k$ . See Chapter 9 and this Manual and Douglas (2002).                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
| <b>WGS84</b>                                     | An acronym for World Geodetic System, 1984. A system of coordinates conventionally used and distributed by GPS as GPS coordinates. Includes an ellipsoid with inverse flattening $1/f = 298.25722$ and semi-major axis = 6378.137 km (DMA, 1987). The flattening $f$ is defined as the difference between the equatorial and polar radius of the Earth divided by the former. All <i>seismic stations</i> should report their positions in WGS84 coordinates. See Feigl (2002).                                                                                                                                                                                                                 |
| <b>wide-angle seismic method</b>                 | A term used for seismic experiments in which the source and receiver are widely separated compared to the target depth. This term is preferred over the traditional “ <i>seismic refraction</i> ” method because of the recognition that much of the information                                                                                                                                                                                                                                                                                                                                                                                                                                |

on crustal structure in modern experiments comes from reflected phases. See Minshull (2002).

**Wiechert-Herglotz inversion** In the case that the *wave velocity*  $v = f(z)$  increases monotonously with depth  $z$  in the Earth a continuous *prograde travel-time curve* exists because all *rays* return to the surface. In this case an exact analytical solution of the *inverse problem* exists, i.e., when knowing the *apparent horizontal velocity*  $v_{\text{app}}$  at any point of *epicentral distance*  $D$ , we can calculate the velocity  $v_{\text{zp}}$  at the turning point of the ray that returns to the surface at  $D$ . The inversion formulas have been derived by Herglotz (1907) and Wiechert (Wiechert and Geiger, 1910). They permit to calculate 1D Earth models by assuming that lateral variations of velocity are negligible as compared to the vertical velocity variations. See section 2.5.3.6 in Chapter 2 of this Manual and Schweitzer (2003).

**Wiener filter** Optimal *filter* after Wiener (1949) to remove *noise* from *signal* under the assumptions that the time series is ergodic, that noise and signal somehow differ, and that certain information on noise and/or signal are given, at best the signal itself or the *autocorrelation functions* of signal and noise. Wiener proposed to minimize the expectations of the squared differences between desired and actual filter output.

**Wilson cycle** A successive recurrence of opening and closing of ocean basins due to *plate motions* on a time scale of about 100 million years. It was proposed by Wilson (1968). See Uyeda (2002, p. 63) and Dickinson (2002).

**WKBJ approximation** A method for approximate solution of wave propagation through an *inhomogeneous* medium, in which the *velocity* variations are relatively weak and occur over distances large compared with the *wavelengths* of interest. This method is named after Gregor Wentzel, Hendrik Kramers, Leon Brillouin, and Harlod Jeffreys. See Aki and Richards (2002 or 2009, Box 9.6, p. 434-437), and the biographies of Wentzel, Kramers, Brillouin and Jeffreys in Howell (2003).

**W phase** The term W phase was introduced by Kanamori (1993) to describe the complex interference of multiple *reflections* and/or *P-to-S* or *S-to-P conversions* of long-period *body waves* of large earthquakes. The W phase appears as a long-period (approximately 100 -1000 s) phase in the *seismogram* between *P* and *S* waves. Recently, this phase came into focus as a strong candidate for deriving in real-time operation stable and non-saturating *moment magnitude* and *source mechanism* estimates of large earthquakes without waiting for *surface wave* information. This is of great advantage for *tsunami early warning systems*. See Kanamori and Rivera (2008) and Hayes et al. (2009).

**WWSSN** An acronym for the World-Wide Standardized Seismograph Network, established in the early 1960s and equipped with narrow-band short-*period* and long-period *seismographs* (see *magnification curves* in Fig. 3.20 of Chapter 3 in this Manual). Later, several stations were upgraded for digital data acquisition, also in an intermediate period band, and operated as DWWSSN between 1981 and 1993. See Hutt et al. (2002).

## X

**xenolith** A piece of preexisting solid *mantle* or crustal rock that is picked up, and incorporated into, flowing *magma*; also sometimes called inclusion. It is a direct sample of the uppermost mantle and the *crust*.

## Y

**Young`s modulus** See *stretch modulus*.

## Z

**zero-frequency earthquakes** A term sometimes used for *silent earthquakes*, i.e., very slow *fault* movements which release energy without perceptible seismic radiation over periods of hours to months rather than within seconds to minutes that are characteristic of “normal” earthquakes.

**zero-initial-length spring** A helical spring that is pre-stressed so that the force it exerts in extension is proportional to the distance between the points of attachment, rather than the difference between that distance and the initial length of the spring. Originally developed by L. LaCoste for use in gravimeters, it found important applications in the design of long-period vertical *seismographs* that can theoretically achieve an infinite *period* without instability. See Aki and Richards (2002 or 2009, p. 602-603).

**zero-phase filter** A *filter*, which causes no time shift (or time shift is corrected for) and no *phase* distortions of the *signal*. Usually, such filters have symmetrical *impulse response* and are therefore called two-sided or *acausal filters*. When applied to a very impulsive signal, the symmetry of the filter may lead to precursory ('acausal') oscillations, making it hard to identify the true *onset* (see Scherbaum, 2002 and Figure 7 in IS 11.4).

**z-transform** A discrete equivalent of the *Laplace transform*. See Scherbaum (2002) and Claerbout (1976, p. 1-8).

## References

- Abe, K. (1979). Size of great earthquakes of 1837-1974 inferred from tsunami data. *J. Geophys. Res.*, **84**, 1561-1568.
- Adams, R. D. (2002). International Seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *Internat. Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 29-37.
- Agnew, D. C. (2002). History of seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 3-11.
- Akaike, H. (1970). Statistical predictor estimation. *Ann. Inst. Statist. Math.*, **22**, 203-217.
- Akaike, H. (1971). Autoregressive model fitting for control. *Ann. Inst. Statist. Math.*, **23**, 163-180.
- Aki, K. (1987). Magnitude frequency relation for small earthquakes: A clue to the origin of  $f_{max}$  of large earthquakes. *J. Geophys. Res.*, **92**, 1349-1355.
- Aki, K. (2002). Synthesis of earthquake science information and its public transfer: A history of the Southern California Earthquake Center. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 39-49.
- Aki, K., and Chouet, B. (1975). Origin of coda waves: Source, attenuation and scattering effects. *J. Geophys. Res.*, **80**, 3322.
- Aki, K., and Kaminuma, K. (1963). Phase velocity of Love waves in Japan (Part 1). *Bull. Earthquake Res. Inst.*, **41**, 243-259.
- Aki, K. and Lee, W. H. K. (1976). Determination of three-dimensional velocity anomalies under a seismic array using first *P* arrival times from local earthquakes, Part I, A homogeneous initial model. *J. Geophys. Res.*, **81**, 4381-4399.
- Aki, K., and Lee, W. H. K. (2003). Glossary of interest to earthquake and engineering seismologists. In: *International Handbook of Earthquake and Engineering Seismology, Part B*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 1793-1856.
- Aki, K., and Richards, P. G. (1980). *Quantitative seismology – theory and methods*, Volume II, Freeman and Company, ISBN 0-7167-1059-5.
- Aki, K., and Richards, P. G. (2002). *Quantitative seismology*. Second Edition, ISBN 0-935702-96-2, *University Science Books*, Sausalito, CA., xvii + 700 pp.
- Aki, K., and Richards, P. G. (2009). *Quantitative seismology*. 2. ed., corr. Print., Sausalito, Calif., Univ. Science Books, 2009, XVIII + 700 pp., ISBN 978-1-891389-63-4 (see [http://www.ldeo.columbia.edu/~richards/Aki\\_Richards.html](http://www.ldeo.columbia.edu/~richards/Aki_Richards.html)).
- Aki, K., Christofferson, A., and Husebye, E. S. (1977). Determination of the three-dimensional seismic structure of the lithosphere. *J. Geophys. Res.*, **82**, 277-296.
- Allen, R. V. (1978). Automatic earthquake recognition and timing from single traces. *Bull. Seism. Soc. Am.*, **68**, 1521-1532.
- Allen, Q. M., Gasperini, P., Kamigaichi, O., and Böse, M. (2009). The status of earthquake early warning around the world. An introductory overview. *Seism. Res. Lett.*, **80**, 682-693; doi: 10.1785/gssrl.80.5.682.
- Ambraseys, N. N., Jackson, J. A., and Melville, C. P. (2002). Historical seismicity and tectonics: The case of the eastern Mediterranean and the Middle East. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 747-763.
- Anderson, J. A., and Wood, H. O. (1925). Description and theory of the torsion seismometer. *Bull. Seism. Soc. Am.*, **15**, 1-72.

- Anderson, J. G. (2003). Strong-motion seismology. In: *International Handbook of Earthquake and Engineering Seismology, Part B*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 937–965.
- Anderson, J. G., and Hough, S. E. (1984). A model for the shape of the Fourier amplitude spectrum of acceleration at high frequencies. *Bull. Seism. Soc. Am.*, **74**, 1969–1993.
- Atkinson, G. M., and Wald, D. J. (2007). “Did You Feel It?” Intensity Data: A Surprisingly Good Measure of Earthquake Ground Motion. *Seism. Res. Lett.*, **78**, 362–368.
- Baisch, S., Ceranna, L., and Harjes, H.-P. (2008). Earthquake cluster: What can we learn from waveform similarity? *Bull. Seism. Soc. Am.*, **98**, 2806–2814.
- Bak, P. (1996). *How Nature Works – The Science of Self-Organized Criticality*. Springer, New York.
- Ballou, G. (1987). Filters and equalizers. In: *Handbook for Sound Engineers*. Edited by G. Ballou, p. 569–612, Howard W. Sams, Indianapolis.
- Banzhaf, W. (2002). Self-organizing systems. In: *Encyclopedia of Science and Technology*. Third edition, Volume 14, p. 589–598, Academic Press, San Diego.
- Bardet, J.-P. (2003). Advances in analysis of soil liquefaction during earthquakes. In: *International Handbook of Earthquake and Engineering Seismology, Part B*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 1175–1201.
- Båth, M. (1954). The elastic waves Lg and Rg along Euroasiatic paths. 1954. *Arkiv för Geofysik*, **2**(13), 295–342.
- Ben-Menahem, A. (1959). Elastodynamics of rupturing finite earthquake faults. *Ph.D. thesis*, California Institute of Technology.
- Ben-Menahem, A., and Singh, S. J. (1981). *Seismic Waves and Sources*. Springer-Verlag, New York, xxi+1108 pp; ISBN 978-1-4612-5858-2.
- Ben-Menahem, A., and Singh, S. J. (2000). *Seismic Waves and Sources*. 2<sup>nd</sup> edition, *Dover Publications*, New York.
- Ben-Menahem, A., Smith, S. W., and Teng, T. L. (1965). A procedure for source studies from spectrums of long-period body waves. *Bull. Seism. Soc. Am.*, **55**, 203–235.
- Ben-Zion, Y. (2003). Appendix 2. Key formulas in earthquake seismology. In: *International Handbook of Earthquake and Engineering Seismology, Part B*, edited by W.H.K. Lee, H. Kanamori, P.C. Jennings, and C. Kisslinger, Academic Press, Amsterdam, 1857–1875.
- Benz, H. M., Okubo, P., and Villasenor, A. (2002). Three-dimensional crustal P-wave imaging of Mauna Loa and Kilauea Volcanoes, Hawaii. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 407–420.
- Bevington, P. R., and D. K. Robinson (2003). *Data reduction and error analysis for the physical sciences*. Third Edition, McGraw-Hill Higher Education, xi+320 pp.; ISBN 0-07-247227-8.
- Birch, F. (1966). Compressibility; Elastic constants. In: “Handbook of Physical Constants.” Edited by S. P. Clark, Revised edition, p. 97–173, Geological Society of America, New York.
- Blackman, R. B., and Tukey, J. W. (1958). *The Measurement of Power Spectra*. Dover, New York, 190 pp.
- Bleistein, N. (2002). Wave phenomena. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 17, p. 789–804, Academic Press, San Diego.
- Bodin, P., Gomberg, J., Singh, S. K., and Santoyo, M. (1997). Dynamic deformations of shallow sediments in the valley of Mexico, part I: Three dimensional strains and rotations recorded on a seismic array. *Bull. Seism. Soc. Am.*, **87**, 528–539.
- Bolt, B. A., and Abrahamson, N. A. (2003). Estimation of strong seismic ground motions. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003).

- International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 983-1001.
- Bonilla, M. G. (1970). Surface faulting and related effects. In: *Earthquake Engineering*. Edited by R. L. Wiegel, p. 47-74, Prentice-Hall, Englewood Cliffs.
- Boore, D. M. (1983). Stochastic simulation of high-frequency ground motions based on seismological models of the radiated spectra. *Bull. Seismol. Soc. Am.*, **73**, 1865-1894.
- Boore, D. M. (2003). Simulation of ground motion using the stochastic method. *Pure Appl. Geophys.*, **160**, 635-676.
- Borcherdt, R. D., Hamburger, R. O., and Kircher, C. A. (2003). Seismic design provisions and guidelines in the United States: A prologue. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1127-1132.
- Bormann, P. (ed.) (2002 and 2009). Glossary. In: IASPEI New Manual of Seismological Observatory Practice. ISBN 3-9808780-0-7, *GeoForschungsZentrum Potsdam*, Vol. 2, 26 pp.; see <http://nmsop.gfz-potsdam.de>; doi:10.2312/GFZ.NMSOP\_r1\_glossary.
- Bormann, P. (2010). From earthquake prediction research to time-variable seismic hazard assessment applications. *Pure Appl. Geophys.*, **198**(1-2), 329-366, doi: 10.1007/s00024-010-0114-0.
- Bormann, P. (2011). Earthquake magnitude. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 207-218; doi: 10.1007/978-90-481-8702-7.
- Bormann, P., and Di Giacomo, D. (2011). The moment magnitude  $M_w$  and the energy magnitude  $M_e$ : common roots and differences. *J. Seismology*, **15**, 411-427; doi: 10.1007/s10950-010-9219-2.
- Bormann, P., and Saul, J. (2008). The new IASPEI standard broadband magnitude  $m_B$ . *Seism. Res. Lett.* **79** (5), 699-706.
- Bormann, P., and Saul, J. (2009). Earthquake magnitude. In: *Encyclopedia of Complexity and Systems Science*, edited by A. Meyers, Springer, Heidelberg, Vol. 3, 2473-2496.
- Bormann, P. (2011). Earthquake magnitude. In: Harsh Gupta (ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, 207-218; doi: 10.1007/978-90-481-8702-7.
- Bormann, P., Liu, R., Ren, X., Gutdeutsch, R., Kaiser, D., Castellaro, S. (2007). Chinese national network magnitudes, their relation to NEIC magnitudes, and recommendations for new IASPEI magnitude standards. *Bull. Seism. Soc. Am.* **97**, 114-127.
- Bormann, P., Liu, R., Xu, Z., Ren, K., Zhang, L., Wendt S. (2009). First application of the new IASPEI teleseismic magnitude standards to data of the China National Seismographic Network. *Bull. Seism. Soc. Am.* **99** (3), 1868-1891; doi: 10.1785/0120080010
- Bouchon, M. (2003). A review of the Discrete Wavenumber Method. *Pure Appl. Geophys.*, **160**, 445-465.
- Box, G. E. P. Jenkins, G. M., and Reinsel, G. C. (2008). *Time Series Analysis, Forecasting and Control* (4th ed.). Hoboken, NJ: Wiley; [ISBN 9780470272848](https://doi.org/10.1002/9780470272848).
- Bullen, K. (1950). An Earth model based on a compressibility-pressure hypothesis. *Mon. Not. R. Astr. Soc., Geophys. Suppl.*, **6**, 50-59.
- Buwalda, J. P. (1937). Shutteridges, characteristic physiographic features of active faults (abstract). *Geol. Soc. Am. Proc.* 1936, p. 307.
- Brigham, E.O. (1988). *The Fast Fourier Transform and its Applications*. Prentice Hall, Englewood Cliffs.
- Brune, J. N. (1970). Tectonic stress and the spectra of shear waves from earthquakes. *J. Geophys. Res.*, **75**, 4997-5009.
- Brune, J. N., and Engen, G. R. (1969). Excitation of mantle Love waves and definition of mantle wave magnitude. *Bull. Seism. Soc. Am.*, **59**, 923-933.
- Brune, J. N., and Thatcher, W. (2002). Strength and energetics of active fault zones. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International*

- Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 569-588.
- Burridge, R., and Knopoff, L. (1964). Body force equivalents for seismic dislocations. *Bull. Seism. Soc. Am.*, **54**(6A), 1875-1888.
- Burridge, R., and Knopoff, L. (1967). Model and theoretical seismicity. *Bull. Seism. Soc. Am.*, **57**, 341-371.
- Campbell, K. W. (2003). Strong-motion attenuation relations. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1003-1012.
- Cara, M. (2002). Seismic anisotropy. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 875-885.
- Castellaro, S., and Bormann, P. (2007). Performance of different regression procedures on the magnitude conversion problem. *Bull. Seism. Soc. Am.*, **97**, 1167-1175.
- Castellaro, S., Mulargia, F., and Kagan, Y. Y. (2006). Regression problems for magnitudes. *Geophys. J. Int.*, **165**, 913-930.
- Červený, V. (2001). *Seismic ray theory*. ISBN 0-521-36671-2, Cambridge University Press, New York, VII + 713 pp.
- Chapman, C. H. (2002). Seismic ray theory and finite frequency extensions. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 103-123.
- Chouet, B. (1996a). Long-period volcano seismicity: its source and use in eruption forecasting. *Nature*, **380**, 309-316.
- Chouet, B. (1996b). New methods and future trends in seismological volcano monitoring. In: Scarpa, R., and Tilling, R. (1996), 23-97.
- Chouet, B. A. (2003). Volcano seismology. *Pure Appl. Geophys.*, **160**, 739-788.
- Christensen, N. I., and Stanley, D. (2003). Seismic velocities and densities of rocks. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1587-1594.
- Claerbout, J.F. (1976). *Fundamentals of Geophysical Data Processing*. McGraw-Hill, New York.
- Clark, S.P. (Editor) (1966). *Handbook of Physical Constants*. Revised edition, Geological Society of America, New York.
- Cochard, A., Igel, H., Schuberth, B., Suryanto, W., Velikoseltsev, A., Schreiber, U., Wassermann, J., Scherbaum, F., and Vollmer, D. (2006). Rotational motions in seismology: theory, observations, simulation. In: *Earthquake source asymmetry, structural media and rotation effects*. Edited by Teyssere, R., Takeo, M., and Majewski, E., p. 391-412, Springer-Verlag.
- Cooley, J. S., and Tukey, J. W. (1965). An algorithm for the calculation of complex Fourier Series. *Math. Comp.*, **19**, 297-301.
- Cosserat, E., and Cosserat, F. (1909). *Théorie des Corps Déformables*, Paris: Hermann (available from the Cornell University Library Digital Collections) (in French).
- Cox, A. (ed.) (1973). *Plate tectonics and geomagnetic reversals*. W. H. Freeman, San Francisco.
- Crary, A. P. (1955). A brief study of ice tremors. *Bull. Seismol. Soc. Am.*, **45**, 1-10.
- Curtis, A., and Snieder, R. (2002). Probing the earth's interior with seismic tomography. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 861-874.

- Das, S., and Aki, K. (1977). Fault plane with barriers: A versatile earthquake model. *J. Geophys. Res.*, **82**(36), 5658-5670.
- Das, S., and Kostrov, B. V. (1988). *Principles of earthquake source mechanics*. Cambridge University Press.
- Decker, R. and Decker, B. (1998). *Volcanoes*. Third Edition, W.H. Freeman, New York.
- Dickinson, W.R. (2002). Plate tectonics. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 12, p. 475-489, Academic Press, San Diego.
- Dieterich, J. H. (1978). Time-dependent friction and the mechanics of stick-slip. *Pure Appl. Geophys.*, **116**, 790-806.
- Dieterich, J. H. (1979). Modeling of rock friction I. Experimental results and constitutive equations. *J. Geophys. Res.*, **84**, 2161-2168.
- Dietz, R. S. (1961). Continent and ocean basin evolution by spreading of the sea floor. *Nature*, **190**(4779), 854-857.
- Douglas, A. (2002). Seismometer arrays – their use in earthquake and test ban seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (eds) (2002), *Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 357-367.
- Dowrick, D. (1977). *Earthquake Resistant Design for Engineers and Architects*. Second Edition, Wiley, New York.
- Dragert, H., Wang, K., and James, T. S. (2001). A silent slip event on the deeper Cascadia subduction interface. *Science*, **292**, 1525-1528.
- Duff, P. M. D. (Editor) (1993). “Holmes’ Principles of Physical Geology.” Fourth edition, Chapman and Hall, London.
- EERI (1989). Glossary by the EERI Committee on Seismic Risk. *Earthq. Spectra*, **5**, 700-702.
- Eicher, D.L. (2002). Geologic time. In: *Encyclopedia of Science and Technology*. Third edition, Volume 6, p. 623-648, Academic Press, San Diego.
- Endo, E. T., and Murray, T. L. (1991). Real-time seismic amplitude measurement (RSAM): a volcano monitoring and prediction tool. *Bull. Volcanol.*, **53**, 533-545.
- Engdahl, E. R. (2003). International Association of Seismology and Physics of the Earth’s Interior. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1541-1549.
- Engdahl, E. R., and Villasenor, A. (2002). Global seismicity: 1900–1999. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 665-690.
- Eringen, A. (1999). *Microcontinuum Field Theories. I. Foundation and Solids*. Springer, Berlin.
- Evans, J. R., and the International Working Group on Rotational Seismology (2009). Suggested notation conventions for rotational seismology. *Bull. Seism. Soc. Am.*, **99**(2B), 1429–1436.
- Evans, J. R., Hutt, C. R., Nigbor, R. L., and de la Torre, T. (2010). Performance of the new R2 Sensor. Second IWGoRS Workshop, October 11-13, Prague, Czech Republic. [Presentation file available at: <http://karel.troja.mff.cuni.cz/~vackar/IWGoRS/Evans.pdf>, last accessed 21 January 2011].
- Ewing, W. M., Jardetzky, W. W., and Press, F. (1957). *Elastic waves in layered media*. McGraw-Hill, New York.
- Faccioli, E., and Pessina, V. (2003). Use of engineering seismology tools in ground shaking scenarios. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1031-1048.



- Fedotov, S. A. (1965). Regularities of the distribution of strong earthquakes in Kamtchatka, the Kurile Islands, and north-east Japan. *Tr. Inst. Fiz. Zemli Akad. Nauk SSSR*, **36**, 66-93.
- Feigl, K. L. (2002). Estimating earthquake source parameters from geodetic measurements. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 607-620.
- Fichtner, A., and Igel, H. (2009). Sensitivity densities for rotational ground-motion measurements. *Bull. Seism. Soc. Am.*, **99**(2B), 1302-1314.
- Fleming, K., Picozzi, M., Milkereit, C., Kühnlenz, F., Lichtblau, B., Fischer, J., Zulfikar, C., Özel, O., and the SAFER and EDIM working groups (2009). The self-organizing seismic early warning information network (SOSEWIN). *Seismol. Res. Lett.*, **80**(5), 755-771; doi: 10.1785/gssrl.80.5.755,
- Freundt, A., Wilson, C. J. N., and Carey, S. N. (2000). Ignimbrites and block-and-ash flow deposits. In: "Encyclopedia of Volcanoes," p. 581-599, Academic Press, San Diego.
- Galiana-Merino, J. J., Parolai, S., and Rosa-Herranz, J. (2003). Seismic wave characterization using complex trace analysis in the stationary wavelet packet domain. *Soil Dynamics Earthq. Engineer.*, **31**, 1565-1578.
- Gasperini, P., Manfredi, G., and Zschau, J. (2007). *Earthquake early warning systems*. Springer Berlin-Heidelberg, 349 pp.
- Geiger, L. (1910). Herdbestimmung bei Erdbeben aus den Ankunftszeiten. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen*, Mathematisch-Physikalische Klasse, 331-349. (1912 translated to English by F. W. L. Peebles and A. H. Corey: Probability method for the determination of earthquake epicenters from the arrival time only. *Bulletin St. Louis University* **8**, 60-71).
- Geller, R. J., and Mueller, C. S. (1980). Four similar earthquakes in central California. *Geophys. Res. Lett.*, **7**, 821-824.
- Giardini, D., Gruenthal, G., Shedlock, K., and Zhang P. (2003). The GSHAP Global Seismic Hazard Map. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1233-1239.
- Goldstein, H. (1950). *Classical Mechanics*. Addison-Wesley, Cambridge.
- Grant, L. B. (2002). Paleoseismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 475-489.
- Grekova, E. F., and Lee, W. H. K. (compilers) (2009). Suggested readings in continuum mechanics, elasticity, and earthquakes, *Bull. Seism. Soc. Am.*, **99**, 1423-1428
- Grünthal, G. (Ed.) (1998). European Macroseismic Scale 1998, *Cahiers du Centre Européen de Géodynamique et de Seismologie*, **15**, Conseil de l'Europe, Luxembourg, 99 pp.
- Gutdeutsch, R., Castellaro, S., and Kaiser, D. (2011). The magnitude conversion problem: Further insights. *Bull. Seism. Soc. Am.*, **101**, 379-384; doi: 10.1785/0120090365.
- Gutenberg, B. (1945). Amplitude of *P*, *PP* and *S* and magnitude of shallow earthquakes. *Bull. Seism. Soc. Am.*, **35**, 57-69.
- Gutenberg, B., and Richter, C. F. (1941). *Seismicity of the Earth*. Geological Society America, Special Paper **34**, New York 1941, vii + 131 pp.
- Gutenberg, B., and Richter, C.F. (1949). *Seismicity of the Earth and Associated Phenomena*. 1<sup>st</sup> edition, Princeton Univ. Press, Princeton.
- Gutenberg, B., and Richter, C. F. (1956). Magnitude and energy of earthquakes. *Annali di Geofisica*, **9**, 1-15.
- Hagiwara, T. (1964). Brief description of the project proposed by the earthquake prediction research group of Japan. *Proc. U.S.-Japan Conf. Res. Relat. Earthquake Prediction Probl.*, pp. 10-12.

- Hall, J. F. (2003). Finite element analysis in earthquake engineering. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1133-1158.
- Hancock, P. L. and Skinner, B. J. (Editor) (2000). "The Oxford Companion to the Earth." Oxford University Press, Oxford.
- Handin, J. (1966). Strength and ductility. In: *Handbook of Physical Constants*. Edited by S.P. Clark, Revised edition, p. 223-289, Geological Society of America, New York.
- Hanks, T. C. (1979).  $b$ -values and  $\omega^{-\gamma}$  seismic source models: Implications for tectonic stress variations along active crustal fault zones and the estimation of high frequency strong ground motion. *J. Geophys. Res.*, **84**, 2235-2241.
- Hanks, T.C. (1982).  $F_{\max}$ . *Bull. Seism. Soc. Am.*, **72**, 1867-1879.
- Harris, R. A. (2003). Stress triggers, stress shadows, and seismic hazard. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B. Academic Press, Amsterdam, 1217-1232.
- Haskell, N. A. (1964). Total energy and energy spectral density of elastic wave radiation from propagating faults. *Bull. Seism. Soc. Am.*, **54**, 1811-1841.
- Hauksson, E., Jones, L. M., and Shakal, A. F. (2003). TriNet: Modern ground-motion seismic network. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1275-1284.
- Havskov, J., and Ottemöller, L. (2010). *Routine data processing in earthquake seismology*. Springer, xi + 347 pp.
- Hayes, G. P., Rivera, L., and Kanamori, H. (2009). Source inversion of the W-phase: Real-time implementation and extension to low magnitudes. *Seism. Res. Lett.*, **80**(5), 817-822.
- Herglotz, G. (1907). Über das Benndorfsche Problem der Fortpflanzungsgeschwindigkeit der Erdbebenwellen. *Physikalische Zeitschrift*, **8**, 145-147.
- Hess, H. H. (1962). History of ocean basins. In: *Petrologic Studies*. Buddington Volume, p. 599-620, Geological Society of America, New York.
- Hill, D. P., Fisher, F.G., Lahr, K.M., and Coakley, J.M. (1976). Earthquake sounds generated by body wave ground motion. *Bull. Seism. Soc. Am.*, **66**, 1159-1172.
- Hill, D. P., Ellsworth, W. L., Johnston, M. J. S., Langbein, J. O., Oppenheimer, D. H., Pitt, A.M., Reasenber, P. A., Sorey, M. L., and McNutt, S.R. (1990). The 1989 earthquake swarm beneath Mammoth Mountain, California – An initial look at the 4 May through 30 September activity. *Bull. Seism. Soc. Am.*, **80**, 325-339.
- Housner, G.W. (1952). Spectral intensity of strong motion earthquakes. Proc. Symp. Blast and Earthquake Effects on Structure, Earthquake Engineering Research Institute. [A computer readable file of this symposium proceeding is placed on the attached Handbook CD #2. of Lee et al., 2002]
- Housner, G. W. (1970). Design Spectrum, Chapter 5 of *Earthquake Engineering*, edited by Robert W. Weigel, Prentice Hall, Englewood Cliffs, p. 93-106.
- Howell, B. F., (Ed.) (2003). Biographies of interest to earthquake and engineering seismologists. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B. Academic Press, Amsterdam, 1725-1787.
- Husseini, M. I. (1977). Energy balance for formation along a fault. *Geophys. J. R. Astron. Soc.*, **49**, 699-714.
- Hutt, C. R., Bolton, H. F., and Holcomb, L.G. (2002). US contribution to digital global seismograph networks. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 319-332.

- IASPEI (2005) Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/summary\\_of\\_WG\\_recommendations\\_2005.pdf](http://www.iaspei.org/commissions/CSOI/summary_of_WG_recommendations_2005.pdf).
- IASPEI (2013). Summary of Magnitude Working Group recommendations on standard procedures for determining earthquake magnitudes from digital data. [http://www.iaspei.org/commissions/CSOI/Summary\\_of\\_WG\\_recommendations\\_20130327.pdf](http://www.iaspei.org/commissions/CSOI/Summary_of_WG_recommendations_20130327.pdf).
- Ishimoto, M., and Iida, K. (1939). Observations sur les seismes enregistres par le microseismographe construit dernièrement. *Earthq. Res. Inst. Bull., Tokyo Univ.*, **17**, 443-478.
- Ito, K. (1990). Regional variations of the cutoff depth of seismicity in the crust and their relation to heat flow and large earthquakes. *J. Phys. Earth*, **38**, 223–250.
- Jacobs, J. A. (1987). *The Earth's core*. 2nd ed. Academic Press, London and Orlando, x + 416 pp.
- Jackson, J. A. (Editor) (1997). *Glossary of Geology*. Fourth Edition. American Geological Institute, Alexandria, Virginia.
- Jackson, J. A. (2002). Using earthquakes for continental tectonic geology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 491-503.
- Jackson, M. (2002). WWW (World Wide Web). In: *Encyclopedia of Science and Technology*. Third Edition, Volume 17, p. 875-885, Academic Press, San Diego.
- Jaeger, J. C., and Cook, N. G. W. (1971). *Fundamental of Rock Mechanics*. Chapman and Hall, London.
- James, G., and James, R. C. (1976). "Mathematics Dictionary." Fourth Edition. Van Nostrand Reinhold, New York.
- Jeanloz, R. (2002). Earth's mantle. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 4, p. 783-799, Academic Press, San Diego.
- Jennings, P. C. (2003). An introduction to the earthquake response of structures. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1097-1125.
- JMA (Japan Meteorological Agency) (1996). Explanation Table of the JMA Seismic Intensity Scale (February 1996), 4pp.
- JMA (Japan Meteorological Agency) (2009). Explanation Table of JMA Seismic Intensity Scale, <http://www.jma.go.jp/jma/kishou/known/shindo/kaisetsu.html> (in Japanese).
- Johnston, M. J. S. (2002). Electromagnetic fields generated by earthquakes. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 621-635.
- Johnston, M. J. S., and Linde, A. T. (2002). Implications of crustal strain during conventional, slow, and silent earthquakes. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 589-605.
- Julian, B. R., Miller, A. D., and Foulger, G. R. (1998). Npon-double-couple earthquakes. *I. Theory Rev. Geophys.*, **36**, 525-549.
- Kalkan, E., and Graizer, V. (2007). Coupled tilt and translational ground motion response spectra. *J. Struct. Eng. ASCE*, **133**, 609-619.
- Kanamori, H. (1972). Mechanism of tsunami earthquakes. *Phys. Planet. Earth Inter.*, **6**, 346-359.
- Kanamori, H. (1977). The energy release in great earthquakes. *J. Geophys. Res.*, **82**, 2981-2987.
- Kanamori, H. (1983). Magnitude scale and quantification of earthquakes. *Tectonophysics*, **93**, 185-199.
- Kanamori, H. (1993). W phase. *Geophys. Res. Lett.*, **20** (16), 1691-1694.

- Kanamori, H. (2003). Earthquake prediction: An overview. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1205-1216.
- Kanamori, H., and Brodsky, E. E. (2004). The physics of earthquakes. *Rep. Progress Physics*, **67**, 1429-1496.
- Kanamori, H., and Rivera, L. (2008). Source inversion of W phase: speeding up seismic tsunami warning. *Geoph. J. Int.*, **175**, 222–238, doi: 10.1111/j.1365-246X.2008.03887.x
- Kawase, H. (2003). Site effects on strong ground motions. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1013-1030.
- King, C.-Y., and Igarashi, G. (2002). Earthquake-related hydrological and geochemical changes. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 637-645.
- Kinoshita, S. (1992). Local characteristics of the fmax of bedrock motion in Tokyo metropolitan area. *J. Phys. Earth*, **40**, 487-515.
- Kinoshita, S. (2003). Kyoshin Net (K-NET), Japan. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1049-1056.
- Klein, F. W. (2003). The HYPOINVERSE2000 earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1619-1620.
- Knopoff, L. (1958). Energy released in earthquakes. *Geophys. J. Roy. Astron. Soc.*, **1**, 44-52.
- Knopoff, L., and Chen, Y. T. (2009). Single-couple component of far-field radiation from dynamical fractures. *Bull. Seism. Soc. Am.*, **99**(2B), 1091–1102.
- Knott, C.G. (1899). Reflexion and refraction of elastic waves, with seismological applications. *The London, Edinburgh, and Dublin Philosophical Magazine and Journal of Science, Series 5*, **48**, 64-97 + 567-569.
- Kostrov, B.V. (1966). Unsteady propagation of longitudinal shear cracks. *J. Appl. Math. Mech.*, **30**, 1241-1248.
- Kulhanek, O. (2002). The structure and interpretation of seismograms. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 333-348.
- Lacroix, A. (1904). *La Montagne Pelée et ses éruptions*. Masson et Cie., Paris.
- Lahr, J. C. (2003). The HYPOELLIPSE earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1617-1618.
- Lahr, J. C., and Van Eck, T. (2003). Global inventory of seismographic networks, In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1657-1663 and CD #3 (87\_Lahr).
- Langston, C. A. (2007). Wave gradiometry in two dimensions. *Bull. Seism. Soc. Am.*, **97**, 401–416.
- Larson, R. L. (1991). Geological consequences of superplumes. *Geology*, **19**, 963-966.
- Larson, R. L., and Pitmann, W. C. (1972). World-wide correlation of Mesozoic magnetic anomalies and its implications. *Geol. Soc. Am. Bull.*, **83**, 3627-3644.
- Lay, T. (2002). The Earth's Interior. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 829-860.

- Lay, T., and Wallace, T. C. (1995). *Modern global seismology*. ISBN 0-12-732870-X, Academic Press, 521 pp.
- Lay, T., and Kanamori, H. (1981). An asperity model of great earthquake sequences. In: *Earthquake Prediction - An International Review*. Edited by D. Simpson and P. Richards, American Geophysical Union, Washington, D.C., p. 579-592,
- Lee, V. W., M. D. Trifunac, and A. Amini (1982). Noise in earthquake accelerograms. *ASCE, EMD*, **108**(12), 1121-1129.
- Lee, V. W., and Trifunac, M. D. (2009). Empirical scaling of rotational spectra of strong earthquake ground motion. *Bull. Seism. Soc. Amer.*, **99**(2B), 1378-1390; doi: 10.1785/0120080070.
- Lee, W. H. K. (2002). Challenges in observational seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 269-281.
- Lee, W. H. K. (Compiler) (2009). A glossary for rotational seismology. *Bull. Seism. Soc. Am.*, **99**(2B), 1,082–1,090.
- Lee, W. H. K., and Stewart, S. W. (1981). Principles and applications of micro-earthquake networks, *Advances in Geophysics*, Supplement 2, *Academic Press*, New York.
- Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C., (Editors), (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 933(+23) pp and 1 CD-ROM.
- Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C., (Editors), (2003a). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1012 pp and 2 CD-ROMs.
- Lee, W. H. K., Lahr, J. C., and Valdes, C. M. (2003b). The HYPO71 earthquake location program. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1641-1642.
- Lee, W. H. K., Celebi, M., Todorovska, M. I., and Igel, H. (Eds.) (2009). Special Issue on Rotational Seismology and Engineering Applications, *Bull. Seism. Soc. Am.*, **99**(2B).
- Lee, W. H. K., Celebi, M., Igel, H., and Todorovska, M. I. (2009a). Introduction to the Special Issue on Rotational Seismology and Engineering Applications. *Bull. Seism. Soc. Am.*, **99**(2B), 945–957.
- Lee, W. H. K., Huang, B. S., Langston, C. A., Lin, C. J., Liu, C. C., Shin, T. C., Teng, T. L., and Wu, C. F. (2009b). Review: Progress in rotational ground-motion observations from explosions and local earthquakes in Taiwan. *Bull. Seism. Soc. Am.*, **99**(2B), 958–967.
- Lide, D. R. (2002). *CRC Handbook of Chemistry and Physics*. 83<sup>rd</sup> edition, CRC Press, Boca Raton.
- Linehan, D. (1940). Earthquakes in the West Indian region. *Trans. Am. Geophys. Union, EOS*, **30**, 229-232.
- Lockner, D. A., and Beeler, N.M. (2002). Rock failure and earthquakes. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*. *Academic Press, Amsterdam*, 505-537.
- Lognonne, P., and Clevede, E. (2002). Normal modes of the Earth and planets. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 125-147.
- Lolli, B., and Gasperini, P. (2012). A comparison among general orthogonal regression methods applied to earthquake magnitude conversions. *Geophys. J. Int.*, **190**, 1135–1151 doi: 10.1111/j.1365-246X.2012.05530.x
- Longuet-Higgins, M. S. (1950). A Theory of the Origin of Microseisms. *Phil. Trans. R. Soc. London, Series A, Mathematical and Physical Sciences*, **243**(857), 1-35.



- Love, A. E. H. (1911). *Some problems of geodynamics*. Cambridge Univ. Pr., Cambridge, xxvii + 180 pp.
- Love, A. E. H. (1927; 1944). *A Treatise on the Mathematical Theory of Elasticity*. Fourth edition, Cambridge University Press, Cambridge; Republication by Dover Publications, New York in 1944.
- Macdonald, G. A. (1972). *Volcanoes*. Prentice-Hall, Englewood Cliffs.
- Madariaga, R. (1976). Dynamics of an expanding circular fault. *Bull. Seism. Soc. Am.*, **66**, 639-666.
- Madariaga, R., and Olsen, K. B. (2002). Earthquake Dynamics. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 175-194.
- Maslov, V.P. (1965): Teorija vozmuschenij i asimptoticheskie metody. Mosk. gos (Univ. im. M. V. Lomonosova). Izd. Univ. Moskva, Moscow, **1965**, 549 pp.
- Maxwell, J. C. (1864). A dynamical theory of the electromagnetic field. [A reprint of this article was published In: *The World of Physics*. Edited by J.H. Weaver, Volume 1, p. 846-858, Simon and Schuster, New York, 1987].
- Maxwell, J. C. (1891). *A Treatise on Electricity and Magnetism*. Third Edition, Clarendon Press. Reprinted by Dover Publications, New York, in 1954.
- Majewski, E. (2006). Seismic rotation waves: spin and twist solitons. In: *Earthquake Source Asymmetry, Structural Media and Rotation Effects*, edited by R. Teisseyre, M. Takeo and E. Majewski, p. 255-272, Springer, Berlin.
- Majewski, E. (2008). Twistors as spin and twist solitons. In: *Physics of Asymmetric Continuum: Extreme and Fracture Processes: Earthquake Rotation and Soliton Waves*, edited by R. Teisseyre, H. Nagahama and E. Majewski, p. 273-284, Springer, Berlin.
- Malischewsky, P. (1987). Surface waves and discontinuities. *Elsevier*, Amsterdam, ISBN 0-444-98959-5, 229 pp.
- Malischewsky, P., and Scherbaum, F. (2004). Love's formula and H/V ratio (ellipticity) of Rayleigh waves. *Wave Motion*, **40**(1), 57-67.
- Mandelbrot B. B., and Frame, M. (2002). Fractals. In: *Encyclopedia of Science and Technology*. Third edition, Volume 6, p. 185-207, Academic Press, San Diego.
- Maruyama, T. (1963). On the force equivalents of dynamical elastic dislocations with reference to the earthquake mechanism. *Bull. Earthq. Res. Inst.*, **41**, 467-486.
- Maxwell, J. C. (1991). [\*A Treatise on Electricity and Magnetism, Vol. 1 and 2, unabridged 3rd ed.\*](#) of the original publication in 1873; New York, Dover.
- McCalpin, J. P., and Nelson, A. R. (1996). Introduction to paleoseismology. In: *Paleoseismology*. Edited by J. P. McCalpin, p. 1-32, Academic Press, San Diego.
- McClean, S. I. (2002). Data mining and knowledge recovery. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 4, p. 229-246, Academic Press, San Diego.
- McGarr, A., Simpson, D., Seeber, L. (2002). Case histories of induced and triggered seismicity. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A. Academic Press, Amsterdam, 647-661.
- McNutt, S. R. (2002). Volcano seismology and monitoring for eruptions. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 383-406.
- Merrill, R.T., and McFadden, P. L. (2002). Geomagnetism. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 6, p. 663-674, Academic Press, San Diego.
- Merrill, R. T., McElhinny, M.W., and McFadden, P.L. (1998). *The Magnetic Field of the Earth Paleomagnetism, the Core, and the Deep Mantle*. Academic Press, San Diego.

- Medvedev, S. V. (1962). Engineering seismology (in Russian). *Academy of Sciences, Inst. of Physics of the Earth*, Publ. house for literature on Civil Engineering, Architecture and Building Materials, Moscow.
- Menke, W. (1984). *Geophysical Data Analysis- Discrete Inverse Theory*. Academic Press, Orlando.
- Meyers, R. A. (Editor-in-chief) (2002). *Encyclopedia of Science and Technology*. Third Edition, 18 volumes, Academic Press, San Diego.
- Minakami, T. (1961). Study of eruptions and earthquakes originating from volcanos, I. *Int. Geol. Rev.*, **3**, 712-719.
- Minshull, T. A. (2002). Seismic structure of the oceanic crust and passive continental margins. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 911-924.
- Mooney, W. D., Prodehl, C., and Pavlenkova, N. I. (2002). Seismic velocity structure of the continental lithosphere from controlled source data. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 887-910.
- Morgan, P. (2002). Heat flow. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 7, p. 265-278, Academic Press, San Diego.
- Morley, L. W., and Larochelle, A. (1964). Palaeomagnetism as a means of dating geological events. In: *Geochronology in Canada*. Edited by F. F. Osborne, p. 39-51, Royal Soc. Canada, Special Publ. No. 8.
- Morris, C. (1992). *Dictionary of Science and Technology*. Academic Press, San Diego.
- Mosegaard, K., and Tarantola, A. (2002). Probabilistic approach to inverse problems. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 237-265.
- Murray, T. L., Ewert, J. W., Lockhart, A. B., and LaHusen, R. G. (1996). The integrated mobile volcano-monitoring system used by the Volcano Disaster Assistance Program (VDAP). In: *Monitoring and Mitigation of Volcano Hazards*. Edited by R. Scarpa and R.I. Tilling, p. 315-362, Springer-Verlag, Heidelberg.
- Musson, R. M. W. (2002). Historical earthquakes of the British Isles. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 803-806.
- Musson, R. M. W., and Cčić, I. (2002). Macroseismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 807-822.
- Musson, R. M. W., Grünthal, G. and Stucchi, M. (2010). The comparison of macroseismic intensity scales. *J. Seismol.*, **14**, 413-428.
- Naeim, F. (Editor) (2001). *The Seismic Design Handbook*. Second Edition, Kluwer Academic Publishers, Boston.
- NBS (1981). The International National System of Units (SI). *National Bureau of Standards Special Publication 330*, U.S. Government Printing Office, Washington, D.C.
- Newhall, C. G., and Self, S. (1982). The volcanic explosivity index (VEI) -An estimate of explosive magnitude for historical volcanism. *J. Geophys. Res.*, **87**, 1231-1238.
- Neunhöfer, H., and Güth, D. (1989). Detailed investigation of the great earthquake swarm in Western Bohemia by the local Vogtland network. In: Bormann, P. (ed.). *Monitoring and analysis of the earthquake swarm 1985/86 in the region Vogtland/Western Bohemia*. Akademie der Wissenschaften der DDR, Zentralinstitut für Physik der Erde, Veröffentlichung No. 110, 124-164.

- Nigbor, R. L., Evans, J. R., and Hutt, C. R. (2009). Laboratory and field testing of commercial rotational seismometers. *Bull. Seism. Soc. Am.*, **99**(2B), 1215–1227.
- Nur, A. (2002). Earthquakes and archaeology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 765-774.
- Nur, A., and Byerlee, J. D. (1971). An exact effective stress law for elastic deformation of rock with fluids. *J. Geophys. Res.*, **76**, 6414-6419.
- Obara, K., Hirose, H., Yamamizu, F., and Kasahara, K. (2004). Episodic slow slip events accompanied by non-volcanic tremors in southwest Japan subduction zone. *Geophys. Res. Lett.*, **31**, doi:10.1029/2004GL02084.
- O'Brien P. W., and D. S. Miletic, D. S. (2003). The sociological dimensions of earthquake mitigation, preparedness, response, and recovery. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1241- 1252.
- Okada, Y. (1985). Surface deformation due to shear and tensile faults in a half-space. *Bull. Seism. Soc. Am.*, **75**, 1135-1154.
- Okal, E. A., and Talandier, J. (1989).  $M_m$ : A variable-period mantle magnitude. *J. Geophys. Res.*, **94**, 4169-4193.
- Okal, E. A., and Talandier, J. (1990).  $M_m$ : Extension to Love waves of the concept of a variable-period mantle magnitude. *Pure Appl. Geophys.*, **134**, 355-384.
- Olson, P. (2002). Mantle convection and plumes. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 9, p. 77-94, Academic Press, San Diego.
- Omori, F. (1894). On the aftershocks of earthquakes. *J. College Sci., Imperial Univ. Tokyo*, **7**, 111-200.
- Papageorgiou, A. (2003). The barrier model and strong ground motion. *Pure Appl. Geophys.*, **160**(3-4), 603-634.
- Papageorgiou, A., and Aki, K. (1983). A specific barrier model for the quantitative description of inhomogeneous faulting and the prediction of strong motion, Part I Description of the model. *Bull. Seism. Soc. Am.*, **73**, 693-722; Part II Application of the model. *Bull. Seism. Soc. Am.*, **73**, 955-978.
- Park, R., and Paulay, T. (1975). *Reinforced Concrete Structures*. Wiley, New York.
- Penzien, J., and M. Watabe (1975). Characteristics of 3-dimensional earthquake ground motion. *Earthq. Eng. Struct. Dyn.*, **3**, 365-373.
- Peters, R. (2009). Tutorial on gravitational pendulum theory applied to seismic sensing of translation and rotation, *Bull. Seism. Soc. Am.* **99**(2B), 1050-1063; doi: 10.1785/0120080163.
- Pham, D. N., Igel, H., Wassermann, J., Cochard, A., and Schreiber, U. (2009). The effects of tilt on interferometric rotation sensors. *Bull. Seism. Soc. Am.*, **99**(2B), 1352–1365.
- Polet, J., and Kanamori, H. (2009). Tsunami earthquakes; in: Meyers, R. (ed), *Encyclopedia of complexity and systems science*, Vol. 10, 9577-9592.
- Press, F., and Siever, R. (1994). *Understanding Earth*. W. H. Freeman, New York (and many later editions).
- Pujol, J. (2009). Tutorial on rotations in the theories of finite deformation and micropolar (Cosserat) elasticity. *Bull. Seism. Soc. Am.*, **99**(2B), 1011–1027.
- Purcaru, G., and Berckhemer, H. (1982). Quantitative relations of seismic source parameters and a classification of earthquakes. *Tectonophysics*, **84**, 57-128.
- Raff, A., and Mason, R. (1961). Magnetic surveys off the west coast of North America, 40° N latitude to 50° N latitude. *Bull. Geol. Soc. Am.*, **72**, 1267-70.
- Rayleigh, Lord (J. W. S.) (1887). On waves propagated along the plane surface of an elastic solid. *Proc. London Math. Soc.*, **17**, 4-11.



- Reid, H. F. (1910). The Mechanics of the Earthquake. *The California Earthquake of April 18, 1906, Report of the State Earthquake Investigation Commission, Volume 2*. Carnegie Institution of Washington, 43–47.
- Reiter, L. (1990). *Earthquake Hazard Analysis - Issues and Insights*. Columbia University Press, New York.
- Rice, J. R., and Ruina, A. L. (1983). Stability of steady frictional slipping. *J. Appl. Mech.*, **50**, 343-349.
- Richards, P. G. (2002). Seismological methods of monitoring compliance with the Comprehensive Nuclear Test-Ban Treaty. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 369-382.
- Richards, P., and Wu, Z. (2011). Seismic monitoring of nuclear explosions. In: Gupta, H. (Ed.) (2011). *Encyclopedia of Solid Earth Geophysics*. Springer, Dordrecht, Vol. 1144-1156.
- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bull. Seism. Soc. Am.*, **25**, 1-32.
- Richter, C. F. (1958). Elementary seismology. *W. H. Freeman and Company*, San Francisco and London, viii + 768 pp.
- Rikitake, T., and Hamada, K. (2002). Earthquake prediction. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 4, p. 743-760, Academic Press, San Diego.
- Rogers, G., and Dragert, H. (2003). Episodic tremor and slip on the Cascadia subduction zone: The chatter of silent slip. *Science*, **300**, 1942-1943.
- Romanowicz, B. (2002). Inversion of surface waves: A review. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 149-173.
- Rosenblueth, E. (1951). *A basis for aseismic design of structures*. PhD thesis, Univ. of Illinois, Urbana, Illinois
- Ruff, L. J. (2002). State of stress within the Earth. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 539-557.
- Ruina, A. L. (1983). Slip instability and state variable friction laws. *J. Geophys. Res.*, **88**, 10.359-10.370.
- Runcorn, S. K. (Editor) (1967). "International Dictionary of Geophysics." 2 volumes, Pergamon Press, Oxford.
- Salmon, C. G., and Johnson, J. E. (1996). *Steel Structures -- Design and Behavior*. Fourth Edition, Harper Collins, New York.
- Sambridge, M. (2003). Nonlinear inversion by direct search using the neighbourhood algorithm. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1635-1637.
- Satake, K. (2002). Tsunamis. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 437-451.
- Sato, H., and Fehler, M. (1998). *Seismic Wave Propagation and Scattering in the Heterogeneous Earth*. AIP Press/ Springer Verlag, New York.
- Sato, H., Fehler, M., and Wu, R.-S. (2002). Scattering and attenuation of seismic waves in the lithosphere. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 195-208.
- Savage, J. C. (1971). A theory of creep waves propagating along a transform fault. *J. Geophys. Res.*, **76**, 1954-1966.

- Savage, J. C. (1980). Dislocations in seismology. In: *Dislocations in Solids*. Edited by F.R.N. Nabarro, Volume 3, North Holland Publishing Co., Amsterdam, p. 251-339.
- Savage, J. C., and Wood, M.D. (1971). The relation between apparent stress and stress drop. *Bull. Seism. Soc. Am.*, **61**, 1381-1388.
- Scarpa, R., and Tilling, R. (eds) (1996). *Monitoring and Mitigation of Volcano Hazards*. Springer, Berlin Heidelberg New York, 841 pp.
- Scherbaum, F. (2002). Analysis of digital earthquake signals. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A., Academic Press, Amsterdam, 349-355.
- Scholz, C. H. (2002). *The Mechanics of Earthquakes and Faulting*. 2<sup>nd</sup> edition, Cambridge University Press, Cambridge.
- Schweitzer, J. (2003). Early German contributions to modern seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part B., Academic Press, Amsterdam, on Handbook CD #2 (CD ISBN 0-12-440654-8) file /79\_24germany/ch79\_24parta.pdf, 58 pp.
- Schweitzer, J., and Lee, W. H. K. (2003). Old seismic bulletins to 1920: A collective heritage from early seismologists. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1665-1723.
- Shakal, A. F., Huang, M. J., and Graizer, V. M. (2003). Strong-motion data processing. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 967-981.
- Sharp, R. P. (1954). Physiographic features of faulting in Southern California. *California Div. Mines and Geology Bull.*, **170**, Chapter 5, pt. 3, p. 21-28.
- Shelly, D. R., Beroza, G., Ite, S., and Nakamura, S. (2006). Low-frequency earthquakes in Shikoku, Japan, and their relationship to episodic tremor and slip. *Nature*, **422**, 188-191; doi: 10.1038/nature04931.
- Sheriff, R. E., and Geldart, L. P. (1995). *Exploration seismology*. Cambridge University Press, Second Edition, 592 pp.
- Sibson, R. H. (2002). Geology of the crustal earthquake source. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 455-473.
- Sieberg, A. (1912). Über die makroseismische Bestimmung der Erdbebenstärke. *Gerl. Beitr. Geophys.*, **11**, 227-239.
- Sieberg, A. (1923). *Geologische, physikalische und angewandte Erdbebenkunde*. Verlag Gustav Fischer, Jena, 572 S.
- Sieberg, A. (1932). *Geologie der Erdbeben*. *Handbuch der Geophysik*, Gebrüder Bornträger, Berlin, vol. 2, pt 4, 550-555.
- Sigurdsson, H. (Editor-in-Chief) (2000). "Encyclopedia of Volcanoes." Academic Press, San Diego.
- Simkin, T., and Siebert, L. (1994). *Volcanoes of the World -A Regional Directory, Gazetteer, and Chronology of Volcanism During the Last 10,000 Years*, Second Edition, Geoscience Press, Inc., Tucson, Arizona (in association with the Smithsonian Institution, Washington, D.C.).
- Simkin, T., Unger, J.D., Tilling, R.I., Vogt, P.R., and Spall, H. (1994). Second Edition of This Dynamic Planet -World Map of Volcanoes, Earthquakes, Impact Craters, and Plate Tectonics. U.S. Geological Survey Special Map (Mercator projection scale 1 30,000,000 at the equator).

- Sipkin, S. A. (2002). USGS Earthquake Moment Tensor Catalog. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 823-825.
- Sleeman, R., van Wettum, A., and Trampert, J. (2006). Three-channel correlation analysis: a new technique to measure instrumental noise of digitizers and seismic sensors. *Bull. Seism. Soc. Am.*, **96**(1), 258-271, doi:10.1785/0120050032.
- Socolar, J. (2002). *Chaos*. In: Encyclopedia of Science and Technology. Third edition, Volume 2, p. 637-665, Academic Press, San Diego.
- Somerville, P., and Y. Moriwaki, Y. (2003). Seismic hazards and risk assessment in engineering practice. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1065-1080.
- Song, X. (2002). The Earth's core. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 925-933.
- Sponheuer, W., and Karnik, V. (1964). Neue seismische Skala, In: Sponheuer, W., (Ed.), Proc. 7th Symposium of the ESC, Jena, 24-30 Sept. 1962, *Veröff. Inst. f. Bodendyn. u. Erdbebenforsch. Jena d. Deutschen Akad. d. Wiss.*, **No 77**, 69-76.
- Spudich, P., Steck, L. K., Hellweg, M., Fletcher, J. B., and Baker, L. M. (1995). Transient stresses at Parkfield, California, produced by the M 7.4 Landers earthquake of June 28, 1992: Observations from the UPSAR dense seismograph array. *J. Geophys. Res.*, **100**(B1), 675-690.
- Stein, S., and Klosko, E. (2002). Earthquake mechanisms and plate tectonics. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 69-78.
- Stephens, C. D., Chouet, B. A., Page, R. A., Lahr, J. C., and Power, J. A. (1994). Seismological aspects of the 1989-1990 eruptions at Redoubt Volcano, Alaska: The SSAM perspective. *J. Volcan. Geotherm. Res.*, **62**, 153-182.
- Stover, C. W., and Coffman, J. L. (1993). Seismicity of the United States, 1568-1989 (Revised). *United States Government Printing Office*, Washington.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2003). The IASPEI Standard Seismic Phase List. *Seism. Res. Lett.*, **74**, 761-772.
- Storchak, D. A., Schweitzer, J., and Bormann, P. (2011). Seismic Phase Names: IASPEI standard. In: Gupta, H. (Ed.). *Encyclopedia of Solid Earth Geophysics*, Springer, Vol. 2. 1162-1173.
- Stratton, J. A. (1941). *The electromagnetic theory*. McGraw-Hill, New York.
- Stromeyer, D., Grünthal, G., and Wahlström, R. (2004). Chi-square regression for seismic strength parameter relations, and their uncertainties, with application to an Mw based earthquake catalogue for central and northwestern Europe. *J. Seismol.*, **8**(1), 143-153.
- Suyehiro, K., and Mochizuki, K. (2002). Marine seismology. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 421-436.
- Sykes, L. R. (1967). Mechanism of earthquakes and nature of faulting on the mid-oceanic ridges. *J. Geophys. Res.*, **75**, 5041-5055.
- Tarantola, A. (2005). *Inverse problem theory and methods for model parameter estimation*. SIAM - Soc. for Industrial and Applied Math., Philadelphia, Pa. 2005, xii + 342 pp. (Free download: <http://www.ipgp.fr/~tarantola/Files/Professional/Books/index.html>)
- Teisseyre, R. (2009). Tutorial on new developments on the physics of rotational motions. *Bull. Seism. Soc. Am.* **99**(2B), 1028-1039.

- Teisseyre, R., and Majewski, E. (2002). Physics of earthquakes. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 229-235.
- Tilling, R. I. (1989). Volcanic hazards and their mitigation Progress and problems. *Rev. Geophys.*, **27**, 237-269.
- Tilling, R. I. (2002). Volcanic hazards. In: *Encyclopedia of Science and Technology*. Third Edition, Volume 17, p. 559-577, Academic Press, San Diego.
- Trifunac, M. D. (2009). Review: Rotations in structural response. *Bull. Seism. Soc. Am.* **99**(2B), 968-979; doi: 10.1785/0120080068.
- Trifunac, M. D., and Todorovska, M. I. (2001). A note on the useable dynamic range of accelerographs recording translation. *Soil Dyn. & Earthq. Eng.*, **21**(4), 275–286.
- Turcotte, D. L., and Malamud, B. D. (2002). Earthquakes as a complex system. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 209-227.
- Twiss, R., Souter, B., and Unruh, J. (1993). The effect of block rotations on the global seismic moment tensor and patterns of seismic P and T axes. *J. Geophys. Res.*, **98**, 645–674.
- Udias, A. (2002). Theoretical seismology: An introduction. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 81-102.
- Uhrhammer, R. A., and Collins, E. R. (1990). Synthesis of Wood-Anderson seismograms from broadband digital records. *Bull. Seism. Soc. Am.*, **80**, 702-716.
- USGS Photo Glossary (2002). Photo glossary of volcanic terms. In: Website of the U.S. Geological Survey Volcano Hazards Program; provides images and definitions of some common volcano terms <<http://volcanoes.usgs.gov/Products/Pglossary/pglossary.html>>.
- Utsu, T. (2002a). A list of deadly earthquakes in the World: 1500–2000. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 691-717.
- Utsu, T. (2002b). Statistical features of seismicity. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 719-732.
- Utsu, T. (2002c). Relationships between magnitude scales. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 733-746.
- Uyeda, S. (1978). *The New View of the Earth*. W. H. Freeman, San Francisco.
- Uyeda, S. (1982). Subduction zones - an introduction to comparative subductology. *Tectonophysics*, **81**, 133-159.
- Uyeda, S. (2002). Continental drift, sea-floor spreading, and plate/plume tectonics. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 51-67.
- Vallance, J. W. (2000). Lahars. In: *Encyclopedia of Volcanoes*. p. 601-616, Academic Press, San Diego.
- Vere-Jones, D., and Ogata, Y. (2003). Statistical principles for seismologists. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology, Part B*, Academic Press, Amsterdam, 1573-1586.
- Vine, F. J., and Matthews, D. H. (1963). Magnetic anomalies over oceanic ridges. *Nature*, **199**, 947-949.
- von Huene, R., and Scholl, D. (1991). Observation at convergent margins concerning sediment subduction, subduction erosion, and the growth of continental crust. *Revs. Geophys. Space Phys.*, **29**, 279-316.

- von Rebeur-Paschwitz, E. (1892). Das Horizontalpendel und seine Anwendung zur Beobachtung der absoluten und relativen Richtungs-Änderungen der Lothlinie. *Nova Acta der Ksl. Leop.-Carol. Deutschen Akademie der Naturforscher*, **LX**(1), Halle 1892, 216 pp.
- von Rebeur-Paschwitz, E. (1893). Ueber die Möglichkeit, die Existenz von Mondgliedern in der scheinbaren täglichen Oscillation der Lothlinie nachzuweisen. *Astronomische Nachrichten*, **133**(3169), 1-24.
- Waldhauser, F., and Ellsworth, W. L. (2000). A double-difference earthquake location algorithm: method and application to the Northern Hayward Fault, California. *Bull. Seism. Soc. Am.*, **90**, 1353-1368; doi:10.1785/0120000006.
- Webb, S. C. (2002). Seismic noise on land and on the sea floor. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 305-318.
- Wegener, A. (1912): Die Entstehung der Kontinente. *Petermanns Mitteilungen*, 1912, 185-195 + 253-256 + 305-309.
- Wegener, A. (1912): Die Entstehung der Kontinente. *Geologische Rundschau*, **3**(4), 276-292.
- Weinstein, S. A., and Okal, E. A. (2005). The mantle wave magnitude  $M_m$  and the slowness parameter  $\Theta$ : Five years of real-time use in the context of tsunami warning. *Bull. Seism. Soc. Am.*, **95**, 779-799; doi: 10.1785/0120040112.
- WGCEP (Working Group on California Earthquake Probabilities) (1995). Seismic hazards in southern California -probable earthquakes, 1994-2014. *Bull. Seism. Soc. Am.*, **85**, 379-439.
- Wiechert, E., and Geiger, L. (1910). Bestimmung des Weges der Erdbebenwellen im Erdinnern. *Physikalische Zeitschrift*, **11**, 294-311.
- Wielandt, E. (2002). Seismometry. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International Handbook of Earthquake and Engineering Seismology*, Part A, Academic Press, Amsterdam, 283-304.
- Wiener, N. (1949). *Extrapolation, interpolation, and smoothing of stationary time series*. Technology Press of the Mass. Inst. of Technology, Cambridge, Mass.; ix+163 pp.
- Willis, B. (1923). Earthquake risk in California. *Bull. Seism. Soc. Am.*, **13**, 147-154.
- Wilson, J. T. (1965). A new class of fault and their bearing on continental drift. *Nature*, **207**, 343-347.
- Wilson, J. T. (1968). Did the Atlantic close and then re-open again? *Nature*, **211**, 676-681.
- Wolfram, S. (2002). *A new kind of science*. Wolfram Media, Champaign.
- Wood, H. O., and Neumann, F. (1931). Modified Mercalli intensity scale of 1931. *Bull. Seism. Soc. Am.*, **21**, 277-283.
- Wyss, M., and Brune, J. N. (1968). Seismic moment, stress, and source dimensions for earthquakes in the California-Nevada region. *J. Geophys. Res.*, **73**, 4681-4694.
- Yeats, R. S., Sieh, K., and Allen, C. R. (1997). *The Geology of Earthquakes*. Oxford University Press, New York.
- Youd, T. L. (2003). Liquefaction mechanisms and induced ground failure. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2003). *International Handbook of Earthquake and Engineering Seismology*, Part B, Academic Press, Amsterdam, 1159-1173.
- Zembaty, Z. (2009a). Rotational seismic load definition in Eurocode 8, Part 6 for slender tower-shaped structures. *Bull. Seism. Soc. Am.*, **99**(2B), 1040-1049.
- Zembaty, Z. (2009b). Tutorial on surface rotations from the wave passage effects - Stochastic approach. *Bull. Seism. Soc. Am.*, **99**, 1483-1485.
- Zhang, H., Thurber C., and Rowe C. (2003). Automatic P-wave arrival detection and picking with multiscale wavelet analysis for single-component recordings. *Bull. Seism. Soc. Am.*, **93**, 1904-1912.
- Zoback, M. D., and Zoback, M. L. (2002). State of stress in the Earth's lithosphere. In: Lee, W. H. K., Kanamori, H., Jennings, P. C., and Kisslinger, C. (Eds.) (2002). *International*

*Handbook of Earthquake and Engineering Seismology, Part A*, Academic Press, Amsterdam, 559-568.

Zoeppritz, K. (1919). Erdbebenwellen VII. VIIb. Über Reflexion und Durchgang seismischer Wellen durch Unstetigkeitsflächen. *Nachrichten von der Königlichen Gesellschaft der Wissenschaften zu Göttingen, Mathematisch-physikalische Klasse*, 66-84.